The future sea-level rise contribution of Greenland’s glaciers and ice caps

Machguth, H.; Rastner, P.; Bolch, T.; Mölg, N.; Sørensen, Louise Sandberg; Aðalgeirsdottir, G.; van Angelen, J. H.; van den Broeke, M. R.; Fettweis, X.

Published in:
Environmental Research Letters

Link to article, DOI:
10.1088/1748-9326/8/2/025005

Publication date:
2013

Document Version
Publisher's PDF, also known as Version of record

Link back to DTU Orbit

Citation (APA):
Machguth, H., Rastner, P., Bolch, T., Mölg, N., Sørensen, L. S., Aðalgeirsdottir, G., ... Fettweis, X. (2013). The future sea-level rise contribution of Greenland’s glaciers and ice caps. Environmental Research Letters, (8), 025005. https://doi.org/10.1088/1748-9326/8/2/025005
The future sea-level rise contribution of Greenland’s glaciers and ice caps

This article has been downloaded from IOPscience. Please scroll down to see the full text article.
2013 Environ. Res. Lett. 8 025005
(http://iopscience.iop.org/1748-9326/8/2/025005)

View the table of contents for this issue, or go to the journal homepage for more

Download details:
IP Address: 130.208.103.74
The article was downloaded on 12/04/2013 at 11:01

Please note that terms and conditions apply.
The future sea-level rise contribution of Greenland’s glaciers and ice caps

H Machguth1,2, P Rastner1, T Bolch1,3, N Mölg1, L Sandberg Sørensen4, G Aðalgeirsdottir5, J H van Angelen6, M R van den Broeke6 and X Fettweis7

1 Department of Geography, University of Zurich, Switzerland
2 Geological Survey of Denmark and Greenland (GEUS), Copenhagen, Denmark
3 Institute for Cartography, Technical University Dresden, Germany
4 Technical University of Denmark, Kongens Lyngby, Denmark
5 Danish Meteorological Institute (DMI), Copenhagen, Denmark
6 Institute for Marine and Atmospheric Research (IMAU), Utrecht, The Netherlands
7 Department of Geography, University of Li`ege, Li`ege, Belgium

E-mail: horst.machguth@geo.uzh.ch

Received 31 July 2012
Accepted for publication 13 March 2013
Published 11 April 2013
Online at stacks.iop.org/ERL/8/025005

Abstract

We calculate the future sea-level rise contribution from the surface mass balance of all of Greenland’s glaciers and ice caps (GICs, ∼90 000 km²) using a simplified energy balance model which is driven by three future climate scenarios from the regional climate models HIRHAM5, RACMO2 and MAR. Glacier extent and surface elevation are modified during the mass balance model runs according to a glacier retreat parameterization. Mass balance and glacier surface change are both calculated on a 250 m resolution digital elevation model yielding a high level of detail and ensuring that important feedback mechanisms are considered. The mass loss of all GICs by 2098 is calculated to be 2016 ± 129 Gt (HIRHAM5 forcing), 2584 ± 116 Gt (RACMO2) and 3907 ± 108 Gt (MAR). This corresponds to a total contribution to sea-level rise of 5.8 ± 0.4, 7.4 ± 0.3 and 11.2 ± 0.3 mm, respectively. Sensitivity experiments suggest that mass loss could be higher by 20–30% if a strong lowering of the surface albedo were to take place in the future. It is shown that the sea-level rise contribution from the north-easterly regions of Greenland is reduced by increasing precipitation while mass loss in the southern half of Greenland is dominated by steadily decreasing summer mass balances. In addition we observe glaciers in the north-eastern part of Greenland changing their characteristics towards greater activity and mass turnover.

Keywords: Greenland, glaciers and ice caps, sea level rise contribution, climate model output, glacier retreat parameterization

1. Introduction

Glaciers and ice caps (GICs) of most regions in the world are currently undergoing strong changes. Retreat and mass loss are also observed on the GICs of Greenland (e.g. Knudsen and Hasholt 2008, Bolch et al 2013). The recently published glacier inventory of Greenland (Rastner et al 2012) revealed that the surface area of GICs amounts to ∼90 000 km² (∼12% of the total GICs on earth), considerably more than previously estimated. Hence there is a considerable potential for sea-level rise that has so far received limited attention compared to other Arctic regions such as Svalbard (e.g. Nuth et al 2010) or the Canadian Arctic (e.g. Gardner et al 2011).

In the majority of studies focusing on future changes of GICs, these are treated as abstracted and simplified...
objects. For instance Raper and Braithwaite (2006) address GICs as size distributions of glacier areas and volumes in 1° latitude/longitude cells. Radić and Hock (2011) approximate the hypsometry of mountain glaciers by linearly increasing the area per elevation-band from zero at the terminus to a maximum at the mean altitude and a linear decrease above. Such abstractions of glaciers allow for a computationally efficient handling of very large glacier samples. However, it is difficult to assess to what degree abstract representations of glaciers are representative for real conditions.

These issues motivate the development of modeling approaches aiming at a more realistic representation of large glacier samples including important mechanisms such as the feedback of glacier thinning on glacier mass balance. The impact of this feedback process strongly depends on the hypsometry of each individual glacier: flat glaciers can lose considerable portion of their accumulation area due to surface lowering while steep glaciers are less sensitive to this effect (Jiskoot et al. 2009).

The above considerations point out the importance of taking characteristics of individual glaciers into account when modeling future extent and volume changes. On GICs a high level of detail is required for accurate representation of complex glacier topography and surface mass balance distribution. Sufficient input data for comprehensive modeling of ice dynamics and mass balance is only available for a few well studied (and mostly small) glaciers (e.g. Zwinger and Moore 2009). For larger glacier samples more simple approaches are applied: Huss et al. (2008, 2010a), Salzmann et al. (2012) combined glacier mass balance models with the so called ‘Δh glacier retreat parameterization’ (Huss et al. 2010b) to model future glacier extent and volume of glacierized catchments in the Swiss Alps.

In the present study we apply a similar approach as in Salzmann et al. (2012) to calculate future scenarios of Greenland’s GICs from the year 2000 to 2098 and their contribution to sea-level rise. We combine DEMs, satellite derived glacier inventory data, surface elevation changes measured from ICESat, mass balance and ice thickness measurements as well as gridded climate model output to achieve a more realistic representation of Greenland’s GICs in future-scenario calculations. Glacier surface mass balance and glacier retreat scenarios are computed for six selected regions and ultimately upscaled to all of Greenland’s GICs.

2. Study area and data

The newly established Greenland glacier inventory (Rastner et al. 2012) was used as baseline information of glacier extents. The inventory includes all GICs on Greenland at a spatial resolution of 30 m and is derived from Landsat scenes acquired mostly in between 1999 and 2002. GICs are divided into three classes of connectivity to the ice sheet: CL0 (no connection), CL1 (weakly connected) and CL2 (strongly connected). CL0 and CL1 glaciers are glaciers dynamically independent of the ice sheet, CL2 ice bodies are of local character but dynamically not independent from the ice sheet. According to Rastner et al. (2012), the area of Greenland’s GICs amounts to ~90 000 km² (CL0 and CL1) or ~130 000 km² (all three connectivity classes). The area of CL0 and CL1 glaciers amounts to ~150% of earlier estimates (e.g. Weidick and Morris 1998, Radić and Hock 2011).

Six regions of Greenland were selected for detailed modeling of glacier mass balance and glacier retreat (table 1 and figure 1). The regions were chosen to represent the different climatic regions of Greenland and to achieve an optimum availability of input and validation data for the modeling. All of the six regions include glaciers with mass balance observations, ice thickness measurements and meteorological stations of the Danish Meteorological Institut (DMI): (1) the southernmost tip of Greenland (henceforward called ‘South’), (2) the Sukkertoppen area in the south-west dominated by two larger ice caps (‘Sukkertoppen’), (3) a region in the south-east including Mittivakkat glacier, Greenland’s only GIC with a longer-term mass balance series (‘Mittivakkat’), (4) the Stauning Alper in the central east comprising rugged mountain terrain with valley glaciers but also ice caps (‘Stauning’), (5) the area around the Zackenberg research station in the north-east (‘Zackenberg’) and (6) in the far north the area around Hans Tausen ice cap (‘North’).

In total these six regions include 17 660 km² of GICs, corresponding to 20% of all CL0 and CL1 glaciers in Greenland.
For all regions we used the 90 m resolution DEM (down-sampled to 250 m resolution) of the Greenland Ice Mapping Project (Howat et al. 2013), currently one of the best DEM available for Greenland (Rastner et al. 2012). The mass balance model is driven from gridded climate model output and we use regional climate model (RCM) data from three different sources: (1) a ~25 km resolution run of HIRHAM5 (e.g. Ádalgeirsdóttir et al 2009) forced by ECHAM5 (A1B scenario) at the boundaries, (2) a ~25 km resolution run of MAR (e.g. Fettweis 2007) (identical forcing as HIRHAM5), and (3) an ~11 km resolution RACMO2 run (e.g. Angelen et al. 2012), forced by HadGEM2 under the RCP4.5 scenario. All three model runs cover the whole of Greenland and are available for the years 1980–2098 at a temporal resolution of one day. Data from all DMI weather stations on Greenland are available for validation of the RCM data and for the correction of biases therein. The data (Boas and Wang 2011) cover the time period 1958–2010 and include 81, almost exclusively coastal, stations. However, numerous stations have been operating only for a limited period in the past.

While Mittivakkat glacier in south eastern Greenland is the only glacier with continued longer-term mass balances (1995–now) (cf Knudsen and Hasholt 2008) a large number of short-term mass balance observations exist (cf figure 1) and were obtained from publications (e.g. Hammer 2001, Ahlstrøm et al 2007), technical reports (e.g. Clement 1982), or gathered from the archives at the Geological Survey of Denmark and Greenland (GEUS), Copenhagen. Finally, measurements of ice thickness are available for a number of larger ice caps and valley glaciers on Greenland from the IceBridge project carried out by NASA (e.g. Gogineni et al 2001) and from other campaigns (e.g. Johnsen et al. 1992, Hammer 2001, Cittero and Mottram 2008) (cf figure 1).

### Table 1. Overview of the six selected regions and the central region specific model calibration parameters. $P_{\text{corr}}$ values vary depending on whether $\sum B$, minimum, maximum or intermediate is used in calibration. For simplicity only values of $P_{\text{corr}}$ for the intermediate calibration are given. A dash (‘—’) indicates that no correction was applied.

| Region          | North | Zackenberg | Stuuning | Mittivakkat | South | Sukkertoppen |
|-----------------|-------|------------|----------|-------------|-------|--------------|
| $h_{\text{max}}$ (m) | 400   | 350        | 500      | 800         | —     | 500          |
| $C_0$ (W m$^{-2}$) | $-45$ | $-45$      | $-45$    | $-45$       | $-45$ | $-45$        |
| $C_1$ (W m$^{-2}$ °C$^{-1}$) | 12    | 12         | 12       | 14          | 14    | 12           |
| $\alpha$        | 0.4   | 0.4        | 0.4      | 0.35        | 0.35  | 0.4          |
| $\sum B$, minimum (m w.e.) | $-1.3$ | $-4.4$    | $-2.9$   | $-10$       | $-5.1$ | $-3.6$        |
| $\sum B$, intermediate (m w.e.) | $-1.8$ | $-6.1$    | $-4.1$   | $-14$       | $-7.9$ | $-5$          |
| $\sum B$, minimum (m w.e.) | $-2.3$ | $-7.8$    | $-5.3$   | $-18$       | $-10.7$ | $-6.4$        |
| Bias correction T | Yes   | Yes        | No       | Yes         | Yes   | Yes          |
| RACMO2 mean $P_{\text{corr}}$ | 1.06  | 1.01       | 1.02     | 0.83        | 0.84  | 1.18         |
| HIRHAM5 mean $P_{\text{corr}}$ | 0.93  | 0.89       | 1.12     | 1.13        | 0.84  | 0.88         |
| MAR mean $P_{\text{corr}}$ | 0.94  | 1.14       | 0.82     | 1.49        | 0.79  | 0.88         |
| RACMO2 $T_{\text{offset}}$ (K) | —     | —          | $-0.5$   | 0.75        | —     | $-1.0$       |
| HIRHAM5 $T_{\text{offset}}$ (K) | —     | —          | $-0.5$   | $-1.5$      | 0.5   | —            |
| MAR $T_{\text{offset}}$ (K) | —     | —          | 1.5      | $-1.5$      | 1.5   | —            |

3. Mass balance modeling

Mass balance distribution for all glaciers of the six selected regions (section 2) is computed at a spatial resolution of 250 m. The applied glacier mass balance model is a simplified version of more sophisticated energy balance approaches. Here we briefly summarize the model as a detailed description is given by Machguth et al (2009).

The model runs at daily steps, and the cumulative mass balance $b_c$ on day $t+1$ is calculated for every time step and over each grid cell of the DEM according to Oerlemans (2001):

$$b_c(t+1) = b_c(t) + \left\{ \frac{\Delta t \cdot (-Q_m)}{l_m + P_{\text{solid}}} \right\} \text{if } Q_m > 0$$

$$= \left\{ \frac{\Delta t \cdot (-Q_m)}{P_{\text{solid}}} \right\} \text{if } Q_m \leq 0$$

where $t$ is the discrete time variable, $\Delta t$ is the time step (one day), $l_m$ is the latent heat of fusion of ice (334 kJ kg$^{-1}$) and $P_{\text{solid}}$ is solid precipitation in meter water equivalent (m w.e.). The energy available for melt ($Q_m$) is calculated as follows:

$$Q_m = (1 - \alpha)S_{\text{in}} + C_0 + C_1 T$$

where $\alpha$ is the albedo of the surface, $T$ is the air temperature (in °C at 2 m above ground and outside the glacier boundary layer), and $C_0 + C_1 T$ is the sum of the long-wave radiation balance and the turbulent exchange linearized around the melting point (Oerlemans 2001). Global radiation ($S_{\text{in}}$) is calculated from potential clear sky global radiation ($S_{\text{in,clr}}$) and fractional cloud cover ($n$). Both the direct and diffuse part of $S_{\text{in,clr}}$ are computed in a preprocessing routine according to Corripio (2003); Iqbal (1983), considering all effects of surface topography including shading and assuming standard atmospheric transmission coefficients for clear sky conditions. During the mass balance model run, $S_{\text{in}}$ of every individual time step is computed from preprocessed $S_{\text{in,clr}}$ and attenuation of clouds ($\tau_\text{cl}$). The latter is derived from RCM fractional cloud cover at the actual time step according to an empirical relationship derived from observations on the Greenland ice sheet (GrIS) (Konzelmann et al 1994):

$$\tau_\text{cl} = 1.0 - bn^2 - \exp(-cz)$$

with $b = 0.78$ and $c = 0.00085$ being constants and $z$ the surface elevation above sea-level.
The source of accumulation (P) and a threshold temperature (T_{snow}) of 1.5°C in combination with a transition range of 0.5°C (i.e., linear increase of the rain fraction from 0% at 1°C to 100% at 2°C) is used to distinguish P_{solid} from rain. Redistribution of snow by wind or avalanches is not considered in the model. Refreezing is calculated according to Pfeffer \textit{et al} (1991), expressing the amount of melt water (M) that can be retained as a ratio of annual accumulation (C):

$$M \geq \frac{c}{C} [\frac{\rho_{PC} + \rho_c}{\rho_c}] \left[1 + \frac{\rho_{PC} + \rho_c}{\rho_c}\right]^{-1}.$$

Thereby c refers to the heat capacity (1950 J K^{-1} kg^{-1}) of ice and ρ_{PC} is the pore close-off density (830 kg m^{-3}). The initial firn temperature (T_i) at the onset of the melt season (in positive degrees Celsius below freezing, (cf Pfeffer \textit{et al} 1991)) is here calculated as the mean annual air temperature of the preceding three years. The initial firn density (ρ_i) is set equal to snow density and is derived from an empirical relationship based on observations from the GrIS (Reeh \textit{et al} 2005):

$$ρ_i = 0.625 + 18.7T_i + 0.2937 T_i^2.$$

Glaciers are regarded as debris-free which is a reasonable assumption for the bulk of Greenland’s GICs (cf Rastner \textit{et al} 2012). Depending on the surface characteristics (snow, firm or ice) three different constant values for the onset of the ablation model: α_s = 0.75, α_f = 0.55 or α_i = 0.40 (α_i calibrated to 0.35 for the regions ‘South’ and ‘Mittivakkat’, cf table 1). Accumulated snow is assigned α_i when its age exceeds 1 year and after 2 years its ablation is lowered to α_f. This parameterization aims at approximating the ablation lowering related to the snow aging and is not meant to simulate the actual conversion from snow to ice that take much longer on most of Greenland’s glacier surfaces (e.g. Reeh \textit{et al} 2005).

4. Coupling of mass balance model to RCM data

The mass balance model is forced from daily gridded RCM output. Rugged high mountain topography is not represented by the coarse spatial resolution of RCMs which is a main reason for biases or shifted pattern in meteorological variables (e.g. Franco \textit{et al} 2012). The biases are addressed through (1) downscaling, (2) bias correction and (3) mass balance model calibration (cf Salzmann \textit{et al} 2012, Machguth \textit{et al} 2012). The three stages are described in the following sub-sections and their position in the modeling process is shown in figure 2.

4.1. Downscaling of the RCM data and bias correction

The simple downscaling procedure includes two steps: (1) The RCM grids (n, T and P) are interpolated to the resolution of the DEM (here 250 m) by means of inverse distance weighting (IDW) followed by (2) the application of sub-grid parameterizations (cf Machguth \textit{et al} 2009). Prior to the interpolation of T the strong dependence on altitude is removed by reducing T to a standard altitude z_{ref} = 0 m a.s.l. by means of altitudinal gradients γ_T where i denotes the number of the month (1–12). The atmospheric lapse rates γ_T were derived according to the empirically derived findings from Fausto \textit{et al} (2009). We use the values derived by Fausto \textit{et al} (2009) including the land stations. Fausto \textit{et al} (2009) published only annual-mean and July γ_T and thus we reconstruct the lapse rates for the remaining months using linear interpolation.

In step (2) simple parameter specific (T, P, S_m) sub-grid parameterizations are applied to account for the major components of the small-scale influences of the rugged topography. Sub-grid scale variability of T is addressed by adjusting T from z_{ref} to the elevation of the DEM using the lapse rates γ_T. The same method is applied for P but the difficulty is to define a vertical gradient in precipitation (γ_P) as reliable observations of P in Greenland’s mountain areas do not exist. As a proxy for γ_P we use observed winter mass balance distributions from the only two longer-term
mass balance series: The 10 year series from Amitsuloq ice cap in the south-west reveals a weak negative trend with altitude (cf Ahlstrøm et al 2007) while on Mittivakkat an increase in accumulation with elevation is observed (Knudsen and Hasholt 2008). The two series cannot be regarded representative for the whole of Greenland, but due to a lack of other data an intermediate value of 0.3 m precipitation increase per 100 m elevation increase was applied for all regions. It should be noted that $\gamma_T$ and $\gamma_P$ address only variability on the sub-grid scale and their influence decreases when the topography at the resolution of the mass balance model approaches the topography at the RCM resolution. Hence their influence is at maximum in rugged alpine topography and limited on larger ice caps.

The preprocessed grids of clear sky global radiation (section 3) account for the topographic sub-grid scale variability (i.e. exposition, slope and shading) while the temporally and spatially variable influence of cloudiness is derived directly from the interpolated fields of $n$ (cf Machguth et al 2009). Cloudiness is directly interpolated to the resolution of the mass balance model by means of IDW because daily mean $n$ is distributed rather smoothly in space.

Downscaled RCM fields require additional bias correction in $T$, $S_M$ and $P$ for accurate mass balance calculations (Machguth et al 2012). On Greenland, however, bias correction is hampered by the limited number of meteorological parameters being measured (only few weather stations do record $S_M$ or $n$) but also by the fact that most weather stations are located directly at the coast and are not representative for higher elevations (e.g. due to frequent inversions, Hansen et al 2008)). In addition, comparing a coastal station with coarse resolution (11–25 km) RCM grid cells that might be of surface type ‘ocean’, bears the risk to establish unreasonable bias corrections. Hence a direct bias correction is performed (i) only for $T$ and (ii) only for selected regions where suitable weather stations (located away from the open ocean) are available. Where these conditions are fulfilled (table 1) the mean deviation between observed and modeled $T$ (1980–2010) for every day of the year was calculated. At every time step during the model run, $T$ was then corrected by adding the offset to the downscaled temperature field. Note that this kind of bias correction does not change the future trends in $T$ as prescribed by the RCM data.

4.2. Mass balance model calibration

Figure 2 visualizes the process of mass balance model calibration in the context of the full modeling chain. Calibration of the mass balance model is challenging because the driving RCMs are forced from GCM data, and thus modeled mass balance must be compared to observations representing climatological means. Hence model output (driven from downscaled and de-biased RCM data) for the time period 1980–2010 was compared to our ‘best guess’ in mean summer mass balance and total cumulative mass balance $\sum B_0$. We emphasize the term ‘best guess’ because in no case a complete 30 years time series is available: Mittivakkat (Knudsen and Hasholt 2008) is the longest series (15 years in the time frame 1995–2010) and Amitsuloq (1981–1990) (Ahlstrøm et al 2007) the second longest.

Modeled summer-melt is adjusted to the measurements in a first calibration step (‘melt calibration’ in figure 2). However available summer mass balance observations do in most regions only cover one to three years which is too short for an automated and systematic adjustment of the model parameters. Thus calibration is restricted to a manual qualitative comparison and adjustment of model parameters $C_0$, $C_1$ and $\alpha_B$ (cf table 1).

After RCM downscaling, correction of $T$-bias and the aforementioned summer mass balance calibration, there is still considerable inaccuracy in modeled glacier mass balances. A main reason is the generally large differences of RCM precipitation pattern and the largely unknown real-world precipitation distribution (Machguth et al 2012). In addition, the RCMs are driven by ECHAM5 and HadGEM2 that fail to simulate the current climate over Greenland (Fettweis et al 2013). Such biases in the forcing GCMs impact the RCM results and affects our mass balance reconstruction. This explains why corrections are needed here. The comparison with the observations would be better if the RCMs were to be driven by reanalyses but no future projection is available in this case.

Hence it is justified to use $P$ as an ultimate mean of model calibration (‘calibration $M_P$’ in figure 2). It was assumed that the modeled $\sum B_0$ 1980–2010 of each glacier must be identical to our best guess of observed $\sum B_0$ and $P$ was tuned to achieve agreement. Our best guess of $\sum B_0$ was derived as follows: For the decade 1980–1990 we assumed that Greenland GICs were in a balanced state as it was roughly the case for the GrIS (Ohmura et al 1999) and Mittivakkat glacier (Knudsen and Hasholt 1999). For the decade 2000–2010 we assumed that the mass balance is well represented by the regional specific mean ICESat derived mass balance for the time period 2003–2008 (Bolch et al 2013). Comparably little is known for 1990–2000 and we assumed that the mass balance has linearly decreased from the 80s to the value of the decade 2000–2010. The sum of the three decadal values is the region specific $\sum B_0$ which is assumed to be valid uniformly for all glaciers of a region. The latter assumption is a strong simplification but there is no data available to assess the variability within a region. The precipitation adjustment is performed individually for each modeled glacier to achieve $\sum B_0$ for 1980–2010. Only the cumulative mass balance is forced to be $\sum B_0$, variability of annual mass balance is not affected. Precipitation is varied iteratively by adjusting a temporally constant precipitation multiplication factor ($M_P$) until for each glacier a cumulative mass balance of $\sum B_0$ with a tolerance of $\pm 0.5$ m w.e. results.

The calibration involves considerable uncertainties because in part, the GCMs that force the RCMs are not able to simulate the current mean climate and variability over Greenland. To assess their impact, model calibrations to three different values of $\sum B_0$ (minimum, maximum and intermediate, table 1) are performed. The three values are calculated for each selected region based on estimated uncertainty in the mass balance assumption for 1980–2010.
Glacier retreat is simulated based on a $\Delta h$ glacier retreat approach according to Huss et al. (2010b). Glacier surface elevation change originating from ice dynamics and surface mass balance is parametrized by distributing volume gain or loss (resulting from the surface mass balance) over the entire glacier surface according to altitude dependent functions. These $\Delta h$ functions (see next paragraph) are derived from previously observed changes in glacier thickness that incorporate both the influence of ice dynamics and surface mass balance. Glacierized grid cells become ice-free when their elevation falls below the altitude of the glacier bed. Glacier advance is not possible in the given approach. The $\Delta h$ approach is mass-conserving with respect to the surface mass balance, i.e. mass loss or gain in the year $m$ is converted into glacier thickness change and the DEM is updated at the onset of the year $m + 1$. Glacier surface mass balance calculation in the year $m + 1$ is performed on the updated topography and thus considers the feedback of surface elevation change on mass balance.

The applied $\Delta h$ functions are derived from ICESat data for all GICs (Bolch et al. 2013). For the four major regions of Greenland (cf Bolch et al. 2013) an altitude-dependency of glacier thickness change and surface elevation can be derived (figure 3). The $\Delta h$ functions are calculated from normalized elevation extent and changes in $h$ are forced to be 0 at the highest point and 1 at the lowest. This does not fully agree with the observations in all regions, but is required to guarantee a proper functioning of the $\Delta h$ approach. Huss et al. (2010b) recommend using different $\Delta h$ functions for different glacier size classes. We use the functions displayed in figure 3 uniformly for all glaciers of a region because the number of ICESat points in insufficient to derive glacier-type specific $\Delta h$ functions.

6. Sensitivity analysis

A sensitivity analysis is performed in a similar manner as presented by Salzmann et al. (2012). Thereby (i) the influence of an expected albedo decrease was investigated by gradually lowering $\alpha$ from 0.4 to 0.2 during the time period 2020–2040, (ii) the respective $\Delta h$ function was modified towards a more pronounced mass loss at the tongue (active glacier retreat) and a more uniform mass loss over the entire area (down wasting). To limit computation time the sensitivity experiments were only performed for the regions Sukkertoppen and Zackenberg under HIRHAM5 forcing and $\sum B_a$ intermediate calibration.
Figure 3. Observed \( \frac{dh}{dt} \) values (m a\(^{-1}\)), normalized \( \frac{dh}{dt} \) values and the derived \( \Delta h \) functions averaged over 100 m elevation intervals for the four main regions of Greenland (cf Rastner et al 2012, Bolch et al 2013, for the extent of the four regions). The error bars for \( \frac{dh}{dt} \) are given as 1\( \sigma \) of all \( \frac{dh}{dt} \) values in an elevation class. No standard deviation is calculated where less than five values are present in an elevation class.

7. Results

7.1. Modeled mass balance and glacier shrinkage

Our results indicate a decrease in ice volume for all six investigated regions. Thereby volume changes are smaller (−3% to −50%) from the central east to the very north and largest (−73% to −92%) in the south and south-east (figure 4). Absolute values of initial ice volume (\( V_{2000} \)) and area (\( A_{2000} \)) as well as final ice volume (\( V_{2098} \)) and area (\( A_{2098} \)) are given in table 2.

In general a realistic picture of modeled glacier retreat results (figures 5 and 6): The examples from the Sukkertoppen area show very pronounced retreat in the relatively moist near-coastal areas (figure 5(a)) and moderate changes of the Amitsuloq ice cap (figure 5(b)) which lies in the rain-shadow of larger ice caps and receives less precipitation. These results are in agreement with the expected larger climate sensitivity of more maritime glaciers (e.g. Oerlemans and Fortuin 1992). Mittivakkat glacier (figure 5(c)) has retreated by approx. 5 km by 2098, corresponding to a retreat rate of 50 m a\(^{-1}\). The large valley glaciers in the Stauning Alper (figure 6) have retreated by 4–7 km, yielding an annual retreat rate of roughly 40–70 m.

The MAR forcing results in the strongest mass loss for all regions (figure 4) and differences to the HIRHAM5 forcing are significant although the two RCMs are forced by the same GCM fields (ECHAM5 under the A1B emission scenario). The main reason for the stronger mass loss under MAR forcing is a more pronounced decrease in summer mass balance (\( B_s \)) (figure 7) in the north-eastern and northern regions. Trends in \( B_s \) in the southerly regions (Sukkertoppen, South and Mittivakkat) are more similar for all three forcings but MAR is still the most negative. More negative \( B_s \) must be related to a more pronounced warming. In a comparison study focused on the GrIS, Rae et al (2012) show that when downscaling the same GCM scenario (ECHAM5-A1B) MAR will have stronger future (2000–2099) warming trend than HIRHAM5. Thereby strongest deviations are observed in the north-eastern marginal regions of the ice sheet, close to the regions where this study finds the largest differences. However, the findings of Rae et al (2012) refer to the ice sheet while the mass balance model in the present study is mainly forced with RCM output from grid cells that are ice-free in the RCMs: when using MAR and HIRHAM5 78% of the area of the modeled GICs is located on RCM cells with surface type ‘land’. RACMO2 cells can have fractional ice cover and the average ice cover fraction of the grid cells underlying the modeled GIC is 54%. HIRHAM5 and RACMO2 forcing result in similar glacier changes with mass loss being overall larger under RACMO2 forcing. All three forcing fields obtained with the RCM downscaling runs generate similar trends in winter mass balance (\( B_w \)) despite the different emission scenarios and GCM forcing fields.
$B_{w}$ shows clearly increasing trends from the central east to the very north (Stauning, Zackenberg and North) and no trends for the more southerly regions. When forcing the model with the output from RACMO and HIRHAM5 the increase in $B_{w}$ in the three northern regions almost counterbalances the decrease in $B_{s}$, under MAR the decrease in $B_{s}$ dominates for all regions.

Figure 7 shows for the Mittivakkat region a systematically lower $B_{w}$ under the MAR forcing. The reason is that MAR strongly underestimates precipitation in the area and that $T$ was lowered by 1.5 K to avoid strong precipitation corrections that would impact on the future precipitation trend. However, the temperature lowering results in a smaller glacier mass turnover and thus likely a lower sensitivity to climate change. The issue could only be solved when precipitation is bias corrected without impacting on future trends. The Stauning Alper also show a lower value for $B_{w}$ MAR. The reason is, however, different: $B_{s}$ (MAR) over the calibration period is less negative and hence model calibration results in lower precipitation.

The aforementioned issues in model calibration are one of the reasons to perform three different calibrations for each region and each RCM forcing. The volume changes under calibrations minimum and maximum are shown as deviations (±) in table 2. The sensitivity analysis reveals that gradually lowering $\alpha_{i}$ to 0.2 from 2020 to 2040 and holding the albedo constant afterwards increases the volume loss for Zackenberg by 13 km$^{3}$ or 24% compared to no albedo change. For the Sukkertoppen region the increase in volume loss reaches 78 km$^{3}$ or 29%. Modifying the $\Delta f$ functions has a comparably small impact of approximately ±2% for both regions.

### 7.2. Extrapolation to all Greenland GICs

We divided Greenland into five sectors to upscale modeled volume change from the modeled regions to the entire sectors
Table 2. Overview of modeled initial and final glacier volume and area in the six selected regions ($A_{2000}$, $A_{2098}$ in km$^2$; $V_{2000}$, $V_{2098}$ in km$^3$). The given values are based on the calibration to $\sum B_a$ intermediate while the deviations ($\pm$) are calculated from the calibrations maximum and minimum.

| Region     | Model    | $A_{2000}$ | $A_{2098}$ | $V_{2000}$ | $V_{2098}$ |
|------------|----------|------------|------------|------------|------------|
| North      | HIRHAM5  | 4446       | 4386 ± 25  | 777        | 745 ± 10   |
|            | RACMO2   | 4087 ± 48  | 685 ± 9    |            |            |
|            | MAR      | 3665 ± 57  | 590 ± 10   |            |            |
|            |          | 4087 ± 48  | 685 ± 9    |            |            |
|            |          | 3665 ± 57  | 590 ± 10   |            |            |
| Zackenberg | HIRHAM5  | 1541       | 1210 ± 77  | 206        | 151 ± 10   |
|            | RACMO2   | 1333 ± 56  | 161 ± 9    |            |            |
|            | MAR      | 875 ± 61   | 101 ± 8    |            |            |
| Stauning   | HIRHAM5  | 3884       | 3402 ± 83  | 327        | 255 ± 14   |
|            | RACMO2   | 3490 ± 85  | 266 ± 15   |            |            |
|            | MAR      | 2744 ± 149 | 182 ± 13   |            |            |
|            |          | 3402 ± 83  | 255 ± 14   |            |            |
|            |          | 2744 ± 149 | 182 ± 13   |            |            |
| Mittivakkat| HIRHAM5  | 2222       | 805 ± 142  | 311        | 72 ± 14    |
|            | RACMO2   | 874 ± 114  | 84 ± 13    |            |            |
|            | MAR      | 538 ± 108  | 52 ± 9     |            |            |
|            |          | 874 ± 114  | 84 ± 13    |            |            |
|            |          | 538 ± 108  | 52 ± 9     |            |            |
| South      | HIRHAM5  | 170        | 31 ± 6     | 7.8        | 0.6 ± 0.2  |
|            | RACMO2   | 28 ± 6     | 0.7 ± 0.2  |            |            |
|            | MAR      | 27 ± 6     | 0.7 ± 0.1  |            |            |
|            |          | 31 ± 6     | 0.6 ± 0.2  |            |            |
|            |          | 28 ± 6     | 0.7 ± 0.2  |            |            |
| Sukkertoppen| HIRHAM5 | 5398       | 4325 ± 90  | 1180       | 910 ± 23   |
|            | RACMO2   | 4006 ± 90  | 804 ± 21   |            |            |
|            | MAR      | 3617 ± 100 | 751 ± 21   |            |            |
|            |          | 4006 ± 90  | 804 ± 21   |            |            |
|            |          | 3617 ± 100 | 751 ± 21   |            |            |
Table 3. Volume loss area scaling relationship calculated from the modeled volume loss of the six selected regions. For simplicity only the values based on the calibration to $\sum B_{intermediate}$ are given.

| Selected region | HIRHAM5 | RACMO2 | MAR | HIRHAM5 | RACMO2 | MAR |
|-----------------|---------|--------|-----|---------|--------|-----|
| North           | 0.0045871 | 0.0371580 | 0.029589 | 1.0795 | 0.88507 | 1.0645 |
| Zackenberg      | 0.0336320 | 0.0276310 | 0.048589 | 1.0113 | 1.00800 | 1.1294 |
| Stauning        | 0.0061142 | 0.0055372 | 0.018061 | 1.2330 | 1.23410 | 1.1843 |
| Mittivakkat     | 0.0333000 | 0.0339930 | 0.034105 | 1.2967 | 1.28680 | 1.3084 |
| Sukkertoppen    | 0.0279680 | 0.0251590 | 0.028043 | 1.0928 | 1.19120 | 1.1941 |
| South           | 0.0200270 | 0.0208730 | 0.020783 | 1.3689 | 1.34790 | 1.3582 |

Table 4. The five sectors of Greenland and extrapolated volume loss 2000–2098. The given values are based on the calibration to $\sum B_{intermediate}$ while the deviations ($\pm$) are calculated from the calibrations maximum and minimum.

| Sector       | Selected region | $A_{2000}$ (km$^2$) | HIRHAM5 | RACMO2 | MAR |
|--------------|-----------------|---------------------|---------|--------|-----|
| North        | North           | 39 327              | −277 ± 98 | −829 ± 77 | −1548 ± 93 |
| East-North   | Zackenberg      | 5 789               | −200 ± 46 | −163 ± 36 | −274 ± 9  |
| East-Central| Stauning        | 21 618              | −302 ± 72 | −275 ± 67 | −742 ± 64 |
| East-South   | Mittivakkat     | 10 220              | −970 ± 44 | −952 ± 42 | −1042 ± 14 |
| West         | Sukkertoppen    | 12 765              | −491 ± 41 | −652 ± 33 | −735 ± 37 |
| Total        |                 | 89 719              | −2240 ± 143 | −2871 ± 121 | −4341 ± 120 |

Figure 6. Stauning Alper: modeled glacier retreat (using MAR) for a number of valley glaciers. The elevation contours represent surface topography in 2098.
8. Discussion and conclusions

In the present study sea-level rise contribution from all local glaciers and ice caps on Greenland is modeled and a range of projections is defined by (i) using three different RCM forcings, (ii) performing three different mass balance model calibrations under each RCM forcing, and (iii) performing a sensitivity analysis to explore the effect of a potential decrease in ice albedo and variations in the applied $\Delta h$ functions.

The choice of the forcing field strongly impacts on modeled sea-level rise contribution. Interestingly the results from the two RCM that are forced with the same GCM output fields divert the most. A likely reason for these differences is the positive albedo feedback that enhances the warming and is considered in MAR but not in HIRHAM5. RACMO2 also takes into account the surface albedo feedback and this could partly explain the stronger mass loss under the RCP4.5 scenario (HadGEM2 RACMO2 forcing) compared to A1B (ECHAM5 HIRHAM5 forcing). The mass balance model calibration has for most regions a smaller impact. A statistical uncertainty range cannot be calculated based on three RCM forcings and hence we provide only modeled mass loss for the three forcings with their respective deviations resulting from the calibrations. The albedo sensitivity experiment indicates that lowering the ice albedo to 0.2 by 2040 increases mass loss by roughly 20–30%. A clear lowering trend in surface albedo is currently observed in the ablation area of the GrIS (Box et al. 2012). However, future projections of albedo change are of a speculative nature and therefore we do not consider the impact of albedo lowering in the final numbers. Nevertheless, the sensitivity experiment indicates that the given final numbers of mass loss are lower boundary estimates.

The modeled annual sea-level rise contribution (0.059 ± 0.004 mm HIRHAM5, 0.076 ± 0.003 mm RACMO2 and 0.114 ± 0.003 mm yr$^{-1}$ MAR) are in the range of the current contribution of 0.08 ± 0.03 mm yr$^{-1}$ from 2003 to 2008 (Bolch et al. 2013). Given the continued warming trend a higher contribution than observed today could be expected. The most likely reasons for the modeled rate of mass loss being similar to current observations are (i) the use of mid-range scenarios, (ii) the predicted increase in precipitation over large areas of GICs in the north-east and north, (iii) the fact that the fastest melting parts of the glaciers disappear first which results in an asymptotic behavior of the mass loss rate and (iv) the exclusion of all calving processes from the modeling chain. According to Bolch et al. (2013) the rate of mass loss of calving glaciers is 10% higher than for non-calving glaciers. However, continued glacier retreat might eventually reduce the impact of calving for many regions and a substantial part of the ice lost due to calving is below sea-level and will thus not contribute to sea-level rise.

The sea-level rate computed by Radić and Hock (2011) (0.036 ± 0.02 mm yr$^{-1}$) is only about half (when forced by HIRHAM5 and RACMO output fields) or a third (when forced by MAR output) of our results. However, Radić and Hock (2011) underestimated the surface area of Greenland’s GICs. When comparing mass loss per area-unit (here and in the following mass loss per area-unit is calculated based on the initial surface area used by the various studies) and considering also the uncertainties given by Radić and Hock (2011), the two studies are in reasonable agreement. Then again Marzeion et al. (2012) have calculated a higher sea-level rise contribution from Greenland’s GICs under the
RCPP4.5 scenario (approx. 0.16 mm yr\(^{-1}\) until 2100). In the following we put these scenarios into the context of GrIS future scenarios: modeled future sea-level rise contribution (2000–2100) from the GrIS using a set of GCMs under the A1B scenario indicate 1.5–6.5 cm with a median of 4 cm (according to Fettweis et al 2008, considering only changes in surface mass balance) or 0–17 cm with a mean of 4.5 cm (according to Graversen et al 2010, considering also dynamic mass loss). Fettweis et al (2013) find a sea-level rise contribution from the GrIS surface mass balance of 4 ± 2 cm by 2100 (RCPP4.5 scenario). In this context the modeled contributions from Greenland GICs are remarkably large: according to Rastner et al (2012) the GICs area corresponds to only 5.4% of the GrIS and if the GrIS’ future mass loss per area-unit were the same as modeled in the present study for the GICs, the GrIS’ future (2100) sea-level rise contribution would be roughly 11 cm (HIRHAM5, A1B), 21 cm (MAR, A1B) or 14 cm (RACMO2, RCP4.5). Comparing these numbers with the aforementioned projections for the GrIS (Fettweis et al 2008, Graversen et al 2010, Fettweis et al 2013) shows that our study and Radič and Hock (2011) indicate a higher (by approx. a factor of three to four) sensitivity to climate change for the GICs while the results of Marzeion et al (2012) suggest mass loss per area-unit being roughly seven to eight times larger on the GICs than on the GrIS. According to Bolch et al (2013) the observed (2003–2008) mass loss per area-unit is about two to three times larger on the GICs than on the GrIS.

Glaciers with low accumulation at the ELA are less sensitive to a given change in climate than more maritime glaciers (e.g. Oerlemans and Reichert 2000). Our results are in good agreement with this general law of glacier sensitivity, but the characteristics of glacier climate sensitivity are not the sole reason for the modeled smaller changes in the dry and cold north-east and the north. On the one hand mass loss in these areas would be larger if precipitation remained constant. On the other hand precipitation increase in a warming climate also leads to a change in glacier characteristics. For instance the glaciers in the Stauing Alper at the end of the 21st century show average values of winter and summer balances (RACMO2 and HIRHAM5 forcing) comparable to the glaciers in the Mittivakkat region at the beginning of the century. This change in glacier characteristics towards increased mass turnover and consequently larger climate sensitivity is a major challenge for future projections. Stating that glaciers in cold and dry climates will play a minor role among sea-level rise contributors due to their low climate sensitivity, neglects the impact of a change in climate on glacier’s climate sensitivity.

It is a major strength of the chosen modeling approach to deliver comprehensible maps of glacier change and thereby including important feedback processes. Nevertheless, our sea-level rise projections for Greenland’s GICs are based on several models and assumptions introducing various sources of uncertainty. By means of different calibrations and sensitivity experiments a number of them have been addressed. Further research is required to assess uncertainties related to model-design and thus we conclude by highlighting two major challenges for model improvement:

(1) 2 m air temperatures used for input to the mass balance model are influenced by the surface properties of the RCM grid cells (i.e. bare land, sea or ice). By adjusting the parameters C\(_0\) and C\(_1\) the mass balance model can be calibrated for use with 2 m air temperature from within or outside a glacier boundary layer (cf Citterio et al 2011). Inconsistencies, however, result where the calibration and the RCM surface type do not match. This is a minor issue with the coarse MAR or HIRHAM5 grids where roughly 80% of the modeled GICs are located on RCM cells of the type ‘land’, but it becomes important when using RACMO2 with its more detailed glacier mask. One solution might be the use of a spatially varying calibration. Another approach would be reducing the influence of the RCM’s surface properties on surface air temperatures by reconstructing the latter from temperatures in the free atmosphere using methods similar to Jarosch et al (2012).

(2) Important issues arise from the modeling of the ice caps: firstly the inaccuracies in the simple approach to assess bed topography need further investigation and comparison to alternative approaches (e.g. Huss and Farinotti 2012). Secondly, the application of the \(\Delta h\) approach to ice caps bears the risk of underestimating elevation changes in the accumulation areas. Originally developed for valley glaciers (Huss et al 2010b), the approach assumes that the highest point of a glacier does not change elevation. Observed changes on ice caps, however, are more complex: some ice caps are thickening in their accumulation areas and thinning in the ablation zone (Rinne et al 2011) while others are only thinning (e.g. Bolch et al 2013). None of the existing coupled glacier mass balance and ice dynamics models are capable to fully address the complexity of volume changes of ice caps. Nevertheless benchmarking the \(\Delta h\) approach on selected ice caps against dynamic models would help to quantify possible underestimation in elevation change.

Acknowledgments

Henrik Højmark Thomsen and Anker Weidick provided most valuable help in tracking down almost forgotten data on ice thickness observations and mass balance. The comments and suggestions of three anonymous reviewers helped to improve the manuscript and are highly acknowledged. HM is supported by the PROMICE program for the monitoring of the Greenland ice sheet and GlacioBasis monitoring funded by the Danish Energy Agency. This work is supported by funding from the ice2sea program from the European Union 7th Framework Program, grant number 226375. Ice2sea contribution number ice2sea135.

References

Ahlægisdóttir G, Stendel M, Christensen J H, Cappelen J, Vejen F, Kjær H A, Mottram R and Lucas-Picher P 2009 Assessment of the temperature, precipitation and snow in the RCM HIRHAM4 at 25 km resolution Danish Climate Centre Report 09-08 (Copenhagen: Danish Meteorological Institute)

Ahlstrøm A P, Böggild C E, Olesen O B, Petersen D and Mohr J J 2007 Mass balance of the Amitsulâq ice cap, West Greenland Glacier Mass Balance Changes and Meltwater
Discharge (IAHS Publication vol 318) (Wallington: IAHS Press) pp 107–15
Angelen J H V, Lenaerts J T M, Lhermitte S, Fettweis X, Munneke P, van den Broeke M R, van Meijgaard E and Smeets C J P F 2012 Sensitivity of Greenland ice sheet surface mass balance to surface albedo parameterization: a study with a regional climate model Cryosphere 6 1175–86
Bahr D B, Meier M F and Peckham S D 1997 The physical basis of glacier volume–area scaling J. Geophys. Res. 102 20355–62
Boas L and Wang P R 2011 Weather and climate data from Greenland 1958–2010, observation data with description Technical Report (Copenhagen: Danish Meteorological Institute)
Bolch T, Sandberg Sørensen L, Simonsen S B, Mølgaard N, Boas L and Wang P R 2011 Weather and climate data from
Bolch T, Sandberg Sørensen L, Simonsen S B, Mølgaard N, Boas L and Wang P R 2011 Weather and climate data from Greenland 1958–2010, observation data with description Technical Report 11–15 (Copenhagen: Danish Meteorological Institute)
Boas L and Wang P R 2011 Weather and climate data from Greenland 1958–2010, observation data with description Technical Report 11–15 (Copenhagen: Danish Meteorological Institute)
Citterio M, Fausto R S and Machguth H 2011 Glaciological thermodynamics and atmospheric drivers Cryosphere 5 630–7
Citterio M, Fausto R S and Machguth H 2011 Glaciological thermodynamics and atmospheric drivers Cryosphere 5 630–7
Citterio M and Mottram R 2008 Glaciological investigations at Malmbyg, Stauning Alper, East Greenland: field report and results of GPR surveys Technical Report (Copenhagen: Geological Survey of Denmark and Greenland)
Clement P 1982 Glaciologiske undersøgelser i Johan Dahl Land 1981 Tech. Rep. (Grønlands Geologiske Undersøgelse)
Corripio J 2003 Vectorial algebra algorithms for calculating terrain parameters from DEMs and solar radiation modelling in mountainous terrain Int. J. Geogr. Inform. Sci. 17 1–23
Fausto R S, Ahlstrøm A P, van As D, Bøggild C E and Corripio J 2003 Vectorial algebra algorithms for calculating terrain parameters from DEMs and solar radiation modelling in mountainous terrain Int. J. Geogr. Inform. Sci. 17 1–23
Fettweis X 2007 Reconstruction of the 1979–2006 Greenland ice sheet surface mass balance using the regional climate model MAR Cryosphere 1 21–40
Fettweis X, Franco B, Tedesco M, van Angelen J H, Lenaerts J T M, van den Broeke M R and Gallée H 2013 Estimating Greenland ice sheet surface mass balance contribution to future sea level rise using the regional atmospheric climate model MAR Cryosphere 7 469–89
Fettweis X, Hanna E, Gallée H, Huybrechts P and Erpich M 2008 Estimation of the Greenland ice sheet surface mass balance for the 20th and 21st centuries Cryosphere 2 117–29
Franco B, Fettweis X, Lang C and Erpich M 2012 Impact of spatial resolution on the modelling of the Greenland ice sheet surface mass balance between 1990 and 2010, using the regional climate model MAR Cryosphere 6 959–711
Gardner A S, Moholdt G, Wouters B, Wolken G J, Burgess D O, Sharp M J, Cogley J G, Braun C and Labine C 2011 Sharply increased mass loss from glaciers and ice caps in the Canadian Arctic Archipelago Nature 473 357–60
Gogineni S et al 2001 Coherent radial ice thickness measurements over the Greenland ice sheet J. Geophys. Res. 106 33761–72
Graversen R G, Drijkoningen H, Hwang C, van den Broeke M R and Bintanja R 2010 Greenland’s contribution to global sea-level rise by the end of the 21st century Clim. Dyn. 37 1427–42
Hammer C U (ed) 2001 The Hans Tausen Ice Cap: Glaciology and Glacial Geology (Meddelelser om Grønland: Geoscience vol 39) (Copenhagen: Danish Polar Center)
Hansen B U et al 2008 Present-day climate at Zackenberg Adv. Ecol. Res. 40 111–49
Howat I, Negrete A, Scambos T and Hanar T 2013 A high-resolution elevation model for the greenland ice sheet from combined stereoscopic and photoclinometric data (http://bprc.osu.edu/GIDO/gimpdem.php)
Huss M and Farinotti D 2012 Distributed ice thickness and volume of all glaciers around the globe J. Geophys. Res. 117 F04010
Huss M, Farinotti D, Bauder A and Funk M 2008 Modelling runoff from highly glacierized alpine drainage basins in a changing climate Hydrolog. Process. 22 3888
Huss M, Jouvet G, Farinotti D and Bauder A 2010b Future high-mountain hydrology: a new parameterization of glacier retreat Hydrolog. Earth Syst. Sci. 14 815–29
Huss M, Usellmann S, Farinotti D and Bauder A 2010a Glacier mass balance in the south-eastern Swiss Alps since 1900 and perspectives for the future Erdkunde 64 119–40
Iqbal M 1983 An Introduction to Solar Radiation (Toronto: Academic)
Johansson S J, Clausen H B, Dansgaard W, Gundestrup N S, Hansson M, Jonsson P, Steffensen J P and Sveinbjörnsson A A 1992 A ‘Deep’ Ice Core From East Greenland (Meddelelser om Grønland: Geoscience vol 29) (Copenhagen: Danish Polar Center) pp 1–22
Knudsen N T and Hasholt B 1999 Radio-echo sounding in the Mittivakkat gletscher, southeast Greenland Arct. Antarct. Alp. Res. 31 321–8
Knudsen N T and Hasholt B 2008 Mass balance observations at Mittivakkat glacier, Ammassalik island, southeast Greenland 1995–2006 J. Geog. Tidsskr.—Dan. J. Geogr. 108 111–20
Konzelmann T, de Wal R V, Greuell W, Bintanja R, Hennenek E and Abe-Ouchi A 1994 Parameterization of global and longwave incoming radiation for the Greenland ice sheet Glob. Planet. Change 9 143–64
Linsbauer A, Paul F and Haeberli W 2012 Modeling glacier thickness distribution and bed topography over entire mountain ranges with GlabTop: application of a fast and robust approach J. Geophys. Res. 117 F03007
Lowry H Machguth Haebeler W 2011 Mass balance parameters derived from a synthetic network of mass balance glaciers J. Glaciol. 58 965–79
Machguth H, Haebeler W and Paul F 2012 Mass balance parameters of all glaciers around the globe from combined stereoscopic and photoclinometric data J. Geophys. Res. 117 F04010
Oerlemans J 2001 Glaciers and Climate Change (Lisse: A Balkema Publishers)
Oerlemans J and Fortuin J P F 1992 Sensitivity of glaciers and small ice caps to greenhouse warming Science 258 115–7
Oerlemans J and Reichert B 2000 Relating glacier mass balance to meteorological data by using a seasonal sensitivity characteristic J. Glaciol. 46 1–6
Ohmura A, Calanca P, Wild M and Anklin M 1999 Precipitation, accumulation and mass balance of the Greenland ice sheet Z. Glaziol. Glazialgeol. 35 1–20
Paterson W S B 1994 The Physics of Glaciers (Amsterdam: Elsevier)
Pfeffer W, Meier M and Illangasekare T H 1991 Retention of Greenland runoff by refreezing: implications for projected future sea level change J. Geophys. Res. 96 22117–24
Radić V and Hock R 2011 Regionally differentiated contribution of mountain glaciers and ice caps of future sea-level rise Nature Geosci. 4 91–4
Rae J G L et al 2012 Greenland ice sheet surface mass balance: evaluating simulations and making projections with regional climate models Cryosphere 6 1275–94
Raper S C B and Braithwaite R J 2006 Low sea level rise projections from mountain glaciers and icecaps under global warming Nature 439 311–3
Rastner P, Bolch T, Mölg N, Machguth H, Le Bris R and Paul F 2012 The first complete inventory of the local glaciers and ice caps on Greenland Cryosphere 6 1483–95
Reeh N, Fisher D A, Koerner R M and Clausen H B 2005 An empirical firm-densification model comprising ice lenses Ann. Glaciol. 42 101–6
Rinne E J, Shepherd A, Palmer S, van den Broeke M R, Muir A, Ettema J and Wingham D 2011 On the recent elevation changes at the Flade Isblink ice cap, northern Greenland J. Geophys. Res. 116 F03024
Salzmann N, Machguth H and Linsbauer A 2012 The Swiss Alpine glacier’s response to the 2 °C air temperature target Environ. Res. Lett. 7 044001
Weidick A and Morris E 1998 Local glaciers surrounding the continental ice sheets Into the Second Century of Worldwide Glacier Monitoring—Prospects and Strategies ed W Haeberli, M Hoelzle and S Suter (Paris: UNESCO) pp 197–205
Zwinger T and Moore J C 2009 Diagnostic and prognostic simulations with a full Stokes model accounting for superimposed ice of Midtre Lovénbreen, Svalbard Cryosphere 3 217–29