Temperature thresholds for degree-day modelling of Greenland ice sheet melt rates

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[1] Degree-day factors (DDFs) are calculated for the ice sheet ablation zone in southwest Greenland, using measurements of automatic weather stations and a regional atmospheric climate model. The rapid increase of DDFs for snow and ice towards higher elevations is caused by the increasing dominance of short daytime melting and nocturnal refreezing. This spatial inhomogeneity can be avoided by choosing a lower threshold for daily average 2 m air temperature (268 K instead of 273.15 K) for the degree-day calculation. **Citation:** van den Broeke, M., C. Bus, J. Ettema, and P. Smeets (2010), Temperature thresholds for degree-day modelling of Greenland ice sheet melt rates, *Geophys. Res. Lett.*, *37*, L18501, doi:10.1029/2010GL044123.

1. Introduction

[2] The melt rate of a snow/ice surface is determined by the surface energy balance (SEB), which can be written as:

$$M = SW_{in} + SW_{out} + LW_{in} + LW_{out} + SHF + LHF + G_s \quad (1)$$

where M is the energy available for melting (M = 0 if surface temperature T_s < 273.15 K), SW_{in} and SW_{out} are downward and reflected shortwave radiation fluxes, LW_{in} and LW_{out} are downward and emitted longwave radiation fluxes, SHF and LHF are the turbulent fluxes of sensible and latent heat and G_s is the sub-surface conductive heat flux, evaluated at the surface. When fluxes are directed towards the surface, they are defined as positive. Equation (1) represents the SEB of a 'skin' layer without heat capacity, neglecting subsurface penetration of SW radiation.

[3] Explicitly calculating M by closing the SEB requires information on many parameters, such as radiation components, temperature, humidity, wind speed, surface roughness, snow/ice density and snow temperature. The difficult operation of meteorological equipment on a melting ice surface makes that year-round SEB studies from the Greenland ablation zone are relatively rare [*van den Broeke et al.*, 2008]. That is why the temperature-index or positive degree-day method is an often-used alternative to calculate melt rate for extratropical glaciers.

[4] The most basic formulation of the degree-day method is [*Braithwaite*, 1985; *van de Wal*, 1996; *Hock*, 2003]:

$$T_{2m} > T_0 : \Sigma M_{i,s} = DDF_{i,s}\Sigma(T_{2m} - T_0)\Delta \mathbf{t} \quad (2)$$

where the amount of ice/snow melt $M_{i,s}$ (kg m^{-2}) per time interval Δt is related to the average 2 m air temperature T_{2m}, when T_{2m} exceeds a certain threshold value T_0, often taken to be the melting point of ice/snow, 273.15 K. Melt and T_{2m} are coupled via the degree day factor for ice/snow $DDF_{i,s}$, which is assumed constant in time. Usually, Δt is chosen as one day, so that $DDF_{i,s}$ is expressed in kg m^{-2} day^{-1} K^{-1}. The values of DDF_i and DDF_s can be determined experimentally by simultaneous measurements of T_{2m} and melt, for instance using a sonic height ranger. Because daily melt amounts are hard to measure accurately, DDF_i and DDF_s are mostly determined from melt and positive degree-days that have been accumulated over longer periods, e.g. one or several ablation seasons.

[5] The physical basis for the degree-day model is that net longwave radiation, sensible heat flux and to a certain extent also the latent flux over a melting ice surface are correlated with T_{2m} [*Ohmura*, 2001]. Using parameterizations of T_{2m} based on data from weather stations near and on the ice sheet [*Ohmura*, 1987; *Reeh*, 1991; *Fausto et al.*, 2009], the degree-day method has been used to calculate melt in Greenland [*Braithwaite and Olesen*, 1985; *Braithwaite*, 1995]. Also, because the method is computationally cheap, it has been used to calculate melt rate in dynamical ice sheet models, often in combination with a meltwater retention/refreezing model to translate melt to runoff [*Huybrechts and de Wolde*, 1999; *Janssens and Huybrechts*, 2000].

[6] The obvious disadvantage of the degree-day method is that it cannot be universally applied, because the DDF depends on the atmospheric structure through LW_{in}, on surface roughness through SHF, the different partitioning of the energy fluxes in the SEB and the variability in surface albedo. DDF_i and DDF_s must then be determined experimentally for each location, and as a result, a rather wide range of values for DDF_i and DDF_s has been reported in the literature [*Hock*, 2003]. But there is also a sampling issue: on days with a negative average T_{2m}, the method predicts zero melt if T_0 = 273.15 K is used, while melt may have occurred during a short period. This problem may be avoided by applying the method to hourly T_{2m} values or by applying a lower value for T_0 (see below). Another important issue is whether it is reasonable to assume that DDF_i and DDF_s remain constant in a future changing climate.

[7] In this paper we use observations of automatic weather stations (AWS) in the ablation zone of the Greenland ice sheet (GrIS) near Kangerlussuaq, and output of a regional atmospheric climate model, to assess the spatial and temporal variability of DDF_i and DDF_s in the region with the strongest ablation rates in Greenland [*Ettema et al.*, 2009].

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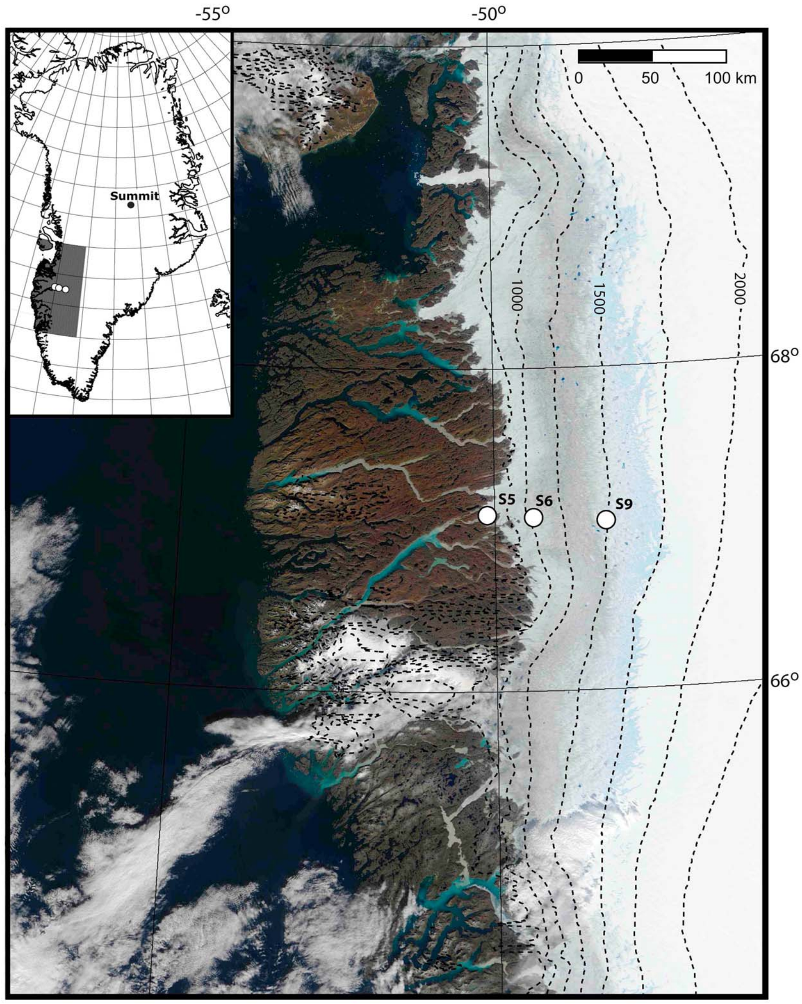

Figure 1. MODIS scene of west Greenland (August 23, 2006) with AWS locations (white dots) and ice sheet elevation contours (dashed lines, height interval 250 m, from [*Bamber et al.*, 2001]).

We demonstrate that by using a lower value for T_0, the applicability of the method to the GrIS is improved.

2. Methods

[8] *Van den Broeke et al.* [2008] used four years (August 2003–August 2007) of continuous hourly AWS data to close the local SEB in the GrIS ablation zone. The three AWS are situated in the lower (S5), middle (S6) and higher (S9) ablation zone in southwest Greenland (Figure 1), S5 near the ice margin and S9 near the equilibrium line. The AWS sites cover the wide range of surface and climate conditions usually encountered in the GrIS ablation zone (Table 1). A combination of sonic altimeter data and albedo measurements was used to decide whether snow or ice was at the surface at the time of melting. Melt and sublimation were calculated using an energy balance model applied to the AWS data, evaluated with sonic height ranger data [*van den Broeke et al.*, 2008]. Next, daily averages of T_{2m} and cumulative melt were determined, and from those cumulative running values of DDF_i and DDF_s.

[9] The regional atmospheric climate model RACMO2/GR was applied to the GrIS and its surroundings at 11 km resolution for the period 1958–2008, using ERA40 and ECMWF operational analysis as lateral forcings. The model gives a faithful representation of GrIS T_{2m} and surface mass balance [*Ettema et al.*, 2009; *van den Broeke et al.*, 2009; *Ettema et al.*, 2010a, 2010b]; here we use cumulative (2003–2007) degree-days and melt directly from RACMO2/

Table 1. AWS Characteristics and Annual Average Climate Parameters

	S5	S6	S9
Location (August 2006)			
Latitude (N)	67° 06′	67° 05′	67° 03′
Longitude (W)	50° 07′	49° 23′	48° 14′
Elevation (m asl)	490	1020	1520
Distance from ice edge (km)	6	38	88
Period of Operation			
Start of observation	28 Aug 2003	1 Sep 2003	1 Sep 2003
End of observation	27 Aug 2007	31 Aug 2007	31 Aug 2007
Annual Mean Climate Variables			
Mass balance (m w.e.)	−3.6	−1.5	~0
Pressure (hPa)	950	887	835
2 m air temperature (K)	267.7	263.4	260.6
2 m relative humidity (%)	75	87	90
2 m specific humidity (g kg^{-1})	2.4	2.2	1.9
10 m wind speed (m s^{-1})	5.0	6.4	7.3

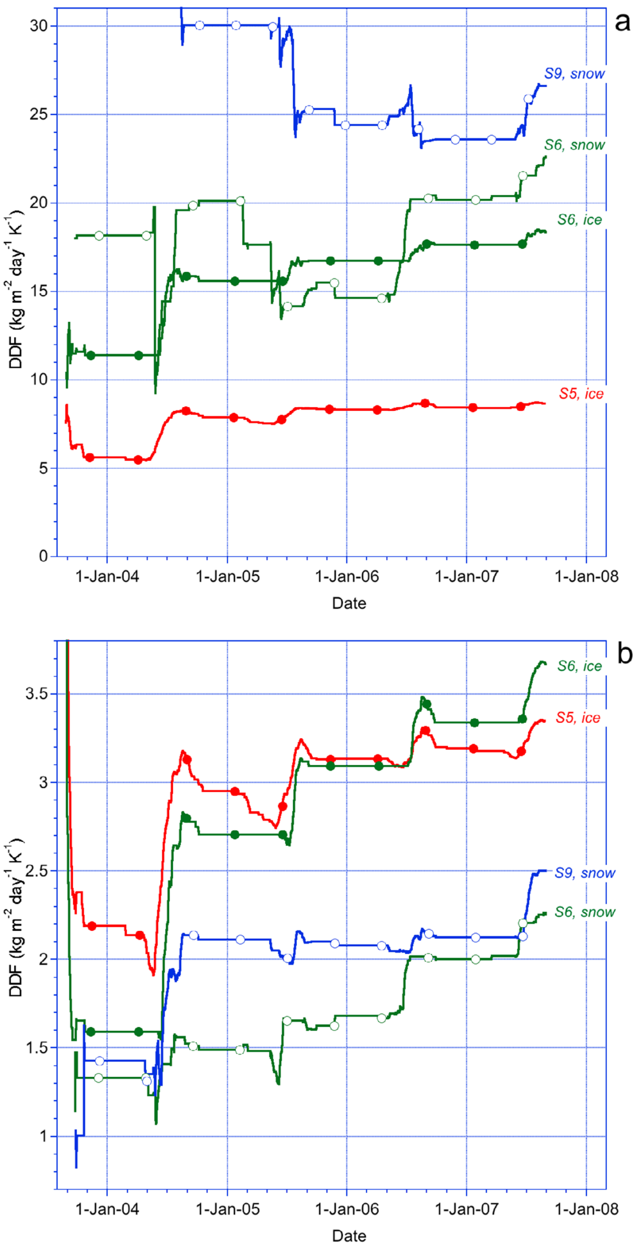

Figure 2. Cumulative Degree Day Factors for snow (DDF_s, open symbols) and ice (DDF_i, filled symbols) using the two threshold values of (a) T_0 = 273.15 K and (b) T_0 = 268 K. $DDF_{i,s}$ are calculated using melt and positive degree-days cumulated up to that moment, which explains the decreasing variability in time. Note that no change in DDF occurs during winter, when the cumulative quantities do not change.

GR to calculate DDF_i and DDF_s in the region surrounding the AWS.

3. Results: AWS

[10] In the period under consideration (2003–2007), the ice (snow) surface melted for 30% (1%), 20% (6%) and 12% (12%) of the time at S5, S6 and S9, respectively. Almost no snow melted at S5 and almost no ice melted at S9. As a result, S5 and S9 only provide values for DDF_i and DDF_s, respectively. The reason for the near-absence of snowmelt at S5 is that winter snow is blown into crevasses and gullies or sublimated during snowdrift events [*van den Broeke et al.*, 2008]. S9 is situated near the equilibrium line, where ice is present at the surface for brief periods only, yielding too little data to calculate DDF_i. Moreover, most of the ice that appears at the surface at S9 presumably is superimposed ice, which is not further considered here because it has radiation characteristics different from glacier ice.

[11] Figure 2a shows DDF_i and DDF_s, based on cumulative degree-days and melt, using T_0 = 273.15 K. At S5, DDF_i is rather constant with a value of ≈8 kg m^{-2} day^{-1} K^{-1} from 2004 onwards, in good agreement with the value of 9.2 kg m^{-2} day^{-1} K^{-1} reported by *van de Wal* [1992] for the GIMEX 1990/91 campaigns at the same site, and other values for the Greenland ablation zone below 1000 m asl as listed by *Hock* [2003]. At S6 the value of DDF_i increases in time from ≈ 11–12 kg m^{-2} day^{-1} K^{-1} in 2004 to ≈ 18 kg m^{-2} day^{-1} K^{-1} in 2007. The latter is close to the value of 20.0 kg m^{-2} day^{-1} K^{-1} reported for the same site in 1990/91 by *van de Wal* [1992] and the value of 18.6 kg m^{-2} day^{-1} K^{-1} reported by *Ambach* [1988] for CAMP-IV (1013 m asl) for the 1959 melt season. The values of DDF_s at S6 and S9 are higher still, considerably higher than those listed in the literature [*Hock*, 2003].

[12] Assuming the same T_{2m} over a snow and an ice surface, one would expect snow melt rate to be generally smaller than that of ice because of the higher surface reflectivity of snow compared to glacier ice [*Kuipers Munneke et al.*, 2010], so that $DDF_i > DDF_s$. Surprisingly, however, $DDF_s > DDF_i$ in Figure 2a. This unexpected behaviour is caused by the temporal sampling problem mentioned before: at the higher sites S6 and S9, melting mainly occurs at daytime, while the surface refreezes during the night under the influence of negative net radiation. This results in numerous melt days with average T_{2m} lower than or only marginally greater than 273.15 K, i.e. the cumulative positive degree-days increase relatively slowly compared to the cumulative melt, resulting in a large value of DDF.

[13] The sampling problem is visualized by plotting the cumulative distribution of T_{2m} for all melt days at the AWS sites (Figure 3). At S5, 90% of the melt days has T_{2m} > 273.15 K. At S6, this is true for only 70% and at S9 for only 40% of the melt days. This means that around the equilibrium line at S9, more than half of the melt days is not counted if T_0 is set at 273.15 K. Higher on the ice sheet, where melt is even less frequent, this percentage could theoretically reach 100%. If, alternatively, we use T_0 = 268 K (lower dashed line in Figure 3), the percentage of melt days included increases to 100% (S5), 99.5% (S6) and 96% (S9). So, by adjusting T_0 downward by 5 K, nearly all melt days are captured and the sampling problem should be strongly reduced. This is confirmed by recalculating DDF_i and DDF_s using T_0 = 268K (Figure 2b). The values of DDF for snow are now lower than those of ice, as expected, and are also more consistent among the stations. Note that the new values of DDF_i and DDF_s cannot be compared to values from literature that use T_0 = 273.15 K.

4. Results: Regional Climate Model

[14] DDF_i and DDF_s were also calculated for 2003–2007 using daily average T_{2m} and cumulative melt from the

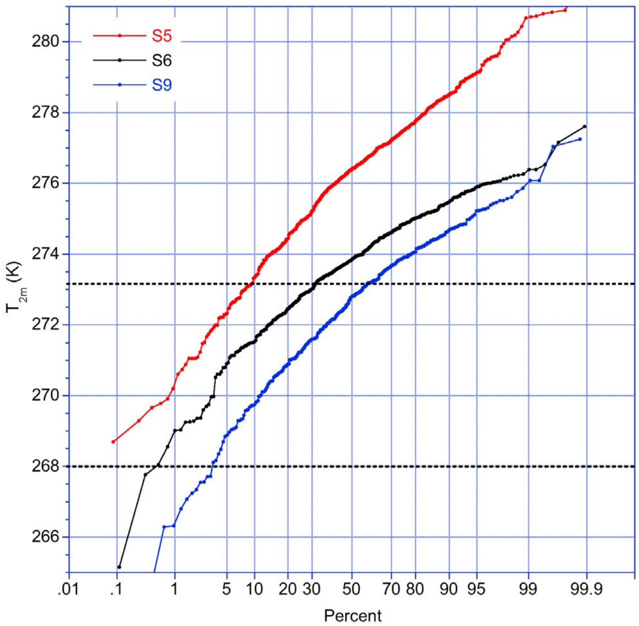

Figure 3. Cumulative distribution of daily average T_{2m} for days with melt at the three AWS (August 2003 – August 2007). The horizontal dashed lines represent the two threshold values for T_0 used in this study.

regional atmospheric climate model RACMO2/GR. Although the model was run over the entire ice sheet, here we limit the analysis to the region of interest in southwest Greenland, in order to reduce the effects of latitudinal insolation gradients. Figure 4 shows DDF_i (upper frames a and c) and DDF_s (lower frames b and d) for T_0 = 273.15 K (left) and T_0 = 268 K (right). Circles indicate DDF values determined from the AWS data, using the same colour code.

[15] Although the absolute values differ somewhat, due to an imperfect simulation of T_{2m} and melt in RACMO2/GR, the model does reproduce the rapid increase of DDF_i and DDF_s with elevation for T_0 = 273.15 K (Figures 4a and 4b). Over highly elevated snow-covered regions with infrequent daytime melt, modelled DDF_s attain values in excess of 300 kg m^{-2} day^{-1} K^{-1} (Figure 4b). For T_0 = 268 K (right), the model does reproduce the AWS-derived values of DDF_i and DDF_s quite well. More importantly, the spatial distribution of DDF_i and DDF_s is now more homogeneous. Using a lower value for T_0 thus eliminates most of the sampling problem, so that a constant value of DDF_i and DDF_s can be reasonably applied to all ice and snow gridpoints, even those presently situated in the dry snow zone (Figure 4d). This is especially favourable for use of the degree-day scheme in climate change scenarios, when the melt zone will shift upwards.

5. Discussion and Conclusions

[16] Using a ~5 K lower threshold value of T_0 removes most of the sampling problem in the application of the positive degree-day model to the ice sheet in southwest Greenland. The resulting values of the degree-day factors for snow and ice are now physically consistent (DDF_i > DDF_s) and spatially more homogeneous.

[17] An outstanding problem is the non-stationarity of DDF_i and DDF_s in Figure 2b, which show an upward trend during 2003–2007. Long-term observations from Alpine glaciers suggest that DDF_s may change significantly in time [*Huss and Bauder*, 2009]. The reason for this is that melt variability from season to season is mainly determined by variations in SW_{net}, through melt season length and the melt-albedo feedback, and less by SEB terms that are (partly) determined by T_{2m}, such as LW_{net}, SHF and LHF (M. van den Broeke, Daily, seasonal and interannual variability of surface energy balance in the ablation zone of the west Greenland ice sheet, manuscript in preparation, 2010). These and other processes that influence albedo in a transient climate [*Oerlemans et al.*, 2009] are not incorporated in the degree-day method, which represents an inherent weakness [*Bougamont et al.*, 2007]. For climate change

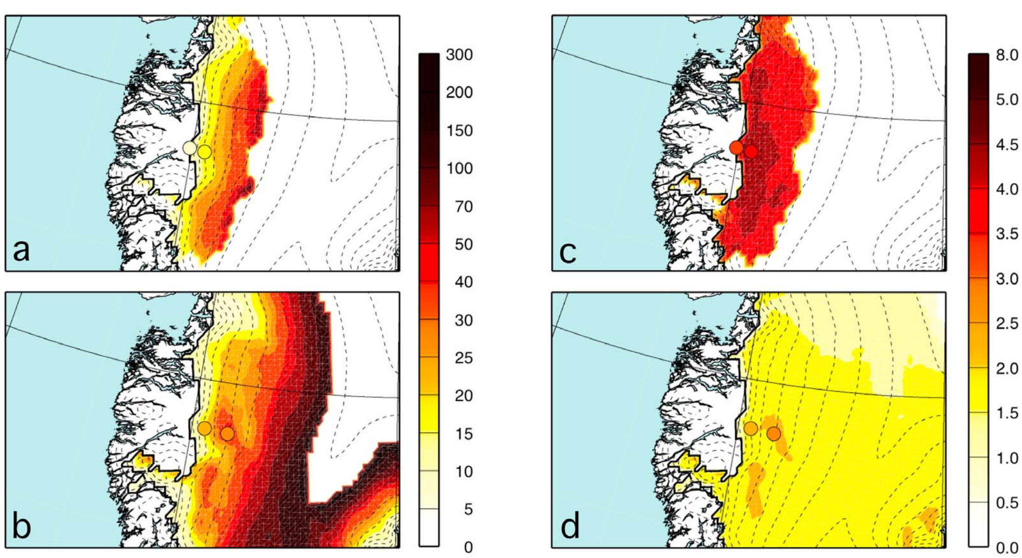

Figure 4. Spatial distribution of DDF_i (Figures 4a and 4c) and (Figures 4b and 4d) DDF_s in kg m^{-2} day^{-1} K^{-1} for (a and b) T_0=273.15 K and (c and d) T_0=268 K from the regional atmospheric climate model RACMO2/GR in the surroundings of the AWS. Circles represent AWS values, using the same colour scale.

scenarios of the GrIS, quantifying the temporal variability of DDF is thus important, but this requires longer time series of temperature and melt than are available at present.

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