Auxiliary Material for Paper 2012GL051466

Recent mass changes of glaciers in the Russian High Arctic

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The following sections provide additional details on the corrections and error analyses that were applied to the ICESat, GRACE and meteorological data sets. Two tables at the end show auxiliary mass budget estimates for individual glaciers and sub-regions (Table S1) and regional GRACE uncertainty components (Table S2).

ICESat corrections and error analysis

The ICESat GLAS determined surface elevations from ~70 m footprints spaced at ~170 m along each ground track [*Zwally et al.*, 2002]. The tracks were repeated 2-3 times a year within a cross-track distance of a few hundred meters. The single-shot elevation accuracy has been demonstrated to be better than 0.05 m under optimal conditions [*Fricker et al.*, 2005], but the performance degrades over sloping terrain [*Brenner et al.*, 2007] and under conditions favourable to atmospheric forward scattering and detector saturation [*Fricker et al.*, 2005]. A saturation range correction available in Release 531 was added to the elevations to account for the delay of the pulse center in saturated returns. We did not apply any parameter-based cloud filtering [e.g. *Smith et al.*, 2009] because it also tends to remove usable data which are essential to achieve a sufficient spatial and temporal data distribution in the Arctic [*Moholdt et al.*, 2010]. Instead, we removed potential outliers in the plane regression by running it iteratively until all elevation residuals were brought within a threshold of 5 m. Seasonal data filtering was also applied in the regression to ensure that the start and end season of each elevation change rate (*dh/dt*) were from the same time of year within ±30 days [*Moholdt et al.*, 2010]. Finally, we removed planes with insufficient data based on criteria of minimum 4 repeat-tracks or 10 data points, as well as a minimum data time span of 2 years. This resulted in an average time span of 4.8 years for the final *dh/dt* estimates.

From the remaining plane-filtered points, we estimated an elevation precision of 0.78 m over Russian Arctic glaciers based on the root-mean-square (RMS) error of 85 crossover points between ascending and descending tracks within individual observation campaigns (*dt* < 30 days). The elevation precision is best over Novaya Zemlya (0.50 m, 32 points) and worst over Franz Josef Land (0.97 m, 28 points), probably because of steeper surface slopes in the latter region (1.5° versus 3.3°). Similarly, we estimated a *dh/dt* precision of 0.26 m a^{-1} from 143 crossover planes. This random error includes uncertainties associated with the determination of surface slopes as well as different temporal sampling of non-linear elevation changes. Additional variation in *dh/dt* is due to spatial gradients in surface mass balance, firn compaction and glacier dynamics. Parts of this variation is typically correlated with elevation (*h*), so by parameterizing *dh/dt* as a function of *h*, one can reduce the uncertainty in spatial extrapolation of *dh/dt* to determine volume change rates (*dV/dt*). Higher order polynomial fitting followed by hypsometric extrapolation is a robust way to do this [*Kääb*, 2008; *Nuth et al.*, 2010]. Each region had an adequate sampling of *dh/dt* across the full elevation range, so the polynom order had very little influence on the end results [*Gardner et al.*, 2011]. We used third order polynomial fits [*Moholdt et al.*, 2010], resulting in RMS errors in the range 0.5-0.6 m a^{-1}. This error reduces after spatial averaging over distances larger than the measurement correlation length. We analysed semivariograms of the difference between *dh/dt* and *dh/dt(h)* to estimate regional correlation lengths of 3.4-6.7 km. We then calculated standard errors (σ_{STDE}) of 0.02-0.03 m a^{-1} for the area-averaged *dh/dt* in each region based on the RMS of the polynomial fits and an average correlation length of 5 km.

In addition to the random errors, we also need to account for potential systematic biases in the observation system and the spatiotemporal sampling. Elevation biases of up to ±0.10 m have been detected for individual ICESat observations campaigns [*Siegfried et al.*, 2011; *Zwally et al.*, 2011], but uncertainty remains whether there is a significant temporal trend in the inter-campaign biases. Crustal uplift may also induce a bias in *dh/dt* estimates. The average uplift rates of the Barents Sea archipelagos are however smaller than 0.002 m a^{-1} [*Forman et al.*, 2004]. Given the small magnitudes and high uncertainty of these potential biases, we did not make any corrections to our *dh/dt* estimates, but instead added an error (σ_{BIAS}) of 0.02 m a^{-1} to the area-averaged *dh/dt* for each region.

Errors in regional *dV/dt* estimates can also arise from uneven spatial and temporal sampling of *dh/dt*. We tested for random spatial biases by doing separate calculations for ascending and descending tracks, which are spatially independent from each other. Ascending and descending results were within 0.02 m a^{-1} of the area-averaged *dh/dt* in each region, which indicate that the number of ICESat tracks is sufficient for spatial extrapolation and that our estimates of σ_{STDE} are realistic. Systematic spatial biases (σ_{SPAT}) can occur if certain parts of glaciers or regions are systematically under-sampled due to more cloud cover, steeper surface slopes or heavier crevassing. We investigated this issue by doing sub-regional *dV/dt* calculations (Table S1) and summing them together to get regional *dV/dt* estimates. Again, we found a correspondence to within 0.02 m a^{-1} of our original *dV/dt* estimates. Note that many ICESat tracks have a complete coverage of *dh/dt* planes, which reduces the likelihood of spatial biases related to meteorology and topography.

Temporal biases in *dh/dt* occur when certain ICESat campaigns have more data than others due to variable cloud cover and campaign duration. The problem can be avoided by calculating campaign-wise or annual *dV* estimates and summing them up for the entire ICESat period [*Gardner et al.*, 2011]. This is however done at the cost of spatial coverage, i.e. spatial biases in individual *dV* estimates may exceed biases in the overall *dV/dt* estimates. We consider the campaign-to-campaign data coverage in the Russian Arctic to be too sparse to follow such an approach and hence only use these data to illustrate temporal variations in area-averaged *dh* (Fig. 2). The average amount of ICESat points that are accepted following the plane regressions is 19,000 ± 7000 for each observation campaign, apart from the October 2009 campaign which has only 1200 points due to the failure of the last laser onboard ICESat. This campaign is therefore under-represented in the *dh/dt* estimates and not included in Fig. 2. Otherwise, there are no clear temporal patterns in the sampling except that winter campaigns generally have more data than other campaigns due to less cloud cover. To account for potential biases related to temporal irregularities, we added an error (σ_{TEMP}) of 0.02 m a^{-1} to the area-averaged *dh/dt* in each region. This number is based on alternative *dV/dt* calculations for the Canadian Arctic using the above mentioned methods [*Gardner et al.*, 2011].

Errors in glacier areas used for extrapolation of *dh/dt* can also cause a bias in *dV/dt.* We expect this error (σ_{AREA}) to be well within ±5% since all glacier outlines were digitized from satellite imagery of the same decade as the ICESat measurements. An additional error is introduced when *dV/dt* is converted to *dM/dt* using an assumed density conversion factor of 0.9 Gt km^{-3}. Unfortunately, there is no available data on firn pack changes in the Russian Arctic from the last decade. The meteorological anomalies for 2003-2009 indicate a slight increase in precipitation, but that is likely compensated by slightly higher summer temperatures. In order to account for this uncertainty, we take the regional volume-to-mass conversion error (σ_{DENS}) to be the greatest of ±10% of *dV/dt* or a constant of 0.5 Gt a^{-1}. This error is likely on the conservative side. A recent study of firn cores from the summit of the Penny Ice Cap on Baffin Island found that the average density of the upper 20 m of the cores changed by only 6% between 1995 and 2010 despite a firn warming of about 10 ºC at 10 m depth [*Zdanowicz et al.*, 2012]. Also, most of the volume loss in the Russian Arctic has occurred at low elevations in Novaya Zemlya where ice melting and glacier dynamics should be the dominating processes.

Finally, we converted all these errors into mass equivalent rates (Gt a^{-1}) and combined them as root sum of squares (RSS) to obtain regional mass budget errors:

$$\sigma_{MB} = \sqrt{\sigma_{STDE}^2 + \sigma_{BIAS}^2 + \sigma_{SPAT}^2 + \sigma_{TEMP}^2 + \sigma_{AREA}^2 + \sigma_{DENS}^2}$$

The total error for the Russian Arctic was calculated from the RSS of the regional σ_{MB} apart from σ_{BIAS} which we added cumulatively to account for its high spatial correlation.

Auxiliary mass budget estimates from ICESat

We also estimated mass budgets and errors for smaller sub-regions where the spatial distribution of ICESat tracks was sufficient. Table S1 summarizes the results for some of the major ice caps and icefields within the larger regions.

GRACE corrections and error analysis

The GRACE data (CSR RL04) were processed as by *Gardner et al.* [2011]: The missing degree-1 coefficients were included following *Swenson et al.* [2008], and the C20 coefficients were replaced by values derived from satellite laser ranging (R. J. Eanes, personal communication, 2008). Data noise was reduced by applying the post-processing method of *Wouters and Schrama* [2007] and a Gaussian smoother with a half-width radius of 250 km. The gravitational effect from Glacial Isostatic Adjustment (GIA) was corrected using a modified version of the ICE-5G (VM2) ice loading history and Earth viscosity model [*Ivins and James*, 2005; *Peltier*, 2004; *Riva et al.*, 2010]. The applied GIA corrections for Franz Josef Land, Severnaya Zemlya and Novaya Zemlya were -1.45 Gt a^{-1}, -1.45 Gt a^{-1} and -0.25 Gt a^{-1}, respectively.

To reduce leakage from non-glacial signals in adjacent areas, we subtracted water storage from the GLDAS-NOAH model [*Rodell et al.*, 2004] in its 0.25 degree configuration, where the glaciated areas of the Russian Arctic were first masked out. Variations in ocean mass were already removed in the level-1 processing by the GRACE science team, using the Ocean Model for Circulation and Tides [*Flechtner*, 2007]. Leakage from remaining variability was suppressed by simultaneously fitting surface mass anomalies to the neighboring regions. The outlines of these basins on land were based on the major river-catchments in the Russian Arctic (such as Pechora, Ob, Yenisey and Lena), determined from the TRIP data set [*Oki and Sud*, 1998], and on bathymetry and the variance of the GRACE signal in the ocean. To investigate if regional mass change signals were properly separated, we calculated correlation coefficients between mass anomaly time series of glaciated and surrounding regions. A positive correlation is expected if the GRACE resolution is too low to isolate the glacier signal. If the iterative basin method does not correctly separate the signal, then compensation will take place between adjacent basins such that the correlation becomes negative. The obtained correlations were typically lower than ±0.2, which is within the range that we can expect from random chance and climatological covariability.

The uncertainty of the applied GRACE data product (CSR RL04) was estimated by comparing it with an independent GRACE solution (GFZ RL04). The GIA uncertainty was derived from a range of realistic viscosity profiles (0.3×10^{21} to 1.0×10^{21} Pa s and 0.3×10^{21} to 1.0×10^{22} Pa s for the upper and lower mantle, respectively) and alternative loading histories from the models ICE-3G [*Tushingham and Peltier*, 1991] and ANU [*Lambeck et al.*, 2004]. We assessed the hydrology correction by replacing the GLDAS-NOAH model by the WGHM model [*Döll et al.*, 2003; 2011]. Simulations have shown that the iterative optimization method of *Wouters et al.* [2008] retrieves regional mass change rates within an 1σ error of 1-1.5 Gt a^{-1} [*Gardner et al.*, 2011]. In addition, we accounted for the standard error of the linear fits to the monthly mass anomalies. All 5 error terms are provided for each region in Table S2. Regional mass budget errors were determined as the RSS of the individual errors. The smoothed mass budget curves in Fig. 3 were not used in any of the calculations, but would have given similar results if they had been used as the basis for the linear fits instead of the unsmoothed monthly values.

Meteorological error analysis

Errors in the climatic anomalies are difficult to assess. Climate conditions at meteorological stations might not be representative for entire glacier regions, and climate reanalysis data might contain significant biases. We have done a simple error analysis based on comparisons between different data sets. The errors of the glacier area-averaged precipitation anomalies (Table 1) were estimated directly from the standard deviation of the three applied precipitation products (Section 2.3). For the summer (JJA) temperature anomalies at meteorological stations, we compared them with "free-air" anomalies at the 700 mb geopotential height of two reanalysis data sets: the NCEP Climate Forecast

System Reanalysis [*Saha et al.*, 2010] and the ERM-Interim reanalysis [*Dee et al.*, 2011]. We used the 700 mb geopotential height instead of the 2 m height to limit the influence of sea surface temperatures in the glacier area-averaged anomalies. The resulting temperature anomalies and standard deviations for the reanalysis data in 2004-2009 are 0.13 ± 0.22 °C for Franz Josef Land, 0.17 ± 0.13 °C for Severnaya Zemlya, and 0.25 ± 0.16 °C for Novaya Zemlya. These values are within 0.25 °C of the station-derived anomalies. We expect the station measurements to be more reliable than the reanalysis products, so we used the RMS difference with respect to the reanalysis temperature anomalies to estimate the error of the station anomalies (Table 1).

References

Brenner, A. C., J. R. DiMarzio, and H. J. Zwally (2007), Precision and accuracy of satellite radar and laser altimeter data over the continental ice sheets, *IEEE Trans. Geosci. Remote*, *45*(2), 321-331.

Dee, D. P., et al. (2011), The ERA-Interim reanalysis: configuration and performance of the data assimilation system, *Q. J. Roy. Meteor. Soc.*, *137*(656), 553-597.

Döll, P., F. Kaspar, and B. Lehner (2003), A global hydrological model for deriving water availability indicators: Model tuning and validation, *J. Hydrol.*, *270*(1-2), 105-134.

Döll, P., H. Hoffmann-Dobrev, F. T. Portmann, S. Siebert, A. Eicker, M. Rodell, G. Strassberg, and B. Scanlon (2011), Impact of water withdrawals from groundwater and surface water on continental water storage variations, *J. Geodyn.*, doi:10.1016/j.jog.2011.1005.1001.

Flechtner, F. (2007), AOD1B Product Description Document for Product Releases 01 to 04, GRACE 327-350, rev. 3.1, *Rep.*, GZF, Potsdam, Germany.

Forman, S. L., D. J. Lubinski, O. Ingolfsson, J. J. Zeeberg, J. A. Snyder, M. J. Siegert, and G. G. Matishov (2004), A review of postglacial emergence on Svalbard, Franz Josef Land and Novaya Zemlya, northern Eurasia, *Quaternary Sci. Rev.*, *23*(11-13), 1391-1434.

Fricker, H. A., A. Borsa, B. Minster, C. Carabajal, K. Quinn, and B. Bills (2005), Assessment of ICESat performance at the Salar de Uyuni, Bolivia, *Geophys. Res. Lett.*, *32*(L21S06), doi:10.1029/2005GL023423.

Gardner, A. S., G. Moholdt, B. Wouters, G. J. Wolken, D. O. Burgess, M. J. Sharp, J. G. Cogley, C. Braun, and C. Labine (2011), Sharply increased mass loss from glaciers and ice caps in the Canadian Arctic Archipelago, *Nature*, *473*(7347), 357-360.

Ivins, E. R., and T. S. James (2005), Antarctic glacial isostatic adjustment: A new assessment, *Antarct. Sci.*, *17*(4), 541-553.

Kääb, A. (2008), Glacier Volume Changes Using ASTER Satellite Stereo and ICESat GLAS Laser Altimetry. A Test Study on Edgeoya, Eastern Svalbard, *IEEE Trans. Geosci. Remote*, *46*(10), 2823-2830.

Lambeck, K., F. Antonioli, A. Purcell, and S. Silenzi (2004), Sea-level change along the Italian coast for the past 10,000 yr, *Quaternary Sci. Rev.*, *23*(14-15), 1567-1598.

Moholdt, G., C. Nuth, J. O. Hagen, and J. Kohler (2010), Recent elevation changes of Svalbard glaciers derived from ICESat laser altimetry, *Remote Sens. Environ.*, *114*(11), 2756-2767.

Nuth, C., G. Moholdt, J. Kohler, J. O. Hagen, and A. Kääb (2010), Svalbard glacier elevation changes and contribution to sea level rise, *J. Geophys. Res.*, *115*(F01008), doi:10.1029/2008JF001223.

Oki, T., and Y. Sud (1998), Design of Total Runoff Integrating Pathways (TRIP) - A global river channel network, *Earth Interact.*, *2*, 1-36.

Peltier, W. R. (2004), Global glacial isostasy and the surface of the ice-age Earth: The ICE-5G (VM2) model and GRACE, *Annu. Rev. Earth Pl. Sc.*, *32*, 111-149.

Riva, R. E. M., J. L. Bamber, D. A. Lavallee, and B. Wouters (2010), Sea-level fingerprint of continental water and ice mass change from GRACE, *Geophys. Res. Lett.*, *37*(L19605), doi:10.1029/2010gl044770.

Rodell, M., et al. (2004), The Global Land Data Assimilation System, *B. Am. Meteorol. Soc.*, *85*(3), 381-394.

Saha, S., et al. (2010), The NCEP Climate Forecast System Reanalysis, *B. Am. Meteorol. Soc.*, *91*(8), 1015-1057.

Siegfried, M. R., R. L. Hawley, and J. F. Burkhart (2011), High-Resolution ground-based GPS measurements show intercampaign bias in ICESat elevation data near summit, Greenland, *IEEE T. Geosci. Remote*, *49*(9), 3393-3400.

Smith, B. E., H. A. Fricker, I. R. Joughin, and S. Tulaczyk (2009), An inventory of active subglacial lakes in Antarctica detected by ICESat (2003-2008), *J. Glaciol.*, *55*(192), 573-595.

Swenson, S., D. Chambers, and J. Wahr (2008), Estimating geocenter variations from a combination of GRACE and ocean model output, *J. Geophys. Res.*, *113*(B08410), doi:10.1029/2007jb005338.

Tushingham, A. M., and W. R. Peltier (1991), ICE-3G - A new global model of late pleistocene deglaciation based upon geophysical predictions of postglacial relative sea-level change, *J. Geophys. Res.*, *96*(B3), doi:10.1029/1090jb01583.

Wouters, B., and E. J. O. Schrama (2007), Improved accuracy of GRACE gravity solutions through empirical orthogonal function filtering of spherical harmonics, *Geophys. Res. Lett.*, *34*(L23711), doi:10.1029/2007gl032098.

Wouters, B., D. Chambers, and E. J. O. Schrama (2008), GRACE observes small-scale mass loss in Greenland, *Geophys. Res. Lett.*, *35*(L20501), doi:10.1029/2008GL034816.

Zdanowicz, C., A. Smetny-Sowa, D. A. Fisher, N. Schaffer, L. Copland, F. J. Eley, and F. Dupont (2012), Summer melt rates on Penny Ice Cap, Baffin Island: Past and recent trends, and implications for regional climate, *J. Geophys. Res.*, *doi:10.1029/2011JF002248, in press.*

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

Zwally, H. J., et al. (2002), ICESat's laser measurements of polar ice, atmosphere, ocean, and land, *J. Geodyn.*, *34*(3-4), 405-445.

Tables

Table S1. Sub-regional mass budget estimates for 2004-2009 based on ICESat

Glacier name (sub-region)	Region	Latitude (°N)	Longitude (°E)	Glacier area (km^2)	Mass budget ($kg\ m^{-2}\ a^{-1}$)
Vostok-3/4 Ice Cap	FJL	81.65	63.09	260	170 ± 100
Vostok-2 Ice Cap	FJL	81.35	58.93	130	50 ± 130
Vostok-1 Ice Cap	FJL	81.00	60.88	390	50 ± 120
Windy Ice Cap	FJL	80.82	64.00	1,170	380 ± 70
Moon Ice Cap	FJL	80.60	46.34	620	-70 ± 140
Tyndall Ice Cap	FJL	80.63	60.79	1,810	-120 ± 80
Prince George Land, icefield	FJL	80.46	49.25	2,050	-160 ± 70
Moscow Ice Cap	FJL	80.01	59.42	850	-60 ± 90
Ushakov Ice Cap	-	80.82	79.47	320	100 ± 70
Schmidt Ice Cap	SZ	81.13	90.90	410	70 ± 70
Academy of Sciences Ice Cap	SZ	80.49	94.99	5,570	-190 ± 70
Rusanov Ice Cap	SZ	79.97	97.10	1,010	-30 ± 60
Albanov Ice Cap	SZ	79.88	95.35	350	-20 ± 70
Karpinsky Ice Cap	SZ	79.59	98.58	2,340	-120 ± 60
Vavilov Ice Cap	SZ	79.32	95.69	1,780	50 ± 60
University Ice Cap	SZ	79.04	98.93	1,690	70 ± 50
Bolshevik Island, icefield	SZ	78.69	102.88	2,140	-10 ± 70
Kropotkin Ice Cap	SZ	78.40	101.44	260	-280 ± 80
Severny Isl. Ice Cap (Barents)	NZ	75.86	61.83	10,010	-410 ± 70
Severny Isl. Ice Cap (Kara)	NZ	75.61	62.08	9,480	-220 ± 50
NZ, other glaciers	NZ	-	-	2,620	-580 ± 140

Table S2. GRACE 1σ uncertainties ($Gt\ a^{-1}$) based on period Oct. 2003 – Oct. 2009

Glacier region	GRACE product[a]	GIA correction[b]	Hydrology correction[c]	Iterative optimization	Linear fits	Total RSS
Franz Josef Land	2.7	0.9	0.1	1.5	1.2	3.4
Severnaya Zemlya	1.2	0.9	1.2	1.5	1.2	3.0
Novaya Zemlya	1.7	1.4	1.2	1.5	1.4	3.0

[a] Difference between GRACE solutions CSR RL04 and GZF RL04, [b] RMS between GIA estimates from different mantle viscosity profiles and loading models, [c] difference between GLDAS-NOAH and WGHM hydrology models.