On the Conversion of Antarctic Ice-Mass Change to Sea Level Equivalent

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Introduction

The ice sheets of Antarctica and Greenland have been experiencing a negative mass balance in recent decades because of melting and accelerated discharge of ice across grounding lines, these losses being only partly offset by increased snowfall (Shepherd et al. 2012; Hanna et al. 2013). Changes of ice-shelf mass are not considered here because they do not contribute to sea-level change. According to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC), the average global sea level rose at an average rate of 1.7 (1.5–1.9) mm per year from 1901 to 2010 and approximately 3.2 (2.8–3.6) mm per year from 1993 to 2010 (Vaughan et al. 2013). Although continental glacier melting, ocean

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thermal expansion and polar ice-sheet changes jointly contribute to the prevailing sea level change rate, it is the Antarctic ice sheet (AIS) that will potentially play a significant role in the future because of its massive volume, which accounts for approximately 90% of the world's ice and 70% of the world's fresh water (McMillan et al. 2014). Recent estimates for the overall mass change rate of the AIS are –71 $\pm$ 53 Gt yr^{-1} (or 0.20 $\pm$ 0.15 mm yr^{-1} SLE) from 1992 to 2011 (Shepherd et al. 2012), and –159 $\pm$ 48 Gt yr^{-1} (or 0.45 $\pm$ 0.14 mm yr^{-1} SLE) from 2010 to 2013 (McMillan et al. 2014).

Generally, these and other estimates are made based on observations using different remote sensing technologies and ground measurement methods, including Interferometric Synthetic Aperture Radar (InSAR) for ice-flow monitoring, gravity change detection for the study of mass balance and satellite altimetry for ice-sheet surface change measurement. Such observations have been applied to calculate the mass changes of the ice sheet in three ways: quantifying the sum of mass changes through snow accumulation, ice melting, and ice flow discharge; calculating the mass change through volume variation detected by surface elevation changes; and weighing of the ice sheet by gravity change measurements (Rignot and Thomas 2002). The resulting mass change estimates from different methods do not always agree because of measurement errors, lack of knowledge of the ice thickness and firn/ice density, inaccurate models for glacial isostatic adjustment (GIA), and other factors (Zwally and Giovinetto 2011; Shepherd et al. 2012; Hanna et al. 2013). However, recent observations and analysis have shown an overall mass loss of the Antarctic ice sheet, with the dominant mass loss mainly from West Antarctica through basal melting and ice discharge. This mass loss exhibits an accelerating trend (Rignot and Thomas 2002; IPCC 2013).

This technical note discusses the conversion of the Antarctic ice-mass change to the globally averaged SLE based on up-to-date available information. Certain assumptions are made to simplify such a complex problem. One basic assumption is that the observed change in the Antarctic mass is totally converted to sea water that becomes equally distributed across the global ocean. Further assumptions are made with regard to the global ocean area, the water density, the density of changed mass and other values required for the conversion, as explained in the following sections.

Basic Assumptions

The transformation of the Antarctic ice-mass change into the global SLE is a complex problem that involves many factors. Due to our limited understanding of the Antarctic system, which belongs to the even more complex Earth system, we must make a number of assumptions in practice to simplify the computational model. However, these assumptions should not significantly alter the resulting computational SLE outcome, given the uncertainties associated with remote sensing techniques, data quality, and model accuracy. A discussion of the sensitivity of the conversion with regard to the assumed conditions is presented in a later section.

The first assumption is that the detected Antarctic mass changes are totally converted into sea-level change. This means that the ice-mass losses that do not directly discharge to the ocean are either adjusted before this conversion or will be neglected. For example, losses by evaporation are usually assumed to enter the ocean indirectly after a short residence in the atmosphere.

The second assumption is that the area of the global ocean is constant for the period of the estimated sea-level change, and the equivalent water calculated from the Antarctic mass change is evenly distributed over the ocean surface. This result is a simplification because the coastal terrain variation would change the calculated SLE as the sea level rises or declines, especially over a long period. Furthermore, the distribution of the amount of the equivalent water over the ocean can be affected by non-homogeneous undulations of the geoid, prevalent regional ocean currents, and other factors.

The third assumption is that the density of the water coming from the ice-sheet melting is constant when distributed over the global ocean surface. Sea water density varies with temperature and salinity. In our computation, we consider the fresh water from the ice-sheet mass change and simplify the water density, ρ_{w}, to 1000 $\mathrm{kg\,m^{-3}}$. We neglect the slight dilution of global-mean ocean salinity due to the addition of this fresh water.

The final assumption, which is not required for gravimetric measurements, is that the density of the changed mass (firn or ice) is constant during conversion from the ice-sheet volume change to the mass change. In Antarctica, the thickness of the firn layer, which is typically a few tens of meters, can reach over 100 m, but we assume that it is small in comparison to the underlying thickness of the ice layer that is typically 2,000 m and exceeds 4,000 m in many places. Within the vertical column of firn and ice the density varies on average between 320 $\mathrm{kg\,m^{-3}}$ and 360 $\mathrm{kg\,m^{-3}}$ for the firn layer. It is generally considered as 917 $\mathrm{kg\,m^{-3}}$ for ice (Kaspers et al. 2004). Previous estimates of the ice-mass change from the measured elevation change have followed diverse criteria. Some authors (Davis et al. 2005; Wingham et al. 2006) adopted an effective density ranging from 350 to 917 $\mathrm{kg\,m^{-3}}$. Kaspers et al. (2004) implemented a parameterization of surface snow density. Rignot et al. (2008) applied a firn-depth correction. Zwally et al. (2011) obtained the density from a firn-compaction model. A high degree of uncertainty lies in the conversion of volume to mass change due to inaccuracy of firn/ice density modeling and the spatial variation thereof (Li and Zwally 2011). Observations from most regions of Antarctica show that the vertical ice velocity (firn compaction and downward ice flow) is almost in balance with the long-term accumulation rate; the mean firn density can then be used because the change mainly occurs in the firn layer (McMillan et al. 2014). Therefore, in the following discussions, unless otherwise specified, we adopted the firn/ice density value of $\rho_{ice} = 400\ \mathrm{kg\,m^{-3}}$ used by Shepherd et al. (2012) simply to provide some computational examples. We understand that this effective density does not represent the situation of many regions in West Antarctica where significant melting takes place. This effective density is also not accurate for every thick firn/ice layer in inland regions where there is a need for a more sophisticated density model. In summary, detailed research on the firn/ice density is outside of the scope of this technical note.

The Conversion Method

Using the above assumptions, the volumetric change, ΔV_{ice}, resulting from n measurements of the ice-sheet elevation change, ΔH_k, at k locations with a surface area of $S_k (k = 1, \ldots, n)$ can be expressed as follows:

$$\Delta V_{ice} = \sum_{k=1}^{n} (S_k \times \Delta H_k). \tag{1}$$

Furthermore, the corresponding mass change, ΔM_{ice}, can be calculated, given the local density of changed mass ρ_k

$$\Delta M_{ice} = \sum_{k=1}^{n} (\rho_k \times S_k \times \Delta H_k). \tag{2}$$

If we use ΔH_{ice} to represent the average surface elevation change in the considered region with a total area of S_{ice}, Eq. (1) can be simplified to

$$\Delta V_{ice} = S_{ice} \times \Delta H_{ice}. \tag{3}$$

Similarly, Eq. (2) can be simplified given the average ice density, ρ_{ice}, of the region:

$$\Delta M_{ice} = \Delta V_{ice} \times \rho_{ice}. \tag{4}$$

Based on the assumption that the mass change of the ice sheet (ΔM_{ice}) is totally transformed into the equivalent mass change of water, the volume of the equivalent water can be obtained as follows:

$$\Delta V_w = \frac{\Delta M_{ice}}{\rho_w}, \tag{5}$$

where ρ_w is the density of water.

Finally, if the global ocean area S_o is given, the corresponding value of the SLE change, ΔH_{SLE}, caused by the ice-mass change can be estimated as follows:

$$\Delta H_{SLE} = \frac{\Delta M_{ice}}{S_o \times \rho_w} = \frac{\Delta V_{ice} \times \rho_{ice}}{S_o \times \rho_w}. \tag{6}$$

Computational Considerations and Implementation

Area of the Global Ocean

The area of the global ocean, S_o, is essential for computing sea-level changes due to exchanges of water between the ocean and the cryosphere. Different values of S_o have been adopted in its computation, and it is often difficult to trace these values back to authoritative sources. There has been little discussion of the impact of the accuracy of the factor on the outcome. As stated in Cogley (2012), there are two values of the global ocean area that are commonly used: 361×10^6 km^2, which is most likely derived from the work of Kossinna (1921), and 362×10^6 km^2, which is most likely derived from the work of Menard and Smith (1966). The main difference between the values is that Kossinna (1921) treated ice shelves as land, while Menard and Smith (1966) give special consideration to ice shelves. In fact, they should be included in the global ocean area because they are made up of floating ice, the equivalent water volume of which is already accounted for by the ocean and thus their changes do not contribute to sea-level change. Cogley (2012) corrected some previous estimates and evaluated the global ocean area to be 362.5×10^6 km^2.

Further investigation has been carried out in this study to estimate the global ocean area using the updated full-resolution global shorelines of GSHHG (Global Self-consistent, Hierarchical, High-resolution Geography Database), version 2.2.4 and version 2.3.1 (Wessel and Smith 1996; National Oceanic and Atmospheric Administration [NOAA] 2014). Because the earlier version of GSHHG database treats ice shelves as land, the area of Antarctic ice shelves, $S_{iceshelves}$, must be subtracted from the land area, S_{land}, when computing the impact of the Antarctic ice-mass change on the global sea level. The area of ocean can be calculated as follows:

$$S_o = S_{Earth} - (S_{land} - S_{ice\ shelves}). \tag{7}$$

For our computation, the WGS84 ellipsoid is used and the total area of the Earth surface, S_{Earth}, is $510,065,621.724$ km^2 based on the following equation:

$$S_{Earth} = 2\pi a^2 + \pi \left(\frac{b^2}{e}\right) \ln\left[\frac{1+e}{1-e}\right], \tag{8}$$

where $a = 6,378,137.0$m and $b = 6,356,752.314$ m are the semi major and minor axes, respectively, $e = \sqrt{1-(b/a)^2}$ is the first eccentricity.

According to Cogley (2012), to calculate the area of each polygon in the GSHHG database, the Lambert azimuthal equal-area projection was applied. The origin of the projection was at the centroid of the polygon. Based on version 2.2.4 of GSHHG, area of the land polygons of the full-resolution S_{land} is $149,060,109.978$ km^2, including the overall area of Antarctica. The area of ice shelves, $S_{ice\ shelves}$, was estimated by Cogley (2012) to be 1.561×10^6 km^2. This value is now updated as 1.595×10^6 km^2 in this study by a calculation based on a separate layer of Antarctic ice sheet in the latest version of GSHHG 2.3.1 that was derived from MODIS images.

Finally, our estimate of the global ocean area using Eq. 7 and GSHHG version 2.2.4 (released in November 2013) is 362.56×10^6 km^2. Using the latest GSHHG version 2.3.1(released in July 2014), the area of the Antarctic ice sheet S_{AIS} is 12.34×10^6 km^2 and the global ocean area S_o is 362.69×10^6 km^2. The updated values are used in the following computation. The difference between S_o and the earlier published value (362×10^6 km^2) is approximately 0.19%. The difference from the Cogley's estimate (362.5×10^6 km^2, based on GSHHG version 1.11) is approximately 0.05%. This difference is proven to have no significance in the conversion.

Conversion Scenarios

Eqs. (1)–(6) can be used in different ways depending on the given parameters and the estimates sought. If the entire AIS with an average ice thickness of 2,000 m would be melted, the global sea level would rise 62.39 m based on our conversion method. The following are a few scenarios in which some of the commonly sought estimates can be calculated based on the assumed conditions.

SLE from ice-mass change. Given an ice-mass change, ΔM_{ice}, that can be calculated from remote-sensing observations or the output of a modeling system, and the assumptions described in this technical note, this ice-mass change can be treated as a water-mass change equivalent, ΔM_w. Furthermore, if the global ocean area, S_o, and the water density, ρ_w, are given, we can estimate the SLE change ΔH_{SLE} using Eq. (6). For example, 500 Gt of ice-mass change commonly equals 1.4 mm SLE in the literature (Shum et al. 2008). If we use the updated global ocean area $S_o = 362.69 \times 10^6$ km^2 and the ocean water density $\rho_w = 1000$ kg m^{-3}, the ice-mass change of 500 Gt would contribute an SLE of 1.38 mm. Therefore, there is a difference of 0.02 mm in the estimated SLEs. As another example, the recently published annual change rate based on the reconciled mass balance estimate of the AIS for 1992 to 2011 is -71 ± 53 Gt yr^{-1} (Shepherd et al. 2012). Based on the above method, this mass loss rate would be equivalent to an SLE increase rate of 0.20 ± 0.15 mm yr^{-1}, which is similar to the observed 0.16 to 0.38 mm contributions of the AIS to global sea level rise between 1993 and 2010 (IPCC 2013).

From surface elevation change to ice-mass/SLE change. Satellite altimetry provides accurate remote sensing data of AIS surface elevation changes. If it is assumed that the elevation change can be simplified as an evenly distributed change over the entire AIS, the calculation of the corresponding mass change and the associated SLE can be made much simpler, and the resulting information may be used for quick and preliminary analysis. For instance, a homogeneous surface melt of 10 cm over the entire AIS would lead to an ice-mass loss of 493.6 Gt using Eqs. (3) and (4) and cause an SLE rise of 1.36 mm. This computation corresponds to the use of the entire area of Antarctica, except ice shelves, given the mass change area of $S_{AIS} = 12.34 \times 10^6$ km^2 and the density of the volume gained or lost from the ice sheet $\rho_{ice} = 400$ kg m^{-3}. From the opposite perspective, an ice-mass change of 100 Gt would be calculated as an AIS surface elevation change of 20.26 mm. Thus, the abovementioned ice-mass loss rate of –71 $\pm$ 53 Gt yr^{-1} (Shepherd et al. 2012) may be imagined as an evenly distributed elevation change rate of 14 $\pm$ 11 mm yr^{-1} on the AIS. Finally, an SLE rise of 1 mm corresponds to an ice-mass loss of –73.48 mm over the entire AIS.

From AIS margin change to ice-mass/SLE change. Another scenario is a uniform change of the AIS margin, which includes grounding lines where the ice flows across it into an ice shelf and ice-sheet boundary lines where ice flows directly into the ocean. Although the majority of the ice discharge takes place at glacier outlets, and some unstable grounding lines sitting on slopes directed inland can cause potential accelerated discharges, for computational simplicity we assume an unrealistic uniform retreat belt around the AIS. When the imagined uniform retreat rate is known, the retreat belt around the AIS can be calculated. We use an average ice thickness and an ice density of 917 kg m^{-3} to estimate the ice volume inside the retreat belt and the corresponding mass loss. Specifically, the AIS boundary data derived from the MODIS Mosaic of Antarctica (MOA) image maps (Haran et al. 2005), and the 1 km $\times$ 1 km resolution ice thickness map of Bedmap2 (Fretwell et al. 2013) are used. The AIS boundary line has a length of 63.948 $\times$ 10^3 km. The average ice thickness around the boundary is approximately 484 m. Based on these parameters, the ice-mass loss rate of –71 $\pm$ 53 Gt yr^{-1} (Shepherd et al. 2012) may be equivalent to an even AIS boundary retreat rate of 2.50 $\pm$ 1.87 m yr^{-1}. Moreover, an SLE rise of 1 mm could be caused by an even AIS boundary retreat of 12.78 m.

Finally, we would like to use the nominal accuracies of two relevant Antarctic missions, ICESat and GRACE, as examples to investigate the corresponding impact on the SLE changes. For the ICESat satellite laser altimeter, the nominal accuracy of the surface elevation measurements is $\pm$15 cm. Under the best conditions, the precision is better than $\pm$3 cm (Schutz et al. 2005). Based on the above analysis, the ice-mass changes corresponding to $\Delta H_{ice} = 15$ cm and 3 cm are 740.4 Gt and 148.1 Gt, respectively. These changes are also equivalent to SLE changes of 2.04 mm and 0.41 mm, respectively. It should be noted that the above calculations assume that the accuracy of 15 cm (or 3 cm) is evenly distributed over the entire continent of Antarctica. In fact, much of the interior has better conditions for altimetric measurements and should result in better accuracies, whereas the accuracy in coastal and high-slope regions may be worse than the nominal accuracy of 15 cm (Schutz et al. 2005; Gu et al. 2014). Furthermore, observations over a long period (multiple years) should provide much better accuracy.

To monitor mass changes in Antarctica using satellite gravity observations of the GRACE mission, a GIA model should be employed to correct for the regional vertical rebound of the bed of the Antarctic ice sheet. Currently, different GIA models produce different correction values that range from approximately 30 Gt yr^{-1} to 130 Gt yr^{-1} (Shepherd et al. 2012; Ivins et al. 2013). This uncertainty would introduce an SLE difference

Table 1
A look-up table of several computational examples of conversion between ice-mass and SLE changes

From	To
100 Gt of ice-mass change	0.28 mm SLE
100 Gt of ice-mass change	20.26 mm elevation change over the entire AIS
100 Gt of ice-mass change	3.52 m retrieval of the AIS margin*
1 mm SLE	362.69 Gt of ice-mass change
1 mm SLE	73.48 mm elevation change over the entire AIS
1 mm SLE	12.78 m retrieval of the AIS margin*
10 cm elevation change over the entire AIS	493.6 Gt of ice-mass change
10 cm elevation change over the entire AIS	1.36 mm SLE

*$\rho_{ice} = 917$ kg m^{-3} was used, elsewhere $\rho_{ice} = 400$ kg m^{-3}.

of 0.08 mm yr^{-1} to 0.36 mm yr^{-1}. Beside the firn/ice density, the uncertainty of GIA is another major source of errors in estimation of Antarctic mass change. More in-depth research on GIA models is expected in the future.

Table 1 lists several computational examples relevant to the conversion between the Antarctic ice-mass change and the SLE using the parameters discussed above and the assumptions of this technical note. It should be noted that we assume that the current mass balance in the AIS is maintained and the resulting SLE rise rate is a few millimeters per year. The global ocean area used above is still valid. Based on this constant global ocean area, the AIS would have the capacity to cause an SLE rise of approximately 62.39 m. However, if we consider the catastrophic scenario of complete melting of major outlet glaciers, for example in West Antarctica, or even the entire AIS, the global ocean area would dramatically grow. A precise computation of the impacted low land regions in the world would require a high resolution coastal DEM, ocean circulation modeling, tide modeling and other data. Such evaluation is outside of the scope of this technical note.

Discussion

The ice-mass imbalance of the Antarctic ice sheet contributes to global sea-level change. This process combines with other global and local phenomena of the planet and impacts our environment and lives. For example, a steadily increasing global sea level accompanied by increasingly frequent and stronger storms has caused coastal disasters with casualties and property losses. The quantification, understanding, and prediction of the SLE change from Antarctic ice-mass changes remain a complex problem that involves the analysis of a number of parameters that are currently determined with multiple uncertainties. This technical note discusses a simplified model for the conversion between Antarctic ice-mass changes and the SLE on the basis of a few generally accepted assumptions. The computational examples can be used as a quick reference in relevant general applications. The result should be helpful for science outreach as well. For glaciologists who undertake research in specific areas of Antarctica, a wide range of literature is available to provide more in-depth knowledge.

A limited sensitivity analysis was carried out to assess the influence of a few uncertain parameters on the conversion process. The accuracy of the area of the global ocean derived from the current global shoreline database is not a significant contributor to the overall uncertainty in this conversion process. However, the difference introduced by different GIA models is considered as the largest uncertainty and can reach a level that is close to the annual contribution of the AIS to the global sea level. Therefore, choosing an appropriate GIA model is critical to the analysis of satellite gravity data in the Antarctic region. Finally, the determination of an effective density of changed mass is very difficult because it depends on spatial location, types of firn layers, and the height change rate of firn/ice layers. The uncertainty introduced by the chosen effective ice density may be the second largest uncertainty in the conversion process. This situation may be improved after a large number of in-situ firn investigations are performed and a firn/ice modeling method coupled with these observations mitigates the situation in which there are insufficient observations.

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References

Cogley, J. G. 2012. Area of the ocean. *Marine Geodesy* 35(4):379–388.

Davis, C. H., Y. Li, J. R. McConnell, M. M. Frey, and E. Hanna. 2005. Snowfall-driven growth in East Antarctic ice sheet mitigates recent sea-level rise. *Science* 308(5730):1898–1901.

Fretwell, P., H. D. Pritchard, D. G. Vaughan, J. L. Bamber, N. E. Barrand, R. Bell, C. Bianchi, R. G. Bingham, D. D. Blankenship, G. Casassa, G. Catania, D. Callens, H. Conway, A. J. Cook, H. F.J. Corr, D. Damaske, V. Damm, F., Ferraccioli, R. Forsberg, S. Fujita, Y. Gim, P. Gogineni, J. A. Griggs, R. C.A. Hindmarsh, P. Holmlund, J. W. Holt, R. W. Jacobel, A. Jenkins, W. Jokat, T. Jordan, E. C. King, J. Kohler, W. Krabill, M. Riger-Kusk, K. A. Langley, G. Leitchenkov, C. Leuschen, B. P. Luyendyk, K. Matsuoka, J. Mouginot, F. O. Nitsche, Y. Nogi, O. A. Nost, S. V. Popov, E. Rignot, D. M. Rippin, A. Rivera, J. Roberts, N. Ross, M. J. Siegert, A. M. Smith, D. Steinhage, M. Studinger, B. Sun, B. K. Tinto, B. C. Welch, D. Wilson, D. A. Young, C. Xiangbin, and A. Zirizzotti. 2013. Bedmap2: Improved ice bed, surface and thickness datasets for Antarctica. *The Cryosphere* 7:375–393.

Gu, Z., T. Feng, M. Scaioni, H. Wu, J. Liu, X. Tong, and R. Li. 2014. Experimental results of elevation change analysis in the Antarctic ice sheet using DEMs from ERS and ICESat data. *Annals of Glaciology* 55(66):198–204.

Hanna, E., F. J. Navarro, F. Pattyn, C. M. Domingues, X. Fettweis, E. R. Ivins, R. J. Nichollos, C. Ritz, B. Smith, S. Tulaczyk, P. L. Whitehouse, and H. J. Zwally. 2013. Ice-sheet mass balance and climate change. *Nature* 498:51–59.

Haran, T., J. Bohlander, T. Scambos, T. Painter, and M. Fahnestock (compilers). 2005, updated 2006. *MODIS mosaic of Antarctica (MOA) image map*. Boulder, CO: National Snow and Ice Data Center. Digital media.

IPCC. 2013. *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Stocker, T. F.,

D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, and P. M. Midgley (eds.), Cambridge: Cambridge University Press.

Ivins, E. R., T. S. James, J. Wahr, E. J.O. Schrama, F. W. Landerer, and K. M. Simon. 2013. Antarctic contribution to sea level rise observed by GRACE with improved GIA correction. *Journal of Geophysical Research: Solid Earth* 118:3126–3141.

Kaspers, K. A., R. van de Wal, M. R. van den Broeke, J. Schwander, N. P. M. van Lipzig, and C. A. M. Brenninkmeijer. 2004. Model calculations of the age of firn air across the Antarctic continent. *Atmospheric Chemistry and Physics* 4:1365–1380.

Kossinna, E. 1921. *Die Tiefen des Weltmeeres. Veroeffentlichungen des Instituts fuer Meereskunde, 9*. Berlin: E.S. Mittler and Son.

Li, J. and H. J. Zwally. 2011. Modeling of firn compaction for estimating ice-sheet mass change from observed ice-sheet elevation change. *Annals of Glaciology* 52(59):1–7.

McMillan, M., A. Shepherd, A. Sundal, K. Briggs, A. Muir, A, Ridout, A. Hogg, and D. Wingham. 2014. Increased ice losses from Antarctica detected by Cryosat-2. *Geophysical Research Letters*. doi:10.1002/2014GL06011.

Menard, H. W. and S. M. Smith. 1966. Hypsometry of ocean basin provinces. *Journal of Geophysical Research* 71(18):4305–4325.

NOAA. 2014. *Homepage of GSHHG*. Available at http://www.ngdc.noaa.gov/mgg/shorelines/gshhs.html. Last accessed 15 July 2014.

Rignot E. and R. H. Thomas. 2002. Mass balance of polar ice sheets. *Science* 297:1502–1506.

Rignot E., J. L. Bamber, M. R. van den Broeke, C. Davis, Y. Li, W. J. van de Berg, and E. van Meijgaard. 2008. Recent Antarctic ice mass loss from radar interferometry and regional climate modeling. *Nature Geoscience* 1:106–110.

Schutz, B. E., H. J. Zwally, C. A. Shuman, D. Hancock, and J. P. DiMarzio. 2005. Overview of the ICESat mission. *Geophysical Research Letters* 32:L21S01.

Shepherd A., E. R. Ivins, A. Geruo, V. R. Barletta, M. J. Bentley, S. Bettadpur, K. H. Briggs, D. H. Bromwich, R. Forsberg, N. Galin, M. Horwath, S. Jacobs, I. Joughin, M. A. King, J. T. Lenaerts, J. Li, S. R. Ligtenberg, A. Luckman, S. B. Luthcke, M. McMillan, R. Meister, G. Milne, J. Mouginot, A. Muir, J. P. Nicolas, J. Paden, A. J. Payne, H. Pritchard, E. Rignot, H. Rott, L. S. Sørensen, T. A. Scambos, B. Scheuchl, E. J. Schrama, B. Smith, A. V. Sundal, J. H. van Angelen, W. J. van de Berg, M. R. van den Broeke, D. G. Vaughan, I. Velicogna, J. Wahr, P. L. Whitehouse, D. J. Wingham, D. Yi, D. Young, and H. J. Zwally. 2012. A reconciled estimate of ice-sheet mass balance. *Science* 338(6111):1183–1189.

Shum C. K., C. Y. Kuo, and J. Y. Guo. 2008. Role of Antarctic ice mass balance in present-day sea-level change. *Polar Science* 2:149–161.

Vaughan, D. G., J. C. Comiso, I. Allison, J. Carrasco, G. Kaser, R. Kwok, P. Mote, T. Murray, F. Paul, J. Ren, E. Rignot, O. Solomina, K. Steffen, and T. Zhang. 2013. Observations: Cryosphere. In *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Stocker, T. F., D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex, and P. M. Midgley (eds.), Cambridge: Cambridge University Press, 317–382.

Wessel, P. and W. H. F. Smith. 1996. A global, self-consistent, hierarchical, high-resolution shoreline database. *Journal of Geophysical Research* 101(B4):8741–8743.

Wingham, D. J., A. Shepherd, A. Muir, and G. J. Marshall. 2006. Mass balance of the Antarctic ice sheet. *Philosophical Transactions of the Royal Society of London, Series A* 364(1844):1627–1635.

Zwally, H. J. and M. B. Giovinetto. 2011. Overview and assessment of Antarctic ice-sheet mass balance estimates: 1992–2009. *Surveys in Geophysics* 32:351–376.

Zwally, H. J., L. Jun, A. C. Brenner, M. Beckley, H. G. Cornejo, J. Dimarzio, M. B. Giovinetto, T. A. Neumann, J. Robbins, J. L. Saba, Y. I. Donghui, and W. Wang, 2011. Greenland ice sheet mass balance: Distribution of increased mass loss with climate warming, 2003–07 versus 1992–2002. *Journal of Glaciology* 57(201):88–102.