RESEARCH ARTICLE

10.1002/2014JD021557

Key Points:

- The annual water vapor flux is toward the surface in the dry snow zone
- Mass turnover from water vapor exchanges will be enhanced in a warming climate
- A shift to sublimation in the dry snow zone will enhance mass loss from the GrIS

Correspondence to:

N. J. Cullen,
nicolas.cullen@otago.ac.nz

Citation:

Cullen, N. J., T. Mölg, J. Conway, and K. Steffen (2014), Assessing the role of sublimation in the dry snow zone of the Greenland ice sheet in a warming world, *J. Geophys. Res. Atmos.*, 119, 6563–6577, doi:10.1002/2014JD021557.

Received 26 JAN 2014

Accepted 23 APR 2014

Accepted article online 29 APR 2014

Published online 12 JUN 2014

Assessing the role of sublimation in the dry snow zone of the Greenland ice sheet in a warming world

Nicolas J. Cullen¹, Thomas Mölg², Jonathan Conway¹, and Konrad Steffen³
¹Department of Geography, University of Otago, Dunedin, New Zealand, ²Chair of Climatology, Technische Universität Berlin, Berlin, Germany, ³Swiss Federal Research Institute WSL, Birmensdorf, Switzerland

Abstract Meteorological and glaciological data obtained over an intensive 2 year measurement period (2000–2002) are used to run a physically based climatic mass balance model to characterize a seasonal variability in mass and energy exchanges at Summit, Greenland. The model resolves the full surface energy balance and the subsurface temperature profile by inclusion of energy release from penetrating shortwave radiation. A Monte Carlo approach using 1000 different parameter combinations is adopted to assess model uncertainty, with output compared to measured surface and subsurface temperatures, changes in surface height, and eddy correlation data. The heat exchanges associated with the change in phase of water are very small in all seasons, with the average turbulent latent heat flux equal to $0.4 (\pm 0.2) \text{ W m}^{-2}$. This suggests that the mean annual water vapor gradient is toward the surface, resulting in a mass gain of $4.1 \text{ mm WE yr}^{-1}$. The mass gain represents only a small fraction of the total accumulation ($< 2\%$), in part because of the change in sign of the water vapor flux from winter (deposition) to summer (sublimation), but if assumed to be typical of the entire dry snow zone (40% of the total ice sheet area) is equivalent to approximately 5.5 Gt yr^{-1} . A simple experiment based on 2012 atmospheric conditions suggests that mass turnover from water vapor exchanges will likely be enhanced in a warming climate, with sublimation increasing more than deposition. Should the sign of the mean turbulent latent heat flux change due to warming, the present mass gain in the dry snow zone could easily become a mass loss of equal proportion, which would further enhance the negative mass balance of the Greenland ice sheet.

1. Introduction

Our understanding of the stability of the Greenland ice sheet (GrIS) has changed considerably over the last 2 decades, and it is now well established that the largest source of freshwater in the northern hemisphere [Bamber *et al.*, 2013] is contributing to sea level rise [van den Broeke *et al.*, 2009; Shepherd *et al.*, 2012]. Evidence has been carefully compiled to demonstrate that surface melting and run off and ice discharge from marine-terminating outlet glaciers have been primarily responsible for the increase in freshwater input into our oceans. Until recently, the retreat and acceleration of glaciers along the margins of the GrIS [e.g., Howat *et al.*, 2007; van de Wal *et al.*, 2008; Joughin *et al.*, 2010; Sundal *et al.*, 2011; Moon and Joughin, 2008; Moon *et al.*, 2012] have contributed to at least half of the observed mass loss. However, it now appears likely that a change in the surface mass balance (MB) has become the most important factor controlling mass loss through the increase in surface melting and run off [e.g., Enderlin *et al.*, 2014]. The change in melt pattern, which is characterized by an increase in the areal extent and duration of surface melt, complicates efforts to establish the mass budget of the GrIS, as the fate of meltwater in firn and snow at higher elevations is poorly constrained [Harper *et al.*, 2012; Forster *et al.*, 2014]. For example, the identification of substantial areas of meltwater in liquid form that have persisted through winter demonstrates that the passage of meltwater from ice sheet to ocean is complex [Harper *et al.*, 2012], which has important implications on both the mass and the energy budget of the GrIS.

The increase in surface melting and run off has been linked to a rise in air temperature, with changes observed in the Greenland region correlated to global warming [Hanna *et al.*, 2008]. Although it has been anticipated that a rise in air temperatures might lead to an upward migration of the percolation zone, leading to changes in the transport and storage of meltwater, a brief melt event that extended across almost the entire ice sheet (approximately 98%) during the summer of 2012 [Nghiem *et al.*, 2012;

Tedesco *et al.*, 2013] was much less expected. The 2012 melt event demonstrated that the dry snow zone, which has traditionally accounted for slightly more than 40% of the ice sheet surface area [Ohmura *et al.*, 1999], is no longer out of harm's way from the impacts of surface melt. Should such events become more frequent in the future [e.g., McGrath *et al.*, 2013], we could witness the demise of the last substantial dry snow facies in the northern hemisphere. To assess the likelihood of this happening, it is essential that we continue to establish linkages between large scale atmospheric circulation and surface melt [Fettweis *et al.*, 2013b; Hanna *et al.*, 2013a, 2013b; Tedesco *et al.*, 2013]. This requires us to also consider whether global warming has affected atmospheric circulation in such a way as to enhance widespread melt events.

To assess whether the GrIS can survive a decrease in the extent of the dry snow zone, it is imperative that we understand the processes that allow mass to accumulate. The first step toward this goal is to establish the variability in snow accumulation for the GrIS, which has been determined using ice/firn cores and coastal precipitation records [e.g., Ohmura and Reeh, 1991; Calanca *et al.*, 2000; Bales *et al.*, 2001a, 2001b, 2009], global climate model simulations [e.g., Thompson and Pollard, 1997; Ohmura *et al.*, 1999], atmospheric reanalysis data [Hanna *et al.*, 2001, 2006, 2011], and regional climate models (RCMs) [e.g., Dethloff *et al.*, 2002; Box *et al.*, 2004, 2006, 2013; Fettweis, 2007; Ettema *et al.*, 2010a, 2010b; Rae *et al.*, 2012; Fettweis *et al.*, 2013a]. The advantage of using RCMs is that mass loss associated with water vapor exchanges can be explicitly accounted for, which is critical when determining variations in accumulation and/or surface MB. The Greenland Climate Network (GC-Net) [Steffen and Box, 2001], the Programme for Monitoring of the Greenland Ice Sheet [van As *et al.*, 2013], as well as the automatic weather stations (AWSs) and glaciological measurements along the Kangerlussuaq transect [van de Wal *et al.*, 2012], have been essential in assessing the skill of RCMs. Despite the impressive advances being made in refining these models, which has allowed the uncertainties in estimating MB to be narrowed, there is still a need for detailed atmospheric and glaciological measurements to be undertaken on the GrIS. Direct observations remain critical in efforts to explicitly resolve the atmospheric controls on the surface energy balance (SEB) that ultimately lead to more robust estimates of MB.

Recent progress has been made in the ablation area of the GrIS [Smeets and van den Broeke, 2008a, 2008b; van As *et al.*, 2012; van den Broeke *et al.*, 2008a, 2008b, 2008c, 2011], but less is known about the atmospheric processes governing the seasonal variability of the SEB in the dry snow zone. Instead, much of the focus has been on characterizing surface-atmosphere exchanges to determine the physical processes responsible for the deposition of chemical species in snow and later ice [e.g., Bottenheim *et al.*, 2002]. A number of detailed experiments carried out at Summit, near the top of the GrIS, has improved our understanding of the characteristics of the atmospheric boundary layer, in particular the weak instability of the atmosphere in summer [Cullen and Steffen, 2001], the intermittent nature of turbulence in stable conditions [Cullen *et al.*, 2007b; Drue and Heinemann, 2007], and the processes controlling the divergence of longwave radiation [Hoch *et al.*, 2007]. The role the shortwave radiation penetration has on the energy budget has also been explored [Kuipers Munneke *et al.*, 2009]. Some attempts have been made to characterize the SEB beyond a summer season [e.g., Albert and Hawley, 2000; Cullen, 2003; Hoch, 2005], but given the recent observed changes in melt extent over the GrIS and warming in its interior [McGrath *et al.*, 2013], it seems timely to carefully reassess the SEB in the dry snow zone and to consider what changes to MB might occur if warming continues. One of the key motivations for doing so is to assess whether mass is being gained or lost at the higher elevations through water vapor exchange, which directly influences the mass budget of the GrIS. The transfers of water vapor in the interior are small and a challenge to account for in RCMs, so there is merit in conducting a careful assessment of the SEB using an observational data set.

Thus, the aim of this study is to describe the seasonal variability of the SEB at Summit using 2 years of high-quality atmospheric and glaciological data. We use the surface layer similarity theory to calculate the turbulent sensible and latent heat fluxes from AWS observations, which are validated using eddy correlation measurements. Our observations and modeling approach allow us to assess the uncertainty surrounding sublimation in the dry snow zone and to demonstrate why determining the sign and magnitude of this flux is important in the assessment of MB under climate change. First, we describe the methods in section 2, including a description of the model used to calculate the energy and mass balance terms, followed by results (section 3), a brief discussion (section 4), and conclusions in section 5.

Table 1. Variables Measured and Sensor Specifications of Each of the Three Automatic Weather Stations Defined in the Text (Variables Labeled 1–3 Indicate the AWS They Were Mounted on)^a

Variable	Instrument	Range	Accuracy
1. Air temperature	Met One T200 PRT	−50 to +50°C	0.1°C
1. Relative humidity	Met One 083D	0 to 100%	2% (<90%), 3% (>90%)
1. Wind speed	Met One 014A	0 to 45 m s ^{−1}	0.1 m s ^{−1}
1. Wind direction	Met One 024A	0 to 360°	5°
1. Surface height	CSI SR50	0.5 to 10 m	0.01 m or 0.4%
1. Air pressure	Vaisala PTB101B	600–1060 hPa	4 hPa
2. Shortwave radiation	Kipp and Zonen CM21	0.35–2.8 μm	estimated accuracy of daily totals (EADT) ± 10% EADT ± 10%
2. Longwave radiation	Kipp and Zonen CG4	4.5–50 μm	
3. 3-D wind speeds	CSI CSAT3	0 to 65 m s ^{−1}	0.001 m s ^{−1} (u_x and u_y) 0.0005 m s ^{−1} (u_z)
3. Air temperature	CSI CSAT3	−50 to +60°C	0.026°C
3. Water vapor	CSI KH20	>0 g m ^{−3}	<0.3 g m ^{−3}

^aThe three AWSs were installed in close proximity to the Science Tower at Summit [Dibb and Fahnestock, 2004, Figure 1]. The accuracy of the radiation instruments is given as estimated accuracy of daily totals (EADT), as indicated by van den Broeke *et al.* [2004b] for similar sensors. The accuracy of the eddy covariance instruments is sensor resolution as defined by the manufacturers.

2. Methods

2.1. Site Description and Instrumentation

Meteorological and glaciological data used in this research were obtained between 18 June 2000 and 17 June 2002 at Summit, Greenland (72.58°N, 38.51°W, 3203 m above sea level), which is located on the northern dome of the GrIS, approximately 29 km west of the highest point. Since the successful extraction of the GISP2 ice core in the 1990s, Summit has become a year-round observatory, providing an ideal location for snow and atmospheric research in the dry snow zone. Our interest in the dry snow zone stems from the fact that more than half of the total net accumulation (or mass gain) of the GrIS occurs in this region [Ohmura *et al.*, 1999]. Thus, to understand the mass balance of the GrIS, it is essential to have a detailed understanding of the physical processes controlling mass gain and loss in this elevated area.

Details of the instruments will only be provided briefly here as they have been described carefully elsewhere [Cullen and Steffen, 2001; Cullen *et al.*, 2007b]. Three AWSs were installed approximately 50 m to the southwest of the long standing Science Tower at Summit to obtain: (1) profile measurements (air temperature, relative humidity, and wind speed) at three levels, as well as barometric pressure, surface height, and wind direction (AWS1); (2) broad spectrum (four component) radiation measurements and snow temperatures (AWS2); and (3) turbulence measurements (eddy correlation) at two heights (AWS3) (Table 1). It should be noted that these AWSs were installed independently from the GC-Net AWS at Summit [e.g., Steffen and Box, 2001], which is located approximately 2 km to the west of the Science Tower and the instruments used during the Greenland Summit experiment by Eidgenössische Technische Hochschule (ETH) Zurich [e.g., Hoch *et al.*, 2007]. The ETH measurements were obtained in close proximity to the AWS installed for this research as well as on the 50 m Swiss Tower (200 m to the east of Science Tower).

Profile measurements of air temperature, humidity, and horizontal wind speed were initially installed at 0.5, 1.0, and 2.0 m above the snow surface (AWS1). The Met One T200 (temperature) and 083D (relative humidity) instruments were housed in motor-aspirated radiation shields (Met One, 077). All instruments, including wind direction, surface height, and barometric pressure were sampled every 5 s and stored as 10 min averages on a CR10X data logger (Campbell Scientific Inc. (CSI)). A four component pyrano-pyranometer housed in a heated ventilation unit (Kipp and Zonen (KZ) CV2) was used to measure downward ($SW_{\downarrow}$) and upward ($SW_{\uparrow}$) broadband shortwave radiation (KZ CM21) and downward ($LW_{\downarrow}$) and upward ($LW_{\uparrow}$) broadband longwave radiation (KZ CG4). The instruments were mounted initially at a height of 1.8 m (AWS2), sampled every second, and averaged every 10 min in the first year of measurements and every minute thereafter using a CR10X data logger (CSI). Snow temperatures were measured using thermistors at 16 depths and sampled every hour using a 21X data logger (CSI). The initial depths of the near-surface thermistors were 0.1, 0.3, 0.5, 0.75, 1.0, 1.5, 2.0, and 2.5 m below the surface. The remaining thermistors were installed between 3 and 10 m below the surface and separated by 1 m. Eddy covariance data used in this study were collected during two

intensive measurement periods in the summers of 2000 (21 June–6 July 2000) and 2001 (16 June–24 June 2001). Two three-dimensional sonic anemometers (CSI CSAT3) with fine wire thermocouples (FW05) and two ultraviolet hygrometers (CSI KH20) mounted at 1 and 2 m above the snow surface (AWS3) were sampled at 50 Hz using a CR5000 data logger (CSI). The steps taken to postprocess the eddy covariance data are described in detail by Cullen *et al.* [2007b]. Treatment of all other data is briefly described in the following section.

2.2. Data Treatment

Problems with the motor-aspirated radiation shields (Met One, 077) in the second year of measurements resulted in a degradation of the air temperature and relative humidity measurements on AWS1, which led to the decision to use air temperature, relative humidity, and wind speed data from a single height from the GC-Net AWS for modeling purposes over the 2 year measurement period. The disadvantage of this approach is that the GC-Net AWS was located about 2 km to the west of the primary measurement site, but a comparison of the first year of data from both sites showed satisfactory agreement between the two locations. The root-mean-square differences (RMSD) for air temperature, relative humidity, and wind speed obtained at the two sites were 1.7 K, 6.3%, and 0.8 m s^{-1} , respectively. The GC-Net AWS design and correction of data, including the procedure to recalculate relative humidity data to account for saturation with respect to ice rather than liquid water, are given by Box and Steffen [2001]. Because the temporal resolution of GC-Net data was 60 min, all other AWS and eddy correlation data were averaged to the same temporal interval.

A record of surface height change was obtained from weekly measurements collected by technicians from an accumulation forest of 100 stakes [Dibb and Fahnestock, 2004], located approximately 200 m to the south of AWS1. Obtaining access to this remarkable and unique data set of surface height change was one of the primary reasons for revisiting the measurement period used in this study. Although surface height records were available from sonic ranging sensors from AWS1 and the GC-Net AWS (2 sensors) over the same period, they were compromised by data gaps due primarily to riming of sensors during winter. Despite having a robust record of surface height change, determining precipitation from the forest data was not attempted (and not possible on a daily basis); instead, daily precipitation was obtained from the regional climate model MAR [Fettweis *et al.*, 2013a]. Further details about the precipitation data are provided in section 2.3.

To reduce errors associated with the cosine response of the $SW\downarrow$ sensor associated with low Sun angles (high zenith angles), albedo (α) was calculated using an accumulated albedo approach using the ratio of the sum of $SW\uparrow$ and the sum of $SW\downarrow$ over a day [van den Broeke *et al.*, 2004b]. The heated ventilation unit for the four component radiation sensors enabled high-quality radiation data to be obtained; although riming on sensors was not completely avoided, compromising the quality of $SW\downarrow$ and $LW\downarrow$ data. Detection of these events was achieved by assessing irregularities in α (>0.95); $SW\downarrow$ over hourly intervals were calculated as $SW\downarrow = SW\uparrow/\alpha$ (accumulated) on such occasions ($<2\%$ of data). $LW\downarrow$ was parameterized during riming events ($<5\%$ of data) by developing a relationship between $LW\downarrow$ and air temperature and vapor pressure measurements during unaffected days using a quadratic polynomial regression [e.g., Gillett and Cullen, 2011].

2.3. Model Description

The MB model used in this study calculates the climatic MB from three mass gain terms (solid precipitation, surface deposition, and refreezing of liquid water in the snow) and three loss terms (surface and subsurface melt and surface sublimation). As it is described in detail by Mölg *et al.* [2008, 2009, 2012], we only provide a brief summary here. The model resolves the full SEB, as well as the temperature profile in the subsurface by inclusion of energy release from penetrating shortwave radiation and refreezing and energy consumption due to subsurface melting. The surface temperature (T_{sf}) represents the interface between surface and subsurface processes and here is chosen to be solved iteratively [Mölg *et al.*, 2008]. This means that it takes a value such that all the energy fluxes at the surface are in equilibrium every time step, assuming an instantaneously reacting surface “skin” layer. This solution is problematic in climates with a strong diurnal cycle like the tropics [Mölg *et al.*, 2008, 2009] but has proven to be reliable in polar climates [e.g., van As *et al.*, 2005; van den Broeke *et al.*, 2006; Hoffman *et al.*, 2008].

To solve the SEB, we used the maximum number of measurements available to us (section 2.1) to force the model at hourly steps (60 min): $SW\downarrow$, $LW\downarrow$, albedo, air temperature, relative humidity and wind speed at 2 m

Table 2. Changes in Model Inputs Based on Estimated Uncertainty^a

Parameter(ization)	Value/Scheme	References/Note
1 Parameterization of stable condition effect on turbulence	two schemes (scheme 2)	<i>Mölg et al.</i> [2012] ^b
2 Density of fresh snow deposition	$250 \pm 25 \text{ kg m}^{-3}$ (266)	<i>Dibb and Fahnestock</i> [2004]
3 Snow density below stake basis	$450 \pm 50 \text{ kg m}^{-3}$ (437)	<i>Albert and Shultz</i> [2002]
4 Roughness length (snow)	$3.1 \pm 0.7 \cdot 10^{-4} \text{ m}$ (3.5)	<i>Cullen et al.</i> [2007b], <i>Kuipers Munneke et al.</i> [2009] ^c
5 Fraction of $S_{\downarrow}$ ($1 - \alpha$) absorbed in surface layer (snow)	0.765 ± 0.135 (0.64)	<i>Bintanja and van den Broeke</i> [1995], <i>Kuipers Munneke et al.</i> [2009]
6 Extinction of penetrating shortwave radiation (snow)	$17.1 \text{ m}^{-1} \pm 20\%$ (19.4)	<i>Bintanja and van den Broeke</i> [1995], <i>Mölg et al.</i> [2012] ^d
7 Thermal diffusivity (snow)	0.5 ± 0.15 $10^{-6} \text{ m}^2 \text{ s}^{-1}$ (0.57)	<i>Garratt</i> [1992]
8 Fixed bottom temperature	$242.2 \pm 0.2 \text{ K}$ (242.3)	measurement (see text)

^aBold is the structural uncertainty; normal is the parametric uncertainty. For the 1000 model runs, the values are assigned randomly from a uniform distribution. Parameter values in parentheses are from run 444, which is the principal run defined in the evaluation (see Section 3.2).

^bScheme 2 dampens the turbulent flux less than scheme 1.

^cThe lower end value (2.4×10^{-4}) is the representative mean from *Cullen et al.*'s [2007b] data.

^dThe uncertainty was assumed by *Mölg et al.* [2012], but it captures the value suggested by *Greuell and Konzelmann* [1994], which is 20 m^{-1} .

above the surface, and air pressure. This input was used to determine $LW_{\uparrow}$ (from the calculated T_{sfc}), the fluxes of turbulent sensible (QS) and latent (QL) heat, the conductive heat flux (QC) and penetrating shortwave radiation (QPS) in the subsurface (the sum of the two yields the ground heat flux, QG), and the energy flux from precipitation (QPRC). The latter was negligibly small in all runs so it is not analyzed further here. The sign convention used in this study ensures that fluxes directed toward the surface are positive. As discussed in section 2.2, it was not possible to determine daily snowfall amounts reliably from the SR50 data, and other measurements of precipitation at the required temporal resolution were not available. To fully resolve MB, daily precipitation input was sourced from the regional climate model MAR (version 2) for the Summit location [Fettweis et al., 2013a] and was distributed equally over the hours of precipitation days. This approach is not new, with output from regional atmospheric models used successfully in other environments where precipitation data have been difficult to obtain [e.g., Mölg et al., 2012; van Pelt et al., 2012].

At the lower boundary, the model used a fixed temperature at 10 m depth (see below) where measurements indicate an almost constant temperature. The other subsurface layers were at depths of 0, 0.1, 0.2, 0.3, 0.4, 0.5, 0.75, 1, 1.5, 2, 3, 5, 7, and 10 m and remain at these depths throughout a model run. Together with the 60 min time step, this setup delivered numerically stable solutions. The temperature profile for model initialization on 17 June 2000 was also specified from the subsurface measurements. The initial snow density in the uppermost 1.2 m was assigned from snow pit measurements obtained on 12 June 2000. This was the depth at which the stakes from the accumulation forest (see section 2.2) were anchored in the ground [Dibb and Fahnestock, 2004], and thus, it was the relative level to which the stake measurements were capturing surface height changes due to snow compaction. Below the initial 1.2 m depth level, the snow density was held constant, but different values were used in numerous runs (see below) since field data showed that density changes were small, increasing marginally, down to the 10 m depth used as the lower boundary in the model [e.g., Albert and Shultz, 2002]. In the course of a model run, subsurface temperature, snow density (here in the mentioned near-surface layer), and snow depth (and thus surface height) evolve in response to the hourly MB and SEB variability.

The MB model was run with 1000 different parameter combinations. If the model is applied in a spatially distributed mode, the number of variable parameters for this Monte Carlo approach is around 30 [Mölg et al., 2012, 2014]. With the large number of measured input variables used in this study, and only the requirement to run the model in point mode, the number of parameters was reduced to 8 (Table 2). Usually, the “density of

fresh snow deposition” represents the density of falling new snow in the MB model, but here we chose a higher value to account for the effects of wind drift [e.g., *Dibb and Fahnestock*, 2004], which is not resolved explicitly in the model. Measurements of snow density in the uppermost centimeters at Summit, which capture the wind drift effect, are typically in the range $250 \pm 25 \text{ kg m}^{-3}$ [*Albert and Shultz*, 2002; *Dibb and Fahnestock*, 2004, Table 2]. The bottom temperature was specified from the available hourly measurements at 10 m depth, which showed a mean of 242.2 K and a standard deviation of only 0.2 K. Sensitivity runs with a deeper model bottom (15 m) showed that the mean conductive heat flux at 10 m depth is indeed negligible (0.03 W m^{-2}). The parameter variations (Table 2) also absorb the uncertainty of the measured input [e.g., *Mölg et al.*, 2012], which we are not accounting for separately.

To carefully assess the performance of the MB model, the 1000 model results were compared to (i) measured radiative surface temperature (T_{sfc}), (ii) measured surface height change, (iii) measured subsurface temperature profiles in summer 2000 and summer 2001, and (iv) eddy correlation measurements. While we found that subsurface temperatures can be modeled more accurately if the number of subsurface layers is increased, which has also been reported in other modeling studies [*Dadic et al.*, 2008], the chosen layer structure is a compromise between the computational demand of 1000 runs and a reasonable representation of the subsurface temperature field. Further details about the modeling evaluation are provided in section 3.2

3. Results

3.1. Meteorological Conditions

The atmospheric conditions at Summit are unique by virtue of its high altitude and latitude. The polar night (defined as $SW\downarrow$ at the top of the atmosphere (TOA) = 0) lasts 87 days between 9 November and 3 February (Figure 1a). The average $SW\downarrow$ during the measurement period was 146 W m^{-2} , which is slightly higher than reported by *van den Broeke et al.* [2008b] for Summit during the period July 2000 to July 2002 and June 2003 to December 2004 (136 W m^{-2}) using ETH data. The physical processes controlling the transmission of $SW\downarrow$ have not been fully investigated but are dependent on atmospheric mass, water vapor content, solar zenith angle, surface albedo (multiple scattering between snow surface and cloud), and cloud optical thickness. The fraction of $SW\downarrow$ that reached the surface compared to TOA was 0.76, which is consistent with *Hoch* [2005] but is higher than reported by others [e.g., *Konzelmann and Ohmura*, 1995; *van den Broeke et al.*, 2008b]. Because of the high albedo at Summit (average 0.82), only 14% of $SW\downarrow$ at TOA is absorbed at the surface.

The variability in atmospheric temperature and emissivity, as captured by $LW\downarrow$, was controlled by changes in cloud cover (Figure 1b). Despite a seasonal variability in $LW\downarrow$, high values during the polar night were not uncommon, demonstrating the importance of cloud cover as an energy source during periods without sunlight. These events are also evident in the air temperature and humidity records (Figures 1c and 1d), which like $LW\downarrow$ also show distinct seasonal variability. The mean air temperature was -29.1°C , which is the same as reported by *Steffen and Box* [2001]. An air temperature reconstruction by *Shuman et al.* [2001] for the period 1987–1999 also established the same mean for Summit. Low air temperatures during winter result in the atmosphere having very little capacity to hold moisture (Figure 1d), while during summer, vapor pressures reached a maximum of just over 5 hPa. The mean vapor pressure at Summit was less than 1 hPa, which is extremely low even when compared to coastal stations in Greenland.

The mean (maximum) wind speed was 4.3 (26.4) m s^{-1} , with the strongest winds occurring during winter (Figure 1e), while wind directions at Summit were mostly between the SSE and SSW sectors (not shown). Precipitation as determined from the regional climate model MAR was $235 \text{ mm WE yr}^{-1}$ (Figure 1f), which is slightly less (15 mm WE yr^{-1}) than estimates of accumulation obtained from density measurements from snow pits in the vicinity of AWS1 during the measurement period. *Ohmura et al.* [1999] reported that the average accumulation for the dry snow zone was $195 \text{ mm WE yr}^{-1}$, while density profiles obtained monthly from shallow snow pits revealed that 250 mm WE accumulated at Summit during the period August 2001 to August 2002 [*Dibb and Fahnestock*, 2004].

3.2. Model Evaluation

The performance of the MB model to reproduce observed surface temperature and surface height change using 1000 different parameter combinations is shown in Figure 2. It is crucial that modeled surface temperature is carefully reproduced as it determines all surface fluxes except net shortwave

Figure 1. The seasonal variability in (a) incoming shortwave radiation (and TOA), (b) incoming longwave radiation, (c) air temperature (T_a), (d) vapor pressure (e_a), (e) wind speed (v), and (f) precipitation. The black lines are the hourly data, the dark green lines show a running average (weekly), and the light gray shaded values in Figure 1a are TOA.

radiation (SW_{NET}), QPS , and $LW\downarrow$. Importantly, Figure 2 (top) shows that modeled surface temperatures are in good agreement with observations, with the RMSD equal to 1.7 K. The small range in uncertainty (0.65 K) in each of the simulations demonstrates that different parameter combinations did not lead to strong negative or positive biases in the modeled surface temperature. The lowest RMSD (1.5 K) was observed in model run 444, which is used to characterize the energy and mass balance in the following section. Run 444 does not show a combination of extreme parameter values (Table 2), which is important in the context of the lowest RMSD. The mean surface temperature for this principal model run is -31.7°C , which is slightly lower than both the observed mean, assuming an emissivity of 1 (-30.3°C) and a reconstruction of average snow temperature at Summit over the past 4000 years that was estimated to be -30.7°C , with a standard deviation of 1.0°C [Kobashi *et al.*, 2011]. Importantly, the temperature gradient between atmosphere and surface typically leads to a strong stability at Summit, although periods of instability in summer do occur [e.g., Cullen and Steffen, 2001].

The MB model also reproduces surface height change, as determined from stake measurements, very well (Figure 2, bottom). The mean RMSD for all runs is only 5.7 cm, which could easily fall within the expected uncertainty associated with the stake forest not being at the same location as the model input data. The MB model shows less increase in surface height during the summer periods as compared to observations, which is difficult to provide a single explanation for, but may also be related to the treatment of compaction in the MB model [Mölg *et al.*, 2012]. Dibb and Fahnestock [2004] observed rapid compaction of surface snow in late spring and early summer compared to other periods, which did result in smaller changes in surface height during the summers of 2001 and 2002 despite observations showing higher than average snowfalls during these periods. The precipitation data used in this study does not show marked increases during summer (Figure 1f), which could also help explain the discrepancy.

Figure 2. (top) Root-mean-square difference (RMSD) between measured and modeled hourly surface temperatures in 1000 model realizations, 17 June 2000 to 18 June 2002 ($N = 17,568$ h). (bottom) Accumulated surface height change in the 1000 model runs and from the stake record relative to 8 July 2000 (first stake point).

temperatures greater than 245 K sits slightly “above” the measured values, suggesting that modeled subsurface temperatures are slightly lower than those observed. This is possibly favored by the usage of a monochromatic extinction coefficient (Table 2) in the model [Mölg *et al.*, 2008]. Nevertheless, the comparison supports the implementation of a penetrating shortwave radiation module in the MB model as suggested by Kuipers Munneke *et al.* [2009]. Without activation of penetrating radiation, the mean modeled subsurface temperature profile (run 444) is 3.4 K and 3.8 K lower on 30 June 2000 and 30 June 2001, respectively.

Figure 3. Measured and modeled (1000 runs) subsurface temperature on 30 June 2000 (14 days into the run) and on 30 June 2001 (379 days into the run, last available day of the subsurface measurement).

Figure 3 shows the comparison between modeled (1000 runs) and observed (temperature string) subsurface temperatures for 30 June in the summers of 2000 and 2001. As shown in Figure 3 (top), subsurface temperatures were reproduced very well at the start of the measurement period, which could only be achieved by allowing a fraction of SW_{NET} to penetrate into the subsurface. As noted by Kuipers Munneke *et al.* [2009], the penetration of shortwave radiation into the snowpack at Summit provides an additional mechanism to deliver energy below the surface, enabling a more rapid transfer of energy into the snow during summer. Despite the depth of the temperature string measurements being less certain after 12 months of accumulation, Figure 3 (bottom) shows that there is a satisfactory agreement between measured and modeled subsurface temperatures 379 days into the model run. The model envelope at

Given the emphasis on water vapor exchanges, in this study, modeled QL is compared to measured values, which were obtained using eddy correlation instruments 2 m above the surface during the summers of 2000 and 2001 [Cullen and Steffen, 2001; Cullen *et al.*, 2007b]. Even though there is a negative bias in modeled QL over all 1000 simulations (mean measured: -1.5 W m^{-2} , mean modeled: -3.5 W m^{-2}) (Figure 4), the performance of the MB model in reproducing the term responsible for water vapor exchange is encouraging. For example, the mean RMSD (3.8 W m^{-2}) is less than that observed by Cullen *et al.* [2007a] over a cold-based glacier on Kilimanjaro, while the correlation coefficient ($r = 0.85$) is higher than that obtained by Kuipers Munneke *et al.* [2009] for QS at Summit for a short duration in summer. The RMSD for the

Figure 4. Measured and modeled hourly turbulent latent heat flux ($N = 523$ h). For the model, all 1000 runs are shown in gray, and the black line indicates the 1:1 line. Statistics are the mean from the 1000 comparisons. The black dots show run 444.

model run used to characterize the energy and mass balance in this study (run 444) compares more favorably (3.0 W m^{-2}) than the mean from all 1000 runs, while the correlation coefficient ($r = 0.84$) is very close to the average. This gives us confidence in our assessment of QL and critically the variability in sublimation and deposition over the measurement period. However, the remaining 999 runs are also considered further in order to characterize the uncertainty of key results, including the mean for QL presented in the following section, as there is no group of runs that shows an unrealistic behavior in the evaluation steps above.

3.3. Seasonal Variability in Energy Exchanges

The seasonal variability in the energy balance at Summit is shown in Figure 5, while monthly changes in individual radiation and energy

balance terms are given in Table 3. The surface radiation budget in summer is characterized by some of the highest $SW_{\downarrow}$ values in the northern hemisphere due to the 24 hour sunlight and persistent clear skies (Table 3). However, the impact of global radiation on the energy balance is to a large extent limited because the dry snow at Summit is so reflective, with SW_{NET} only close to or above 60 W m^{-2} for 3 months yr^{-1} (May, June, and July). Though not investigated further here, the combined effect of changes in cloudiness and albedo are understood to play a role in SW_{NET} peaking in June at Summit [van den Broeke *et al.*, 2008b]. In the absence of sunlight, radiative exchanges are restricted to the infrared region of the electromagnetic spectrum (Figure 5). Despite the very low sky and surface temperatures during winter, the energy loss from net longwave radiation (LW_{NET}) is markedly less than summer, when a rapid increase in surface temperature results in high $LW_{\uparrow}$ values (Table 3). This is important as it restricts the effect of increases in $SW_{\downarrow}$ during summer, with monthly net all wave radiation (R_{NET}) values never exceeding 20 W m^{-2} . R_{NET} is positive between May and August (energy gain to the surface), but the surface is otherwise in radiative deficit, with the average annual loss equal to -7 W m^{-2} over the 2 year period.

The primary energy balance term to compensate for the radiative loss at the surface at Summit is QS, which is, on average, directed toward the surface during all months of the year (Table 3). The magnitude of QS is

Figure 5. Monthly variation in energy balance components between June 2000 and June 2002.

Table 3. Mean Monthly Variability in Atmospheric, Radiation, and Energy Balance Variables^a

Variable	J	F	M	A	M	J	J	A	S	O	N	D	2YR
T_a	−38.5	−44.1	−40.7	−32.7	−23.0	−15.9	−14.3	−14.3	−21.8	−32.7	−37.5	−35.7	−29.1
e	0.2	0.2	0.2	0.4	0.9	1.4	1.7	1.9	1.0	0.4	0.3	0.4	0.7
v	5.2	4.8	4.1	4.6	4.3	3.6	4.3	3.5	4.6	3.5	4.3	4.8	4.3
p	652.9	653.7	659.0	663.1	672.6	679.0	678.2	676.2	670.6	661.9	659.0	662.9	665.9
TOA	0.0	19.2	111.1	255.7	409.3	494.8	455.9	315.7	165.0	48.5	1.3	0.0	191.5
SW↓	0.0	13.4	91.3	210.2	336.8	394.8	334.8	220.9	104.2	30.1	1.1	0.0	146.2
SW↑	0.0	−10.9	−74.6	−173.8	−275.1	−321.1	−275.0	−183.0	−87.3	−24.3	−0.9	0.0	−120.0
SW _{NET}	0.0	2.4	16.7	36.5	61.8	73.7	59.8	37.8	17.0	5.8	0.2	0.0	26.2
LW↓	139.3	122.1	122.4	137.0	164.7	190.1	208.2	222.6	199.4	158.1	145.6	153.2	163.9
LW↑	−161.5	−148.5	−156.7	−182.4	−218.6	−245.7	−252.4	−252.4	−223.4	−181.1	−165.1	−172.5	−197.1
LW _{NET}	−22.3	−26.4	−34.4	−45.4	−53.8	−55.6	−44.2	−29.8	−24.1	−22.9	−19.5	−19.3	−33.2
R _{NET}	−22.3	−24.0	−17.6	−8.9	7.9	18.1	15.6	8.1	−7.1	−17.1	−19.3	−19.3	−7.0
QS	16.1	15.9	13.6	13.3	5.7	2.4	1.9	2.8	8.1	10.7	12.3	13.1	9.6
QL	1.2	0.8	0.8	1.3	0.3	−2.5	−2.9	−0.6	2.0	1.1	1.4	1.6	0.4
QG	4.9	7.3	3.2	−5.7	−13.9	−18.0	−14.6	−10.3	−3.1	5.3	5.5	4.5	−3.0

^aThe monthly values were calculated from all available hourly data ($N = 17,568$ h) for each month interval over the 2 year period (17 June 2000 to 18 June 2002). Averaged hourly data over the entire measurement period is also given (2YR). The months start with January (J) and finish with December (D). Atmospheric variable names are defined in the caption of Figure 1, while all radiation and energy balance variable names are given in the text. All variables are rounded to one decimal place.

greatest during winter (Figure 5), when strongly stable atmospheric conditions lead to persistent temperature inversions. During summer, the atmosphere is often unstable during the daytime [Cullen and Steffen, 2001], but as a monthly average, the heat gradient is still toward the surface, although the diurnal change in the sign of QS results in small monthly values. The largest nonradiative flux away from the surface during summer is QG , which is the sum of the conductive heat flux (QC) and penetrating shortwave radiation (QPS). In the absence of sunlight, the subsurface is a heat source through conduction (QC), with QG being the largest nonradiative term after QS during winter. The heat exchanges associated with the change in phase of water are very small in all seasons, with average QL over the measurement period equal to $0.4 (\pm 0.2) \text{ W m}^{-2}$, where the uncertainty is equal to the standard deviation of all 1000 runs. It should be noted that QL to two decimal precision is $0.37 (\pm 0.19) \text{ W m}^{-2}$, which is used to calculate mass gain in the following section. This small, positive value in QL suggests that on average, the water vapor gradient is toward the surface, resulting in a net gain of mass annually. Figure 5 shows that the water vapor gradient is toward the surface during winter (deposition), while during summer, a small amount is lost to the atmosphere (sublimation). While QL is on average not important energetically, we will explore its role in mass gain and loss at Summit and the dry snow zone more broadly in the following section.

4. Discussion

The calculated MB over the 2 year measurement period was 478.8 mm WE , with total solid precipitation equal to 470.6 mm WE . The mass gain through deposition of water vapor ($4.1 \text{ mm WE yr}^{-1}$) represents only a small fraction of the total accumulation ($<2\%$), in part because of the change in sign of the water vapor flux from winter (deposition) to summer (sublimation). The mass gain (loss) from deposition (sublimation) over the entire period (2 years) was $26.4 (18.2) \text{ mm WE}$, which if assumed to be representative of the entire dry snow zone (40% of the total ice sheet area) corresponds to a mass gain of 5.5 Gt yr^{-1} (this is calculated using an ice sheet area equal to the “common mask” defined by Vernon *et al.* [2012, p. 603]). There is little doubt that the water vapor exchange on the top of the GrIS has a seasonal cycle (Figure 5), but uncertainty exists about the magnitude of QL in summer versus winter, which ultimately controls whether there is a mass gain or loss associated with the annual flux in the dry snow zone. For example, evaporation estimates from high-resolution GCM experiments (ECHAM3 above 1500 m) by Ohmura *et al.* [1999] indicated that water vapor is lost to the atmosphere in the interior of the GrIS over an annual cycle. These modeling results were later supported by Hoch [2005] from an observational period similar to the one presented in this study, with mass loss from sublimation (32 mm WE yr^{-1}) at Summit thought to remove about 10% of the mass gain from annual precipitation.

Figure 6. Histograms of the mean turbulent (left) latent heat flux and daily mass flux of (middle) sublimation and (right) deposition for the study period (gray bars) and the warming scenario (open bars). For each histogram, the range of the 1000 model runs is divided in 10 equal bins.

The challenge of calculating robust estimates of sublimation in the dry snow zone from automated weather station observations is demonstrated by *Box and Steffen* [2001], who used two methods (one (bulk) and two (profile) measurement levels) to characterize seasonal variability in water vapor exchange. As discussed by the authors, the method chosen can influence both the sign and magnitude of the annual sublimation calculated for sites in the dry snow zone. At Summit, the one-level method resulted in a mass loss (10 mm WE yr^{-1}), while application of the two-level method (profile of instruments) produced a mass gain (18 mm WE yr^{-1}) [*Box and Steffen*, 2001, see Table 6]. There is a better agreement in the ablation areas, where mass loss through evaporation and sublimation is larger than any uncertainty associated with method. A careful assessment of the surface MB of the GrIS using a regional climate model (RCM) by *Box et al.* [2006] also involved comparing modeled sublimation values to those determined from AWS data. The evaluation helped identify a negative (deposition) bias in the RCM (Polar MM5) due to overly strong surface temperature inversions, which was corrected based on the two-level method to calculate the water vapor flux [*Box and Steffen*, 2001]. With correction, the Polar MM5 water vapor flux in the dry snow zone of the GrIS was found to be directed toward the surface [*Box et al.*, 2006, Figure 7], with values at Summit similar or slightly less than that reported by *Box and Steffen* [2001] using the two-level method. Other recent regional climate modeling efforts are in agreement with the average annual net surface water vapor flux resulting in a mass gain (deposition) on the higher parts of the ice sheet [e.g., *Ettema et al.*, 2010b; *Fettweis et al.*, 2013a].

It could be argued that the cancelling effect of the water vapor flux in the dry snow zone, from deposition during winter to sublimation in summer, makes the term somewhat unimportant in determining MB, especially as mass loss from drifting snow sublimation is likely to be as large or greater than the losses from surface sublimation [e.g., *Lenaerts et al.*, 2012]. However, as pointed out by *van den Broeke et al.* [2004a, p. 571], when assessing the surface MB in Antarctica, small mean turbulent latent heat fluxes can still be important for the mass “household” of large ice sheets. If QL is positive or negative 1 W m^{-2} annually, it can either add or remove about 11 mm WE yr^{-1} of snow. Importantly, if the sign of an annual water vapor flux with a magnitude equal to 1 W m^{-2} is found to be directed toward the surface (deposition) over the entire dry snow zone of the GrIS, when in fact it should be the same magnitude away from the surface (sublimation), the difference in volume is approximately 15 Gt yr^{-1} . Thus, the sign and magnitude of the small QL values reported by different authors for Summit and elsewhere in the dry snow zone of the GrIS are important and require careful scrutiny.

To evaluate our own estimate of QL , we show the full distribution of values obtained from the 1000 different parameter combinations (Figure 6), which reveals that during the measurement period, the mean annual values range between $\pm 0.5 \text{ W m}^{-2}$, with the largest count slightly positive (deposition) and equal in magnitude to the primary model run used in this study (run 444). It is clear that the water vapor flux at Summit is most likely small but can result in either a mass gain or loss. Figure 6 also shows that the range of values for deposition is more tightly constrained than those for sublimation, suggesting that mass loss in summer might be more vulnerable to changes in climate. To explore this, we assess the qualitative change to the water vapor flux (Figure 6) in a warming climate by considering the extreme summer in 2012, where surface melt extended across almost the entire ice sheet [*Nghiem et al.*, 2012].

In general, anomalously high-pressure conditions and advection of warm air over Greenland were the major contributors to the exceptional melt season in 2012 [Fettweis *et al.*, 2013b; Hanna *et al.*, 2013b]. At Summit, the presence of optically thin and liquid water-containing clouds was shown to play a key role in the modification of the near-surface moisture and radiation budget and thus the near-surface air temperature [Bennartz *et al.*, 2013]. Further, albedo lowering must be expected during enhanced surface melt in the course of a positive feedback [Franco *et al.*, 2013]. In light of this evidence, we repeated the 1000 model runs but modified the following input: $SW\downarrow$, $LW\downarrow$, surface albedo, and 2 m air temperature and relative humidity. This was simply done by adding to every day of the study period the respective daily mean differences of these variables between 2012 and 2000–2002, which occur in the MAR data at Summit [Fettweis *et al.*, 2013a].

The results illustrate that QL is shifted toward more negative values in a warming atmosphere (Figure 6), with sublimation increasing on average by $4.5 \text{ mm WE yr}^{-1}$. However, deposition increases as well ($3.2 \text{ mm WE yr}^{-1}$) due to higher water vapor pressure in the air, but the increase is less than for sublimation resulting in mean QL shifting from a positive (mass gain) to a slightly negative (mass loss) value in the warming scenario. Since surface roughness and wind speed were not altered in the warming experiment, one logical explanation for the increase in mass turnover (Figure 6) is a change in the gradient of water vapor pressure between surface and the air (here 2 m). For the modeled 2000–2002 conditions, this gradient amounts to -0.12 hPa m^{-1} and barely changes in the warming scenario (-0.10 hPa m^{-1}) during sublimation, but an increase in the gradient is noticeable during deposition (from 0.04 to 0.10 hPa m^{-1}). A second possible reason is a change in the thermodynamic stratification of the near-surface layer, and this factor in turn clearly enhances sublimation due to greater instability in the warming scenario (the bulk Richardson number changes from -5.6×10^{-3} to -1.2×10^{-2}). During deposition, by contrast, the stability increases in the warming scenario and restricts changes to the water vapor flux (the Richardson number changes from 2.4×10^{-2} to 3.1×10^{-2}). Regardless of the relative contributions of gradient and stability changes, the warming scenario suggests that the mass fluxes associated with QL will, at the very least, reduce the mass budget at Summit by a few mm yr^{-1} if warming continues in the future. This reduction in the mass budget has the potential to be larger should warming lead to a long-term shift from net deposition to sublimation in the dry snow zone.

5. Conclusions

Two years of meteorological and glaciological data collected at Summit allowed us to characterize the seasonal variability of mass and energy exchanges on top of the GrIS. The high-quality data from 2000 to 2002 were used as input to run a climatic MB model, with a Monte Carlo approach using 1000 different parameter combinations adopted to investigate parametric and structural uncertainties. The realism of the 1000 model results was supported by measured surface and subsurface temperatures, changes in surface height, and eddy correlation data. The subsurface temperatures could not be reproduced without allowing a fraction of SW_{NET} to penetrate into the surface as observed by Kuipers Munneke *et al.* [2009].

The high reflectivity of the dry snow (average 0.82) restricts the impact of $SW\downarrow$ on the energy balance at Summit during summer, with SW_{NET} only close to or greater than 60 W m^{-2} in May, June, and July. The rapid increase in surface temperatures in summer results in high $LW\uparrow$ values, with energy loss from LW_{NET} largest in June. This restricts the effect of increases in $SW\downarrow$ in summer, with R_{NET} small ($< 20 \text{ W m}^{-2}$) but positive between May and August. Radiative exchanges are restricted to the infrared part of the electromagnetic spectrum during the polar night, with the magnitude of $LW\uparrow$ exceeding $LW\downarrow$ (radiation deficit). A persistent temperature inversion during winter results in QS being directed toward the surface and is the primary energy balance term to compensate for the annual deficit in R_{NET} , which over the 2 year measurement period was -7 W m^{-2} . The atmosphere during summer is often characterized by a smaller surface air gradient in temperature, reducing the magnitude of QS , with QG becoming the largest nonradiative energy balance term. The energy associated with QL results in mass loss during summer (sublimation) and mass gain in winter (deposition), with the difference equal to a gain of $4.1 \text{ mm WE yr}^{-1}$. If this observed deposition rate is assumed to be representative of the entire dry snow zone (40% of the total ice sheet area), the net gain in mass is expected to be approximately 5.5 Gt yr^{-1} .

Even though annual mass gain from deposition represents only a small fraction of the snow accumulation ($<2\%$), a change in the sign and magnitude of the water vapor flux over the entire dry snow zone has the potential to have a noticeable effect on MB. Assessment of all 1000 model runs suggests that mean annual QL values are most likely slightly positive (deposition), but it cannot be completely ruled out that sublimation may occur on an annual basis. A warming scenario based on 2012 atmospheric conditions, where an extreme melt event took place, suggests that QL will be shifted toward more negative (sublimation) values, which could result in a change in sign in the annual water vapor flux from deposition (mass gain) to sublimation (mass loss). To fully resolve whether sublimation will in fact dominate water vapor exchanges in a future that is characterized by warming, there is a need to obtain higher quality humidity measurements in the dry snow zone, in particular in winter, to allow the physical processes controlling the deposition of mass during the polar night to be examined. Determining the magnitude and direction of the water vapor flux in the dry snow zone of the GrIS will also not be possible without further efforts to understand the role of drifting snow sublimation [e.g., *Lenaerts et al.*, 2012], which has not been explicitly accounted for in this study.

Acknowledgments

This research has been supported by the Department of Geography, University of Otago, New Zealand. We are grateful for the data provided by Jack Dibb and Xavier Fettweis. NSF Office of Polar Programs provided funding for the field work related to this research. Data used in this research are available upon request to the corresponding author.

References

- Albert, M. R., and R. L. Hawley (2000), Seasonal differences in surface energy exchange and accumulation at Summit, Greenland, *Ann. Glaciol.*, **31**, 387–390.
- Albert, M. R., and E. F. Shultz (2002), Snow and firn properties and air-snow transport processes at Summit, Greenland, *Atmos. Environ.*, **36**, 2789–2797.
- Bales, R. C., J. R. McConnell, E. Mosley-Thompson, and B. Csatho (2001a), Accumulation over the Greenland ice sheet from historical and recent records, *J. Geophys. Res.*, **106**, 33,813–33,825, doi:10.1029/2001JD900153.
- Bales, R. C., E. Mosley-Thompson, and J. R. McConnell (2001b), Variability of accumulation in northwest Greenland over the past 250 years, *Geophys. Res. Lett.*, **28**, 2679–2682, doi:10.1029/2000GL011634.
- Bales, R. C., Q. Guo, D. Shen, J. R. McConnell, G. Du, J. F. Burkhart, V. B. Spikes, E. Hanna, and J. Cappelen (2009), Annual accumulation for Greenland updated using ice core data developed during 2000–2006 and analysis of daily coastal meteorological data, *J. Geophys. Res.*, **114**, D06116, doi:10.1029/2008JD011208.
- Bamber, J. L., et al. (2013), A new bed elevation dataset for Greenland, *Cryosphere*, **7**, 499–510, doi:10.5194/tc-7-499-2013.
- Bennartz, R., M. D. Shupe, D. D. Turner, V. P. Walden, K. Steffen, C. J. Cox, M. S. Kulie, N. B. Miller, and C. Pettersen (2013), July 2012 Greenland melt extent enhanced by low-level liquid clouds, *Nature*, **496**, 83–86, doi:10.1038/nature12002.
- Bintanja, R., and M. R. van den Broeke (1995), The surface energy balance of Antarctic snow and blue ice, *J. Appl. Meteorol.*, **34**, 902–926.
- Bottenheim, J. W., J. E. Dibb, R. E. Honrath, and P. B. Shepson (2002), An introduction to the Alert 2002 and Summit 2000 Arctic research studies, *Atmos. Environ.*, **36**, 2467–2469.
- Box, J. E., and K. Steffen (2001), Sublimation on the Greenland ice sheet, *J. Geophys. Res.*, **106**, 33,965–33,982, doi:10.1029/2001JD900219.
- Box, J. E., D. H. Bromwich, and L.-S. Bai (2004), Greenland ice sheet surface mass balance 1991–2000: Application of Polar MM5 mesoscale model and in situ data, *J. Geophys. Res.*, **109**, D16105, doi:10.1029/2003JD004451.
- Box, J. E., D. H. Bromwich, B. A. Veenhuis, L. S. Bai, J. C. Stroeve, J. C. Rogers, K. Steffen, T. Haran, and S. H. Wang (2006), Greenland ice sheet surface mass balance variability (1988–2004) from calibrated polar MM5 output, *J. Clim.*, **19**, 2783–2800.
- Box, J. E., et al. (2013), Greenland ice sheet mass balance reconstruction. Part I: Net snow accumulation (1600–2009), *J. Clim.*, **26**, 3919–3934, doi:10.1175/JCLI-D-12-00373.1.
- Calanca, P., H. Gilgen, S. Ekholm, and A. Ohmura (2000), Gridded temperature and accumulation distributions for Greenland for use in cryospheric models, *Ann. Glaciol.*, **31**, 118–120, doi:10.3189/172756400781820345.
- Cullen, N. J. (2003), Characteristics of the atmospheric boundary layer at Summit, Greenland, PhD dissertation, Univ. of Colorado at Boulder.
- Cullen, N. J., and K. Steffen (2001), Unstable near-surface boundary conditions in summer on top of the Greenland ice sheet, *Geophys. Res. Lett.*, **28**, 4491–4493, doi:10.1029/2001GL013417.
- Cullen, N. J., T. Mölg, G. Kaser, K. Steffen, and D. R. Hardy (2007a), Energy balance model validation on the top of Kilimanjaro, Tanzania, using eddy covariance data, *Ann. Glaciol.*, **46**, 227–233.
- Cullen, N. J., K. Steffen, and P. D. Blanken (2007b), Nonstationarity of turbulent heat fluxes at Summit, Greenland, *Boundary Layer Meteorol.*, **122**, 439–455.
- Dadic, R., M. Schneebeli, M. Lehning, M. A. Hutterli, and A. Ohmura (2008), Impact of the microstructure of snow on its temperature: A model validation with measurements from Summit, Greenland, *J. Geophys. Res.*, **113**, D14303, doi:10.1029/2007JD009562.
- Dethloff, K., M. Schwager, J. H. Christensen, S. Kiilsholm, A. Rinke, W. Dorn, F. Jung-Rothenhäusler, H. Fischer, S. Kipfstuhl, and H. Miller (2002), Recent Greenland accumulation estimated from regional climate model simulations and ice core analysis, *J. Clim.*, **15**, 2821–2832.
- Dibb, J. E., and M. Fahnestock (2004), Snow accumulation, surface height change, and firn densification at Summit, Greenland: Insights from 2 years of in situ observation, *J. Geophys. Res.*, **109**, D24113, doi:10.1029/2003JD004300.
- Drue, C., and G. Heinemann (2007), Characteristics of intermittent turbulence in the upper stable boundary layer over Greenland, *Boundary Layer Meteorol.*, **124**, 361–381.
- Enderlin, E. M., I. M. Howat, S. Jeong, M.-J. Noh, J. H. van Angelen, and M. R. van den Broeke (2014), An improved mass budget for the Greenland ice sheet, *Geophys. Res. Lett.*, **41**, 866–872, doi:10.1002/2013GL059010.
- Ettema, J., M. R. van den Broeke, E. van Meijgaard, W. J. van de Berg, J. E. Box, and K. Steffen (2010a), Climate of the Greenland ice sheet using a high-resolution climate model – Part 1: Evaluation, *Cryosphere*, **4**, 511–527, doi:10.5194/tc-4-511-2010.
- Ettema, J., M. R. van den Broeke, E. van Meijgaard, and W. J. van de Berg (2010b), Climate of the Greenland ice sheet using a high-resolution climate model – Part 2: Near-surface climate and energy balance, *Cryosphere*, **4**, 529–544, doi:10.5194/tc-4-529-2010.
- Fettweis, X. (2007), Reconstruction of the 1979–2006 Greenland ice sheet surface mass balance using the regional climate model MAR, *Cryosphere*, **1**, 21–40, doi:10.5194/tc-1-21-2007.

- Fettweis, X., B. Franco, M. Tedesco, J. H. van Angelen, J. T. M. Lenaerts, M. R. van den Broeke, and H. Gallée (2013a), Estimating the Greenland ice sheet surface mass balance contribution to future sea level rise using the regional atmospheric climate model MAR, *Cryosphere*, 7, 469–489, doi:10.5194/tc-7-469-2013.
- Fettweis, X., E. Hanna, C. Lang, A. Belleflamme, M. Epicum, and H. Gallée (2013b), Brief communication “Important role of the midtropospheric atmospheric circulation in the recent surface melt increase over the Greenland ice sheet”, *Cryosphere*, 7, 241–248, doi:10.5194/tc-7-241-2013.
- Forster, R. R., et al. (2014), Extensive liquid meltwater storage in firn within the Greenland ice sheet, *Nat. Geosci.*, 7, 95–98, doi:10.1038/NGEO2043.
- Franco, B., X. Fettweis, and M. Epicum (2013), Future projections of the Greenland ice sheet energy balance driving the surface melt, *Cryosphere*, 7, 1–18, doi:10.5194/tc-7-1-2013.
- Garratt, J. R. (1992), *The Atmospheric Boundary Layer*, 316 pp., Cambridge Univ. Press, Cambridge.
- Gillett, S., and N. J. Cullen (2011), Atmospheric controls on summer ablation over Brewster Glacier, New Zealand, *Int. J. Climatol.*, 31, 2033–2048, doi:10.1002/joc.2216.
- Greuell, W., and T. Konzelmann (1994), Numerical modeling of the energy balance and the englacial temperature of the Greenland Ice Sheet, calculation for the ETH-Camp location (West Greenland, 1155 m asl), *Global Planet. Change*, 9, 91–114.
- Hanna, E., P. Valdes, and J. McConnell (2001), Patterns and variations of snow accumulation over Greenland, 1979–98, from ECMWF analyses, and their verification, *J. Clim.*, 14, 3521–3535.
- Hanna, E., J. McConnell, S. Das, J. Cappelen, and A. Stephens (2006), Observed and modelled Greenland ice sheet snow accumulation, 1958–2003, and links with regional climate forcing, *J. Clim.*, 19, 344–358, doi:10.1175/JCLI3615.1.
- Hanna, E., P. Huybrechts, K. Steffen, J. Cappelen, R. Huff, C. Shuman, T. Irvine-Fynn, S. Wise, and M. Griffiths (2008), Increased runoff from melt from the Greenland ice sheet: A response to global warming, *J. Clim.*, 21, 331–341.
- Hanna, E., et al. (2011), Greenland Ice Sheet surface mass balance 1870 to 2010 based on Twentieth Century Reanalysis, and links with global climate forcing, *J. Geophys. Res.*, 116, D24121, doi:10.1029/2011JD016387.
- Hanna, E., J. M. Jones, J. Cappelen, S. H. Mernild, L. Wood, K. Steffen, and P. Huybrechts (2013a), The influence of North Atlantic atmospheric and oceanic forcing effects on 1900–2010 Greenland summer climate and ice melt/runoff, *Int. J. Climatol.*, 33, 862–880, doi:10.1002/joc.3475.
- Hanna, E., X. Fettweis, S. H. Mernild, J. Cappelen, M. H. Ribergaard, C. A. Shuman, K. Steffen, L. Wood, and T. L. Mote (2013b), Atmospheric and oceanic climate forcing of the exceptional Greenland ice sheet surface melt in summer 2012, *Int. J. Climatol.*, doi:10.1002/joc.3743.
- Harper, J., N. Humphrey, W. T. Pfeffer, J. Brown, and X. Fettweis (2012), Greenland ice-sheet contribution to sea-level rise buffered by meltwater storage in firn, *Nature*, 491, 240–243, doi:10.1038/nature11566.
- Hoch, S. W. (2005), Radiative flux divergence in the surface boundary layer - a study based on observations at Summit, Greenland, PhD dissertation, No. 16194, Swiss Federal Institute of Technology (ETH), Zurich.
- Hoch, S. W., P. Calanca, R. Philipona, and A. Ohmura (2007), Year-round observation of longwave radiative flux divergence in Greenland, *J. Appl. Meteorol.*, 46, 1469–1479, doi:10.1175/JAM2542.1.
- Hoffman, M. J., A. G. Fountain, and G. E. Liston (2008), Surface energy balance and melt thresholds over 11 years at Taylor Glacier, Antarctica, *J. Geophys. Res.*, 113, F04014, doi:10.1029/2008JF001029.
- Howat, I. M., I. Joughin, and T. A. Scambos (2007), Rapid changes in ice discharge from Greenland outlet glaciers, *Science*, 315, 1559–1561.
- Joughin, I., B. E. Smith, I. M. Howat, T. Scambos, and T. Moon (2010), Greenland flow variability from ice-sheet-wide velocity mapping, *J. Glaciol.*, 56(197), 415–430.
- Kobashi, T., K. Kawamura, J. P. Severinghaus, J.-M. Barnola, T. Nakaegawa, B. M. Vinther, S. J. Johnsen, and J. E. Box (2011), High variability of Greenland surface temperature over the past 4000 years estimated from trapped air in an ice core, *Geophys. Res. Lett.*, 38, L21501, doi:10.1029/2011GL049444.
- Konzelmann, T., and A. Ohmura (1995), Radiative fluxes and their impact on the energy balance of the Greenland ice sheet, *J. Glaciol.*, 41, 490–502.
- Kuipers Munneke, P., M. R. van den Broeke, C. H. Reijmer, M. M. Helsen, W. Boot, M. Schneebeli, and K. Steffen (2009), The role of radiation penetration in the energy budget of the snowpack at Summit, Greenland, *Cryosphere*, 3, 155–165, doi:10.5194/tc-3-155-2009.
- Lenaerts, J. T. M., M. R. van den Broeke, J. H. van Angelen, E. van Meijgaard, and S. J. Dery (2012), Drifting snow climate of the Greenland ice sheet: A study with a regional climate model, *Cryosphere*, 6, 891–899, doi:10.5194/tc-6-891-2012.
- McGrath, D., W. Colgan, N. Bayou, A. Muto, and K. Steffen (2013), Recent warming at Summit, Greenland: Global context and implications, *Geophys. Res. Lett.*, 40, 2091–2096, doi:10.1002/grl.50456.
- Mölg, T., N. J. Cullen, D. R. Hardy, G. Kaser, and L. Klok (2008), Mass balance of a slope glacier on Kilimanjaro and its sensitivity to climate, *Int. J. Climatol.*, 28, 881–892.
- Mölg, T., N. J. Cullen, D. R. Hardy, M. Winkler, and G. Kaser (2009), Quantifying climate change in the tropical midtroposphere over East Africa from glacier shrinkage on Kilimanjaro, *J. Clim.*, 22, 4162–4181.
- Mölg, T., F. Maussion, W. Yang, and D. Scherer (2012), The footprint of Asian monsoon dynamics in the mass and energy balance of a Tibetan glacier, *Cryosphere*, 6, 1445–1461.
- Mölg, T., F. Maussion, and D. Scherer (2014), Mid-latitude westerlies as a driver of glacier variability in monsoonal High Asia, *Nat. Clim. Change*, 4, 68–73, doi:10.1038/nclimate2055.
- Moon, T., and I. Joughin (2008), Changes in ice front position on Greenland's outlet glaciers from 1992 to 2007, *J. Geophys. Res.*, 113, F02022, doi:10.1029/2007JF000927.
- Moon, T., I. Joughin, B. Smith, and I. Howat (2012), 21st-century evolution of Greenland outlet glacier velocities, *Science*, 336, 576, doi:10.1126/science.1219985.
- Nghiem, S. V., D. K. Hall, T. L. Mote, M. Tedesco, M. R. Albert, K. Keegan, C. A. Shuman, N. E. DiGirolamo, and G. Neumann (2012), The extreme melt across the Greenland ice sheet in 2012, *Geophys. Res. Lett.*, 39, L20502, doi:10.1029/2012GL053611.
- Ohmura, A., and N. Reeh (1991), New precipitation and accumulation maps of Greenland, *J. Glaciol.*, 37, 140–148.
- Ohmura, A., P. Calanca, M. Wild, and M. Anklin (1999), Precipitation, accumulation and mass balance of the Greenland ice sheet, *Z. Gletscherkd. Glazialgeol.*, 35(1), 1–20.
- Rae, J. G. L., et al. (2012), Greenland ice sheet surface mass balance: Evaluating simulations and making projections with regional climate models, *Cryosphere*, 6, 1275–1294, doi:10.5194/tc-6-1275-2012.
- Shepherd, A., et al. (2012), A reconciled estimate of ice-sheet mass balance, *Science*, 338(6111), 1183–1189, doi:10.1126/science.1228102.
- Shuman, C. A., K. Steffen, J. E. Box, and C. R. Stearns (2001), A dozen years of temperature observations at the Summit: Central Greenland automatic weather stations 1987–1999, *J. Appl. Meteorol.*, 40, 741–752, doi:10.1175/1520-0450(2001)040<0741:ADYOTO>2.0.CO;2.
- Smeets, C. J. P. P., and M. R. van den Broeke (2008a), Temporal and spatial variation of momentum roughness length in the ablation zone of the Greenland ice sheet, *Boundary Layer Meteorol.*, 128, 315–338.

- Smeets, C. J. P. P., and M. R. van den Broeke (2008b), Parameterizing scalar roughness over smooth and rough ice surfaces, *Boundary Layer Meteorol.*, **128**, 339–355.
- Steffen, K., and J. E. Box (2001), Surface climatology of the Greenland ice sheet: Greenland Climate Network 1995 – 1999, *J. Geophys. Res.*, **106**, 33,951–33,964, doi:10.1029/2001JD900161.
- Sundal, A. V., A. Shepherd, P. Nienow, E. Hanna, S. Palmer, and P. Huybrechts (2011), Melt-induced speed-up of Greenland ice sheet offset by efficient subglacial drainage, *Nature*, **469**, 521–524, doi:10.1038/nature09740.
- Tedesco, M., X. Fettweis, T. Mote, J. Wahr, P. Alexander, J. E. Box, and B. Wouters (2013), Evidence and analysis of 2012 Greenland records from spaceborne observations, a regional climate model and reanalysis data, *Cryosphere*, **7**, 615–630, doi:10.5194/tc-7-615-2013.
- Thompson, S., and D. Pollard (1997), Greenland and Antarctic mass balances for present and doubled atmospheric CO₂ from the GENESIS version-2 global climate model, *J. Clim.*, **10**, 871–900.
- van As, D., M. R. van den Broeke, C. Reijmer, and R. van de Wal (2005), The summer surface energy balance of the high Antarctic plateau, *Boundary Layer Meteorol.*, **115**, 289–317.
- van As, D., A. L. Hubbard, B. Hasholt, A. B. Mikkelsen, M. R. van den Broeke, and R. S. Fausto (2012), Large surface meltwater discharge from the Kangerlussuaq sector of the Greenland ice sheet during the record-warm year 2010 explained by detailed energy balance observations, *Cryosphere*, **6**, 199–209, doi:10.5194/tc-6-199-2012.
- van As, D., R. Fausto, W. Colgan, J. E. Box, and the PROMICE team (2013), Darkening of the Greenland ice sheet due to the melt-albedo feedback observed at PROMICE weather stations, *Geol. Surv. Den. Greenl. Bull.*, **28**, 69–72.
- van de Wal, R. S. W., W. Boot, M. R. van den Broeke, C. J. P. P. Smeets, C. H. Reijmer, J. J. A. Donker, and J. Oerlemans (2008), Large and rapid melt-induced velocity changes in the ablation zone of the Greenland Ice Sheet, *Science*, **321**, 111–113, doi:10.1126/science.1158540.
- van de Wal, R. S. W., W. Boot, C. J. P. P. Smeets, H. Snellen, M. R. van den Broeke, and J. Oerlemans (2012), Twenty-one years of mass balance observations along the K-transect, West Greenland, *Earth Syst. Sci. Data*, **4**, 31–35, doi:10.5194/essd-4-31-2012.
- van den Broeke, M. R., C. Reijmer, and R. van de Wal (2004a), A study of the surface mass balance in Dronning Maud Land, Antarctica, using automatic weather stations, *J. Glaciol.*, **50**, 565–582.
- van den Broeke, M. R., D. van As, C. Reijmer, and R. van de Wal (2004b), Assessing and improving the quality of unattended radiation observations in Antarctica, *J. Atmos. Oceanic Technol.*, **21**(9), 1417–1431.
- van den Broeke, M. R., C. Reijmer, D. van As, and W. Boot (2006), Daily cycle of the energy balance in Antarctica and the influence of clouds, *Int. J. Climatol.*, **26**, 1587–1605.
- van den Broeke, M. R., P. Smeets, and J. Ettema (2008a), Surface layer climate and turbulent exchange in the ablation zone of the west Greenland ice sheet, *Int. J. Climatol.*, **29**, 2309–2323, doi:10.1002/joc.1815.
- van den Broeke, M. R., P. Smeets, J. Ettema, and P. K. Munneke (2008b), Surface radiation balance in the ablation zone of the west Greenland ice sheet, *J. Geophys. Res.*, **113**, D13105, doi:10.1029/2007JD009283.
- van den Broeke, M. R., P. Smeets, J. Ettema, C. van der Veen, R. van de Wal, and J. Oerlemans (2008c), Partitioning of energy and meltwater fluxes in the ablation zone of the west Greenland ice sheet, *Cryosphere*, **2**, 179–189, doi:10.5194/tc-2-179-2008.
- van den Broeke, M. R., J. Bamber, J. Ettema, E. Rignot, E. Schrama, W. Jan van de Berg, E. van Meijgaard, I. Velicogna, and B. Wouters (2009), Partitioning recent Greenland mass loss, *Science*, **326**, 984–986.
- van den Broeke, M. R., C. J. P. P. Smeets, and R. S. W. van de Wal (2011), The seasonal cycle and interannual variability of surface energy balance and melt in the ablation zone of the west Greenland ice sheet, *Cryosphere*, **5**, 377–390, doi:10.5194/tc-5-377-2011.
- van Pelt, W. J. J., J. Oerlemans, C. H. Reijmer, V. A. Pohjola, R. Pettersson, and J. H. van Angelen (2012), Simulating melt, runoff and refreezing on Nordenskiöldbreen, Svalbard, using a coupled snow and energy balance model, *Cryosphere*, **6**, 641–659, doi:10.5194/tc-6-641-2012.
- Vernon, C. L., J. L. Bamber, J. E. Box, M. R. van den Broeke, X. Fettweis, E. Hanna, and P. Huybrechts (2012), Surface mass balance model intercomparison for the Greenland ice sheet, *Cryosphere*, **7**, 599–614, doi:10.5194/tc-7-599-2013.