

Black carbon and dust accumulation

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Albedo reduction caused by black carbon and dust accumulation: a quantitative model applied to the western margin of the Greenland ice sheet

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Abstract

Ice loss due to surface melt of the Greenland ice sheet has increased in recent years. Surface melt in the ablation zone is controlled by atmospheric temperature and surface albedo. Impurities such as mineral dust and black carbon darken the snow and ice surfaces and therefore reduce the surface albedo which leads to more absorbed solar energy and ultimately amplifying melt. These impurities accumulate on the ice surface both from atmospheric fallout and by melt-out of material which was enclosed in the snowpack or the ice compound. A general impurity accumulation model is developed and applied to calculate the surface albedo evolution at two locations in western Greenland. The model is forced either by regional climate model output or by a parameterisation for temperature and precipitation. Simulations identify mineral dust as the main contributor to impurity mass on ice where the dominating part originates from melt out of englacial dust. Daily reduction of impurities is in the range of one per-mille which leads to a residence time of decades on the ice surface. Therefore the impurities have a prolonged effect on surface melt once they are located on the ice surface. The currently englacially stored mineral dust and black carbon will effect future melt and sea level rise and can be studied with the presented model.

1 Introduction

Sea level rise due to melting and calving of ice sheets and glaciers is a major concern in association with climate change and has attracted considerable attention. Ice loss of the Greenland Ice Sheet (GrIS) has increased in recent years (Vaughan et al., 2013) where until 2009 the ice loss due to surface melt accounted for 50 % (Van den Broeke et al., 2009a) and up to 84 % after (Enderlin et al., 2014).

Absorption of shortwave radiation is the largest energy source for ice and snow melt (i.e. Male and Granger, 1981; Cuffey and Paterson, 2010). Surface albedo and incident radiation are controlling the amount of absorbed shortwave radiation, hence surface

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albedo has been identified as a major component of the surface mass balance for ice sheets (e.g. van Angelen et al., 2012; Bougamont et al., 2005; Tedesco et al., 2011). Despite the importance of albedo it is rarely included in ice sheet model and if it is then treated as constant in time and space (e.g. Konzelmann and Braithwaite, 1995; Hock and Noetzli, 1997; Robinson et al., 2011, 2010; Bougamont et al., 2005). Sensitivity experiments of the GrIS (Machguth et al., 2013) suggest that the present high mass loss could be 20–30 % higher if the surface albedo would be significantly lower in the future. Currently a downward trend in the surface albedo is observed at the GrIS margin with a decrease by up to 18 % in coastal regions (Box et al., 2012).

Impurities such as black carbon (BC), mineral dust (Warren and Wiscombe, 1980) and microbes affect the albedo of snow and ice. Microbes may influence the albedo by aggregating material (Takeuchi et al., 2001) and by producing dark materials (Takeuchi, 2002; Remias et al., 2012). Yet the influence of microbes on surface albedo remains to be quantified (Stibal et al., 2012; Yallop et al., 2012) as its dynamics are currently not well understood. Other impurities are aerosols, which in the context of glaciers, are eolian transported particles which are located on, or inside the snowpack and glacier ice. Only at the very margin surfacing till from the base of the glacier become significant and can locally affect the albedo. Albedo relevant aerosols are BC originating from natural or man-made forest fires or incomplete fuel composition as well as mineral dust produced by weathering of the earth surface. Mineral dust (subsequently referred to as just dust) is the main contributor to total impurity mass (e.g. Takeuchi et al., 2014) but BC is much more effective in reducing albedo (Warren and Wiscombe, 1980). Just a few nano-gram of BC have a significant impact on ice and snow albedo. Termination of the little ice age has been reported to be possibly caused by industrial BC (Painter et al., 2013).

Large ice masses act as a reservoir of dust and BC by entraining the particles in the accumulation zone which are later released at the margin after centuries or millennia of transport. These englacial impurities accumulate (e.g. Oerlemans, 1991; Bøggild et al., 1996; Klok and Oerlemans, 2002; Oerlemans et al., 2009) at the ice surface as

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they melt-out and therefore have a long-time effect on surface melt. Between 1850 to 1951 North America is probably the main source of BC in central Greenland where current emissions are almost back to pre-industrial levels (McConnell et al., 2007). Nevertheless today the main source is probably Asia (McConnell et al., 2007; Koch and Hansen, 2005) also found to be the main source of dust in the past (Biscaye et al., 1997). Asian BC emissions were rising since 2000 and in particular in China by about 30 % (Lei et al., 2011) to 40 % (Lu et al., 2011).

Together with the high concentrations of dust from glacial times currently stored inside the GrIS this forms a potential major source of uncertainty to future sea level rise. Model projections of future melt of glaciers and ice sheets might be underestimated without accounting for the influence of impurities on ice albedo.

Sophisticated snow albedo models like SNICAR (Flanner and Zender, 2006) have existed for some time. While these models are computationally to demanding to include in large scale models they give important insight to efficient parameterisations (Pirazzini, 2009). Global circulation models use different snow albedo models with different levels of complexity and some may include simple parameterisation of impurities (Dolan et al., 2014). However, a physical based model of ice albedo which includes the accumulation of impurities is still missing.

Our goal in this paper is twofold. Firstly, development of a computationally efficient albedo model which takes outcropping and accumulation of impurities on the ice surface into account. Secondly, we examined the relative contribution of outcropping dust and BC to the total impurity mass and investigate its residence time on the surface at two locations on the western margin of Greenland based on local meteorological data.

2 Model description

2.1 Impurity accumulation

The surface albedo of an ice sheet (Fig. 1) is influenced by several processes. In principle there are three factors: optical properties of the surface (albedo), angle of incidence of downward radiation and proportions of direct to diffuse radiation (Cuffey and Paterson, 2010).

The surface property of an ice sheet is primarily effected by surface type (snow or ice) and secondarily by the optical surface properties. This is because the difference between snow and ice albedo is normally greater than between dry and wet snow. The optical properties of snow and ice are determined by impurities and the specific surface area of the material. The specific surface area of snow is determined by snow metamorphism and of ice by enclosed air, cracks and the magnitude of the ablation crust (Müller and Keeler, 1969). The magnitude of diffuse scatter below the surface and therefore the albedo is effected by this specific surface area.

Impurities such as BC, mineral dust and microbes have different origins and accumulating on the ice and snow surface. There are four different sources of impurities at the ice surface in the ablation zone (Fig. 2):

- *atmosphere*: k_I by dry or wet deposition,
- *tundra*: k_{II} by regionally transported material from the surrounding tundra,
- *flow*: k_{III} by transport from the accumulation zone to the ablation zone where impurities melt-out,
- *local*: k_{IV} by local biological production of dark material on the ice or snow surface.

For model purpose the accumulation has to be treated separately for each impurity species with different considerations for snow and ice. Since the microbiological interaction is not yet well understood we limit ourself to BC ($n = 1$) and dust ($n = 2$) only

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in this study. Nevertheless the equations can be applied for other species, including microbes. We treat the impurities l as a mass per area (ng m^{-2}). If snow melts then all the impurities are deposited on the ice surface and all the sources contribute directly to the amount on the ice surface. On the other hand if snow covers the ice surface all the sources contribute to the amount in the snowpack and the impurities amount on the ice surface remains constant. Hence the impurity amount located on top of the ice is preserved from one melt season to the next with the additional influx of material from the melted winter snowpack. The total amount is balanced by reduction ($r_{\text{ice},n}(t)$) from the ice surface due to meltwater removal and we assume no loss of impurities up into the snowpack. Moreover, resuspension of formerly deposited aerosols is not likely due to the fact that the surface is either wet or refrozen with the impurity entrained in the frozen ice. We also assume that once aerosol is in the snowpack it is not resuspended again.

Impurity accumulation inside the snowpack (Eq. 1) and on the ice surface (Eq. 2) are described in general by:

$$\frac{dl_{\text{snow},n}}{dt} = \begin{cases} k_{\text{I},n}(t) + k_{\text{II},n}(t) + k_{\text{IV},n}(t) & d > 0 \\ 0 & d = 0 \end{cases} \quad (1)$$

$$\frac{dl_{\text{ice},n}}{dt} = \begin{cases} 0 & d > 0 \\ k_{\text{I},n}(t) + k_{\text{II},n}(t) + k_{\text{III},n}(t) + \\ + k_{\text{IV},n}(t) - r_{\text{ice},n}(t) & d = 0 \end{cases} \quad (2)$$

The in situ production term $k_{\text{IV},n}(t)$ is only relevant for microbiological activity and is zero otherwise.

When ice is melting the impurities transported with the ice from the accumulation zone resurface as they melt out and therefore contribute to the impurity mass at the ice surface. This is described by the term $k_{\text{III},n}(t)$ in Eq. (2). We assume that superimposed ice ($h_{\text{s,ice}}$) does not contain any impurities since impurities will not be transported downward with melt water during snow melt, but instead accumulate on the

snow surface. Otherwise the source of impurities depends on the englacial impurity concentration nearest to the ice surface, the amount of melt and ice density:

$$k_{\text{Ill},n}(t) = \begin{cases} [l_{n,\text{englacial}}] \Delta h_{\text{ice}} \Delta t_b^{-1} \rho_{\text{ice}} & h_{s,\text{ice}} = 0 \\ 0 & h_{s,\text{ice}} > 0 \end{cases} \quad (3)$$

For a sufficient short time steps in Eq. (3) (Δt_b) ice melt Δh_{ice} is small enough that the englacial aerosol concentration ($[l_{n,\text{englacial}}]$) can be treated as constant. In this study we prescribe the englacial $[l_{n,\text{englacial}}]$ concentration with values from shallow ice cores.

2.2 Surface mass balance

It is necessary to compute ice melt and snow thickness (d) for the accumulation equations and later also for the resulting surface albedo. The daily change of snow pack height d and ice thickness h_i for the melt-out of englacial impurities with Eqs. (4) and (5) (Robinson, 2011) reads:

$$\frac{d}{dt_b} d = P_{\text{solid}} - M_s, d \in (0, d_{\text{max}}) \quad (4)$$

$$\frac{d}{dt_b} h_i = \begin{cases} M_s r_f & d > 0 \\ \min(P_{\text{solid}} - M_s, 0) & d = 0 \end{cases} \quad (5)$$

where P_{solid} is the solid precipitation and M_s the potential melt rate which will be described in the next section in more detail.

If the snow depth exceeds $d_{\text{max}} = 5$ m (consistent with Robinson et al., 2010; Fitzgerald et al., 2012) the excess amount is added to the ice thickness and the snow thickness is reset to 5 m.

The amount of solid precipitation rate P_{solid} in Eq. (5) depends on a temperature dependant fraction $f(T)$ and the total amount of precipitation P :

$$P_{\text{solid}} = P \cdot f(T) \quad (6)$$

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The surface temperature depending fraction $f(T)$ is empirically based on data of Greenland (Bales et al., 2009; Calanca et al., 2000) and states that below a minimum temperature T_{\min} (-7°C) all precipitation is snow and above a maximum temperature ($+7^{\circ}\text{C}$) rain:

$$f(T) = \begin{cases} 1 & T < T_{\min} \\ 0 & T > T_{\max} \\ \cos\left(\frac{T-T_{\min}}{T_{\max}-T_{\min}}\right)\left(\frac{\pi}{2}\right) & T_{\min} < T < T_{\max} \end{cases} \quad (7)$$

Refreezing within a snowpack is a function of the snow depth and the surface temperature T :

$$rf = \begin{cases} 0 & d = 0 \\ r_{\max}f(T) & 0 < d \leq 1 \\ r_{\max} + [(1 - r_{\max})(d - 1)] & 1 < d \leq 2 \\ 1 & d > 2 \end{cases} \quad (8)$$

In the absence of snow the refreezing fraction is 0. Between 0 and 1 m w.e. the refreezing fraction is following the definition in Janssens and Huybrechts (2000) and also depending on $f(T)$. Where r_{\max} is the fraction of melt precipitation that has to that has to refreeze before runoff can occur (Fitzgerald et al., 2012).

2.3 Melt

We use a simplified energy-balance model based on Oerlemans (2001) and recently applied for Greenland (Robinson et al., 2010, 2011; Fitzgerald et al., 2012):

$$M_s = \frac{1}{\rho_w L_m} [\tau_a (1 - \alpha_s) S_{\text{TOA}} + c + \lambda T] \quad (9)$$

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Where ρ_w is the density of water, L_m is the latent heat of melting, τ_a is the transmissivity, S_{TOA} the insolation on top of the atmosphere, c and λ are empirical parameters for short-wave radiation and sensible heat flux and long-wave radiation respectively and T is the surface air temperature. Equation (9) requires only the information of the insolation on top of the atmosphere which can be calculated for any time by simple equations (Liou, 2002). The transmissivity τ_a is based on summer (April–September) radiation data over Greenland and is a linear fit with elevation h in meters (Robinson et al., 2010):

$$\tau_a = 0.46 + 0.00006 h \quad (10)$$

The only variable which remains to be described in Eq. (9) is the surface albedo.

2.4 Surface albedo

2.4.1 General albedo parameterisation

In the absence of snow the surface albedo in Eq. (9) is identical to the ice albedo (see also Fig. 1). From bare ice to a critical snow depth d_{crit} the surface albedo fades linearly to the snow albedo:

$$\alpha_s = \begin{cases} \alpha_{ice} & d = 0 \\ \alpha_{ice} + \frac{d}{d_{crit}} (\alpha_{snow} - \alpha_{ice}) & 0 < d < d_{crit} \\ \alpha_{snow} & d \geq d_{crit} \end{cases} \quad (11)$$

Even with a few centimetres of snow bare ice can be exposed due to meter scale surface undulations of the glacier ice surface and the effect of wind drift.

Equation (11) is a more generalised version of the one used in Robinson et al. (2010, 2011) which uses the parameterisation of Van den Berg et al. (2008). This parameterisation allows the snow albedo to be lower than the ice albedo.

The basic method for snow and ice albedo are the same where the albedo is composed of the base albedo and the sum of albedo reductions due to impurities and the influence of the solar zenith angle $d\alpha_{\theta_z}$:

$$\alpha_{\text{ice|snow}} = \alpha_{\text{base}} + \sum_1^n d\alpha_{l,n} + d\alpha_{\theta_z} \quad (12)$$

The influence of clouds on albedo is already included in the melt scheme and optimised for GrIS (Robinson et al., 2010) via Eq. (10). The albedo change due to sun zenith angle is also parameterised with the scheme of Gardner and Sharp (2010). At high zenith angles close to solstice at Kangerlussuaq the maximum albedo increase with high impurity loadings is below 0.04. Therefore we neglect the influence of solar angle for the moment as our main focus is on ice albedo.

2.4.2 Ice albedo

The base albedo of ice is computed via the specific surface area \hat{S} with Eq. (13) (Gardner and Sharp, 2010):

$$\alpha_{\text{base}} = 1.48 - \hat{S}^{-0.07} \quad (13)$$

The albedo reduction due to BC is calculated with the parameterisation in Eq. (14) (Gardner and Sharp, 2010). Dust is included by adding a BC equivalent mass to the BC amount, as dust is about 1/200 (Gardner and Sharp, 2010) less effective per mass:

$$d\alpha_{l,n} = \max \left(0.04 - \alpha_{\text{base}}, \frac{-[l_{\text{eff},n}]^{0.55}}{0.16 + 0.6\hat{S}^{0.5} + 1.8[l_{\text{eff},n}]^{0.6}\hat{S}^{-0.25}} \right) \quad (14)$$

Where $l_{\text{eff},BC}$ is the effective aerosol concentration in ppm described as:

$$[l_{\text{eff},n}] = [l_{n,\text{englacial}}] + [l_{\text{ice},n}] \cdot \mathcal{F}_{\text{ice},n} \quad (15)$$

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The *active fraction* $\mathcal{F}_{\text{ice},n}$ describes the fact that not all impurities are influencing the albedo. Measurements of cryoconite holes at the GrIS margin (Bøggild et al., 2010) has shown that a considerable mass of impurities, in the order of kilograms per square meter, is located inside the holes. The holes are vertical but the incident solar radiation is mostly non zenith reducing the impact of radiation on impurity melt in the bottom of the hole to nearly zero.

Since Eq. (14) requires a concentration in ppm and the accumulation model handles the total amount in ngm^{-2} a conversion is necessary:

$$[l_n] = \frac{l_n}{\rho d_{\text{eff}}} 10^6 \quad (16)$$

Where the impurity concentration is in ppm when the impurity amount is in ngm^{-2} and the effective depth d_{eff} is in meters and density is in kgm^{-3} . The effective depth and the active fraction on ice are not well constrained and both have the same effect on the results. They both describe how effective the total amount of aerosol is in reducing the albedo. This effective depth is related on the absorption length in ice which is depending on wavelength and impurities. We set the effective depth to 5 m which corresponds to an extinction coefficient of about 1 m^{-1} which is in the range of totally clear and white ice (Male and Gray, 1981) and leave the active fraction as a free parameter.

2.4.3 Snow albedo

In this study we focus manly on ice albedo and therefore left the snow albedo is defined by differentiating between wet and dry snow as in Robinson et al. (2010) although the model includes also snow albedo reduction due to BC and dust. This does not limit the main conclusions of this study as accumulation of BC and dust inside the snowpack is included. On a side note, our snowpack model has no layers but impurities such as BC (Doherty et al., 2013) are usually located close to the surface which can be accounted for with a high active fraction.

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2.5 Model setup

Figure 3 shows the flow chart of the model for one time-step. First the climate parameters are updated and the snow-depth is calculated with the melt amount from the previous time-step. Then the aerosol accumulation of BC and dust in snow and on the ice surface is calculated which are then passed to the snow and ice albedo modules. Where different methods of calculating the albedo from aerosols are implemented and in this study the parameterisation from Gardner and Sharp (2010) is used. Then the resulting surface type and albedo is derived and used as input to the potential melt scheme, which affect the refreezing amount and ultimately in the surface mass balance.

3 The model applied to the western margin of Greenland

3.1 K-transect

The model is tested and forced with data from the K-transect (also known as Søndre Strømfjord transect) on the western Greenland margin near Kangerlussuaq airport (Fig. 4) at a latitude of 67° N. The transect is about 140 km long and reach from an elevation of 460 up to 1850 m i.e. above the equilibrium line which is located around 1500 m (Van de Wal et al., 2005) and KAN_U and S10 are located in the accumulation zone. The Institute for Marine and Atmospheric Research in Utrecht (IMAU) and later the Geological Survey of Denmark and Greenland (GEUS) (Van As et al., 2011) each operate three automatic weather stations (AWS). Even before IMAU installed the first AWS (S5, S6, S9) for continuous measurements since 2003 (Van den Broeke et al., 2009b) surface mass balance studies have been done (Van de Wal et al., 2012) since the beginning of the 1990s. A dark region persists below the equilibrium line (Van de Wal, 1992 as cited in Wientjes and Oerlemans, 2010) in each year. At 67° N the dark region is about 30 km wide and starts at approximately 30 km away from the

margin (Wientjes and Oerlemans, 2010). Previously meltwater was believed to be the cause (Zuo and Oerlemans, 1996) but now is attributed to dust (Wientjes and Oerlemans, 2010) and carbonaceous particles (Wientjes et al., 2012).

Reference surface albedo is calculated from daily mean data from the GEUS stations and hourly data for IMAU stations, where only data with a solar angle above 15° are used. This was found to be in agreement with the data where as 10° was used (Van As et al., 2005) before.

3.2 Experimental design and parameters

In order to compute the impurity and albedo evolution at the K-Transect the model parameters needs to be adjusted specifically for this particular environment. Some standard physical parameters common for all locations are listed in Table 1.

The model is applied to stations KAN_M and S5 because they both have the full cycle with dry snow, wet snow and exposed ice over the annual cycle. Station KAN_M is situated in the dark zone at 1280 m a.s.l and S5 is close to the margin at 460 m a.s.l with more reflectant ice but higher surface ablation due to higher temperatures. Station S5 has the longer record and was chosen over KAN_B and KAN_L. Station S6 has only complete data for the period 2005–2007 and was therefore left out.

Adjustable free parameters are: dry snow albedo, wet snow albedo, critical snow depth, the active fraction of BC and dust on ice, the specific surface area of ice and reduction fraction on ice. Which makes in total six free parameters.

The best parameter combination was derived by a Monte-Carlo method by taking the combination with the lowest value of S after a spin-up:

$$S = \sum_{01.04}^{30.09} |\alpha_{\text{mod}} - \alpha_{\text{dat}}| \quad (17)$$

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Parameter ranges taken into account can be seen in Fig. 6c. The active fraction of dust and BC was further optimised manually to improve the fit between model and observation after the Monte-Carlo search.

We use 100 years for the spin-up in order to guarantee that the model reaches an equilibrium with all parameter combinations. This long spin-up time also allowed to start with no snow and without aerosols on snow and ice. In practice with realistic parameter combinations the equilibrium is reached much earlier and how long it takes is primarily controlled by the reduction fraction.

The englacial BC value from KAN_M is taken from an ice core (Wientjes et al., 2012) at location S8 (7.25 km south-west) and is 4 ng g^{-1} . For S5 the value is set to 1 ng g^{-1} based on the closely located S1 core.

For dust the englacial value at KAN_M is set to 2000 and 100 ng g^{-1} at S5 as it is located in a cleaner region. These values are based on assumptions from the NGRIP core (Ruth, 2007) and the effect of these assumptions is discussed later (Sect. 5.2).

The atmospheric input rate (kII) of BC at both stations is $20 \text{ ng m}^{-2} \text{ a}^{-1}$ (Bauer et al., 2013) which also corresponds to the maximum value found in a snow sample close to Dye-2 (Doherty et al., 2013). For dust a value of $1 \text{ g m}^{-2} \text{ a}^{-1}$ taken everywhere from the mean value of Mahowald et al. (1999). The melt-out rate of aerosols is calculated (kIII) and all the other contributors from tundra and local production are assumed to be zero. This might lead for an underestimation of the dust as a significant amount could be transported from the surrounding tundra to the ice. Nevertheless the tundra input might be neglect-able as the dark region is 30 km away from the margin which indicates the insignificance of direct input from the tundra (Wientjes and Oerlemans, 2010).

The experiments are carried out once with temperature from AWS data and daily precipitation from MAR (Tedesco et al., 2014) model output and once with the parameterisation of those, as presented in the next section.

3.3 Temperature and precipitation parameterisation

As the model should be applicable for studies with long timescale a parameterisation of temperature and precipitation is required. The temperature in the K-transect has a high variability in the cold months (Figs. 5 and 7a) and is stably during the warmer summer season due to the fact that ice is cooling air temperatures during melt. Therefore we choose to parameterise the near surface air temperature not with a sinusoidal shape otherwise used (e.g. Reeh, 1991) but via a trapezoid shape:

$$T = \begin{cases} T_+ & t_{O,start} \leq t \leq t_{O,end} \\ t\zeta - t_{O,start}\zeta + T_+ & t < t_{O,start} \\ -t\zeta + t_{O,end}\zeta + T_+ & t > t_{O,end} \end{cases} \quad (18)$$

Where T_+ is the mean of all temperatures above 0°C and $t_{O,start}t_{O,end}$ are the days when summer starts and ends. At the K-transect we assumed a start of summer of first of May and ends at first of September. The non-summer temperatures are defined by the slope ζ . This is a parameterisation of temperature with only two parameters, namely the slope and the mean positive temperature T_+ .

The precipitation rate is parameterised with a sinusoidal shape in Eq. (19) where ΔP_{jul} is the precipitation rate offset in July \bar{P} is the mean precipitation rate and \mathcal{A} are seconds per year:

$$P = \bar{P} + \Delta P_{jul} \sin\left(\frac{2t\pi}{\mathcal{A}} + 1.5\pi\right) \quad (19)$$

4 Modeling results

4.1 KAN_M in 2010

Figure 5 displays detailed results of simulations of the KAN_M station in the dark zone of the K-transect in the year 2010. Details of the settings can be seen in Fig. 6c (green

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colors). Figure 5 shows both simulations which use weather station temperature and MAR- model precipitation in green and with the temperature and precipitation parameterisation in dotted lines. The temperature parameterisation in panel a is based on all the available data from 2009 until end 2012 by fitting the temperature over the entire period with a mean of 1.39°C and a SD of 0.04. This emphasises the low temperature variability at summer and the good fit of the parameterisation at the melt period. Panel b shows the daily precipitation of the MAR model and the parameterisation. Here we use the mean precipitation over the whole measurement period which results in a similar snow depth time-series (panel e) as with the MAR precipitation. Plots c and d show the BC and dust amount in the snowpack and on the ice surface respectively. Note the different scales for the different species and for snow and ice, respectively. The amount of aerosols in the snowpack is building up over time and is released onto the ice surface as the snow is melting, which coincides with superimposed ice formation. As we assumed no aerosol content in superimposed ice the aerosol content drops as a result of missing influx from ice melt and runoff on the ice surface. Superimposed ice exposure can be seen in panel f. Later in the melt season the aerosol amount on the ice surface rises again as englacial aerosols are released and accumulating until it balances with runoff and reaches its equilibrium level again, which is then preserved under the new snow cover until the next melt season.

The simulated maximum amount of BC on the surface is 0.051 gm^{-2} and of dust is 25.82 gm^{-2} . Hence the amount of dust is 502 times larger than black carbon.

The time series of the albedo is shown in panel g, for comparison the MAR computed albedo is plotted as a dashed line. In 2010 the simulated ice albedo reduction due to accumulated impurities was 0.11 from the base ice albedo.

4.2 Simulations of KAN_M from 2009 until 2013

Figure 6 presents an overview of different parameter values and model results compared to AWS data. Panel a shows the models with temperature from AWS data and MAR precipitation for each day with four different parameter configurations in panel c.

Each one is optimised for a different year. Panel b uses the same parameter configurations but this time the temperature and precipitation are parameterized with mean temperatures and precipitations calculated over the whole period.

In the simulations the optimisation of the year 2009 was problematic as the autumn snow cover was building up earlier than normal. Also the parameters set for this year failed to reproduce the low albedo in 2010 but succeeded partly in 2011 and 2012.

4.3 Stations S5 in 2011

Figure 7 displays the simulation of station S5 in 2011. Station S5 is located close to the margin at low elevation and therefore experiences more surface melt than KAN_M because of higher temperatures there. The parameter setting is shown in Fig. 6c and for the temperature parameterisation the data from 2005 until the end of 2012 is used with the positive temperature of 3.60 °C with a SD of 0.02.

In panel d the aerosol evolution on the ice surface is shown, with an amount of BC on the ice surface of 0.017 and 1.84 g m⁻² dust. Comparing it with the equivalent plot of the KAN_M station Fig. 5d the different scales have to be taken into account. Although the surface melt is about double here as compared to KAN_M the aerosol amount is less because of the lower englacial content. A small peak in the dust amount of the ice surface can be seen after the dust content of the snowpack has been deposited on ice. This also emphasises the relative magnitude of aerosols deposited from atmosphere as opposed to melted out aerosols. Since the time with (panel f) superimposed ice is much shorter here the time evolution of aerosols does not have the drop at the beginning when ice is exposed.

The high ice albedos at S5 makes it necessary to set the active fraction of BC and dust to zero. Hence the already low amount of aerosols may be mostly trapped in cryoconite holes and therefore not directly influence the ice albedo anymore.

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5 Discussion

5.1 General

In this paper we have introduced a model of dust and black carbon accumulation on the ice surface and its effect on ice albedo of a melting surface. In order to achieve this goal the model depends on several parts (see Fig. 3): temperature, precipitation, melt, snow thickness, aerosol accumulation and snow and ice albedo reduction. Each part has its own limitations and assumptions. Since the overall goal was to be applicable for long time-scales, it is desired to use a limited number of input data types and be computationally efficient. Therefore we parameterised the temperature by a trapezoid shape. This shape follows the measurements reasonable well especially during the melt season when temperature is important for the calculated melt and hence the resulting melt-out of englacial aerosols. Although at station S5 the start of the summer in 2011 and other years was set too early. Therefore it may be better to adjust the start and end of summer individually by location or parameterise it with an elevation dependency for example.

The precipitation was directly parameterised by taking the annual mean as a constant influx over the whole year. This is sufficient to get a comparable snow height evolution as to use the MAR model output because the annual signal are not effected by the high frequency variations. Nevertheless this can not simulate significant summer snowfall which would result in a sudden increase from low ice albedo to high dry snow albedos, as can be seen in the data. This could be realised by coupling the model to a regional climate model which would also allow to account for snow metamorphism via the specific surface area. To study the effect of englacial dust and BC simulations over millennia are required which would make this approach computationally too expensive. Nevertheless to study the long term trend this may not be necessary and the simple parameterisation of temperature and precipitation may be sufficient.

The constant annual cycles of surface albedo in Fig. 6b is not just a result of the fixed temperature and precipitation parameterisation, but also because the englacial dust

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100 (Warren and Wiscombe, 1980) to 200 (Gardner and Sharp, 2010) times more absorbent by mass than dust. Since the cores taken by Wientjes et al. (2011) were not analysed for its mineral dust content we choose the englacial dust value of 2000 ng g^{-1} . This is a value of dust concentration from the NGRIP ice core by Ruth (2007) for the late pleistocene. Although the age of the ice by C14 dating and ranges from 3071 BP to modern (Wientjes et al., 2012), and is therefore Holocene. Dust concentration of the Holocene in ice cores is about 100 times lower (Steffensen, 1997). Therefore we also tested the accumulation with just 20 ng g^{-1} at KAN_M and together with a higher active fraction of 0.6 similar albedo values can be reached although with a much lower total impurity amount of below one gm^{-2} . Wientjes et al. (2012) explained the higher concentration by a period of enhanced eolian activity in the Holocene. The source of dust of deep cores in the ice sheet interior revealed Asia as the main source of dust (Biscaye et al., 1997; Svensson et al., 2000) although cores from local ice caps (Renland and Hans Tausend) were found to have a more local source (Bory et al., 2003). There might be a “threshold elevation” up to which local dust is contributing and above that elevation large scale transport is dominating and transporting dust from other continents to the GrIS interior.

A detailed study of relative amount of impurities is required to further investigate the concentration to mass of each species. Our simulations give a quantitative estimate of the involved dynamics and the relative importance of outcropping of aerosols. The amount of outcropped material is relative robust, as it is depending on the amount of melt which simulations yielding similar results as observations (Van As et al., 2012) (see Fig. S1 in the Supplement) and the englacial concentration which resulted in a total amount in comparable to observations.

5.3 Reduced albedo at KAN_M after 2009

The albedo in summer 2009 was around 0.6 while in 2010 (see Fig. 6a) it dropped to around 0.4 and reached even lower values in the following years. We speculate that

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the residence-time of each species. For that purpose cameras on automatic weather stations could be useful as well as routinely taking surface samples at surface mass balance sites. In addition the microbiological activity and its direct effect on ice albedo has yet to be fully understood and once it is it can easily be incorporated into the model framework. Moreover, the specific surface of the ice i.e. the ablation crust is an important parameter which needs to be quantified.

Together with a transport scheme of englacial aerosols the present albedo model can be used to study the long term effects of natural and anthropogenic aerosols on future melt of glaciers and ice sheets. This model framework could be applied to further investigate the role of black carbon on the termination of the little ice age in the alps for example.

The full benefit of the present model is on long term projections of future sea level rise and can shed light on the way the Greenland ice sheet will be influenced by the dust and BC currently transported toward the ablation zone inside the ice. As the current source of BC in Greenland is probably Asia the total englacial amount stored in Greenland is still increasing. Fortunately due to modernisation the emissions are expected to decrease until 2030 (Streets, 2007). Nevertheless the full effect of past emissions still remains to come as the particles that melt free at the ice sheet margin, accumulate and impact the melt for long time to come.

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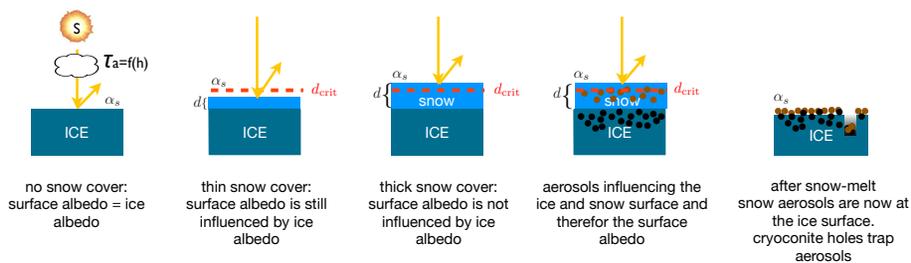


Figure 1. The relationship between ice albedo, snow albedo and aerosol concentration to surface albedo. As ice is exposed to melting, englacial aerosols are released and accumulate on the surface. Some impurities on the ice surface are trapped in cryoconite holes and are therefore not directly contributing to a lower surface albedo (Bøggild et al., 2010). In addition micro-biological activity can produce dark material on the snow and ice surface and can enhance the albedo reduction by aggregating the impurities.

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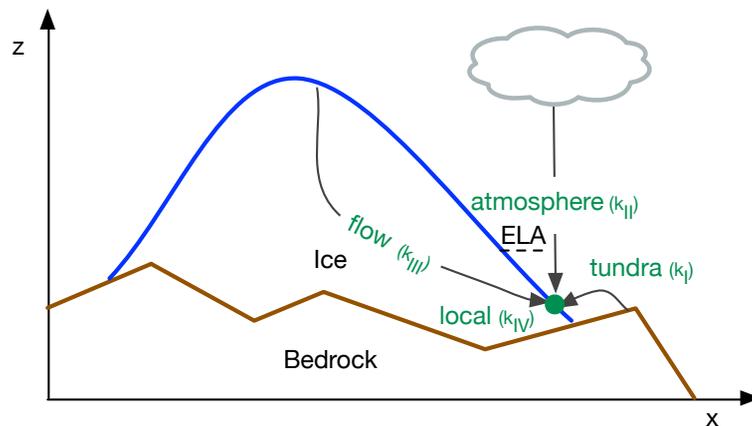


Figure 2. Cross section through an ice sheet: four different mechanisms of aerosol transport to the ablation zone. ELA stands for equilibrium line altitude.

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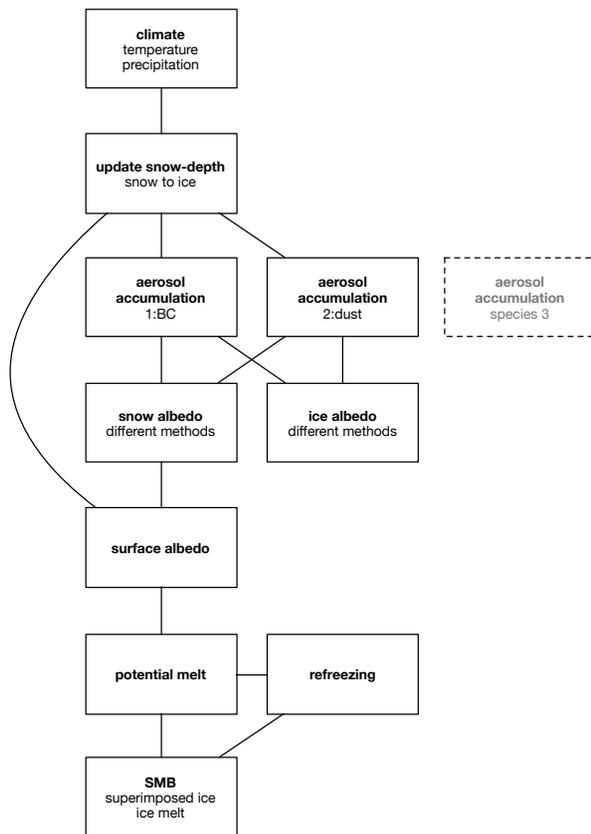



Figure 3. Flow chart of the surface mass balance model inside a loop which is solved for each time-step. Ultimately resulting in an annual SMB which can be used as boundary condition in an ice sheet model.

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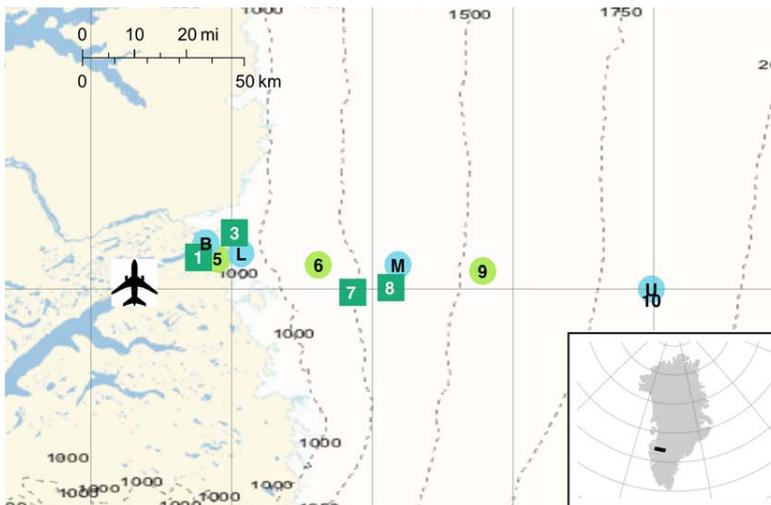


Figure 4. Weather stations and sampling sites at the K-transect on western Greenland. The airplane symbol marks the Kangerlussuaq airport. Blue circles mark the PROMICE automatic weather stations operated by GEUS and the green circles mark the ones operated by IMAU. The squares mark additional mass balance sites where ice samples have been taken (Wientjes et al., 2012). Station 9 is located near the equilibrium line and sites 6, 7, 8 and M are located in the dark region with lower albedos.

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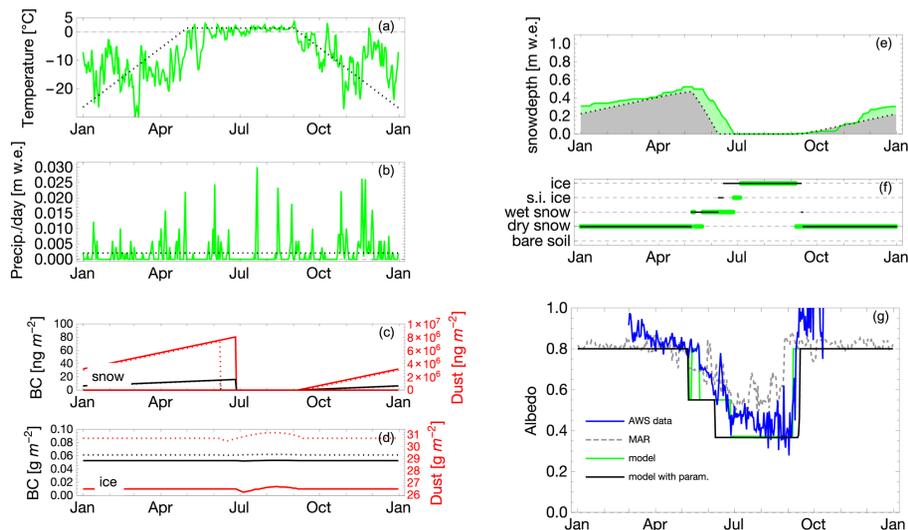


Figure 5. Detailed result of KAN_M station in 2010 with a simulation after spin-up using precipitation from MAR and temperature from the weather station, shown in green and with temperature and precipitation parameterisation (dotted black). Temperature in (a), daily precipitation in (b). Evolution of dust (red) and BC (black) amount in the snow pack and on the ice surface are shown in (c, d). The snow depth evolution in (e) together with the surface type (f) and aerosol content determine the surface albedo evolution in (g).

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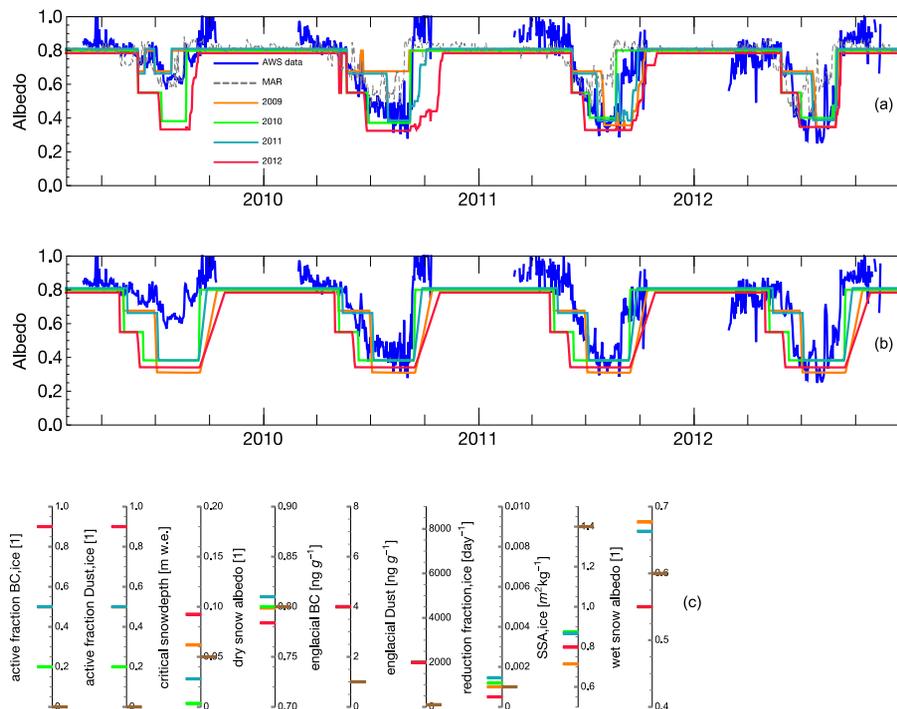


Figure 6. Albedo evolution of the KAN_M station over the years 2009 until the end of 2012. Panels (a, b) show AWS data in blue and different simulations. The parameters are optimised for different years and then used to simulate the entire period. In (a) the precipitation comes from MAR model output and the temperature from AWS data and in (b) those are parameterised. Panel (c) shows the range of settings for each year and for Station S5 in brown on the right side of the scales.

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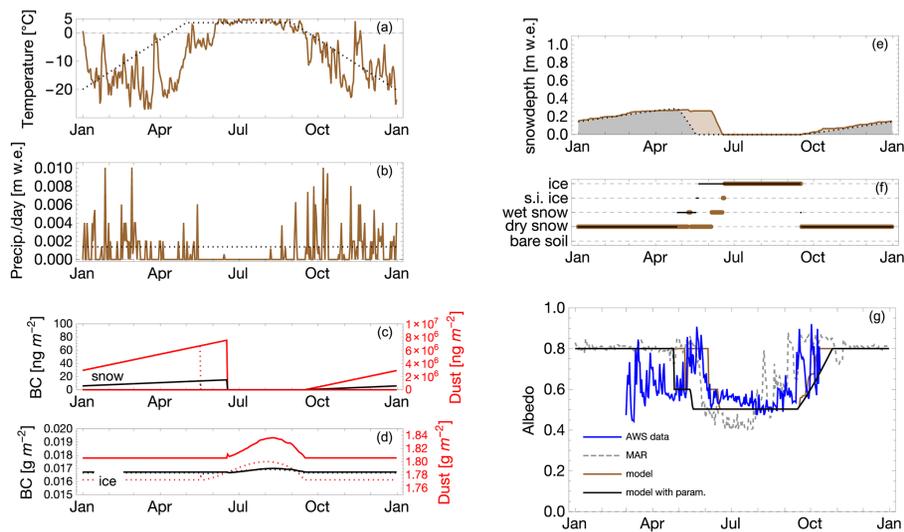


Figure 7. Details of station S5 in the year 2011 with the same plot arrangement as in Fig. 5.

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