Climate change and forest fires synergistically drive widespread melt events of the Greenland Ice Sheet

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In July 2012, over 97% of the Greenland Ice Sheet experienced surface melt, the first widespread melt during the era of satellite remote sensing. Analysis of six Greenland shallow firm cores from the dry snow region confirms that the most recent prior widespread melt occurred in 1889. A firm core from the center of the ice sheet demonstrated that exceptionally warm temperatures combined with black carbon sediments from Northern Hemisphere forest fires reduced albedo below a critical threshold in the dry snow region, and caused the melting events in both 1889 and 2012. We use these data to project the frequency of widespread melt into the year 2100. Since Arctic temperatures and the frequency of forest fires are both expected to rise with climate change, our results suggest that widespread melt events on the Greenland Ice Sheet may begin to occur annually by the end of the century. These events are likely to alter the surface mass balance of the ice sheet, leaving the surface susceptible to further melting.

The massive Greenland Ice Sheet (GIS) experiences annual melting at low elevations near the coastline. However, surface melt is extremely rare over ~600,000 km² of the “dry snow region” in its center (1). In July 2012, ground-based observations confirmed indications from satellite imagery that, for several days, surface melt had occurred over 97% of the ice sheet’s surface (Fig. 1) (2). Melting in the dry snow zone does not contribute to sea level rise; instead, water from these events percolates into the snowpack and refreezes (3). Subsequently, it was determined that the radiative effects of a thin layer of low-level “liquid clouds” added to warming conditions over the GIS in 2012 (4). Evidence from firm cores collected at multiple sites demonstrated that the most recent previous spatially extensive melt occurred in 1889. Thus, at present, widespread melt events are anomalous.

Interestingly, although 1889 and 2012 were relatively warm years, there have been even warmer years over the past century when the snow in the coldest part of the ice sheet did not melt. As such, we hypothesized that, in addition to relatively warm temperatures, the snow albedo may also play a critical role in inducing melt in the dry snow region. In this paper we evaluate the causes of both the 1889 and the 2012 melt events and estimate the expected frequency of these events in the context of continuing climate change in the Arctic.

Results and Discussion

Ice layers formed from refrozen snowmelt in porous firn are prominent upon visual inspection. A prominent ice layer dated to 1889 in multiple firm cores from central Greenland (5–7) and northwest Greenland (8) has been briefly mentioned in the literature. It has been traditionally thought that these melt layers were likely part of a widespread melt event in 1889, but the extent of the melt was unknown. Here, we present physical evidence of a melt layer dating to 1889 in six additional Greenland firm cores from four sites: Summit, NEEM, D4, and ACT 3 (Fig. S1 and Table S1). Thus, the 1889 melt event appears to have been very widespread, and included the dry snow region.

During most summers in the 20th and 21st centuries, it has been too cold to permit snowmelt in the dry snow region of the GIS. Accordingly, the widespread melting in 1889 and in 2012 occurred during periods that were unusually warm (Fig. 2). We used firm core records of stable water isotopes, δ18O, as a proxy for temperature (9), although it should be noted that there was only a moderate correlation (r² = 0.29, Fig. S2) between temperature and δ18O at Summit from 2000 to 2010. As methods for the indirect estimation of temperature improve, more accurate historical temperature records may become available. Nevertheless, these data suggest that 1889 was a particularly warm year (Fig. 2C). However, this was not the warmest year recorded in the firm. Temperatures were warmer in 1785, for example, but melting in the dry snow region did not occur in that year (Fig. 24). Similarly, widespread melting in the dry snow region did not occur during the most recent record-breaking melt extent years of 2002, 2007, or 2010 (10–12). Thus, high temperatures alone are often not enough to cause widespread melt. Indeed, continuous measurements of near-surface conditions at Summit Station from the Greenland Climate Network (GC-Net) instruments show that the temperature reached 0 °C on July 12, 2012 (13), providing ~0.94 MJ m⁻² of energy, although ~1.1 MJ m⁻² of energy is needed to cause melting in the accumulation area of the ice sheet (calculated in SI Materials and Methods). Under present conditions in the dry snow region, this energy threshold cannot be attained from the average summer temperature, irradiance, and albedo. Despite melting in the ablation zone of the ice sheet during summer, unusually warm conditions have not been sufficient to cause widespread melting into the dry snow region.

The albedo of snow is an important factor in snowmelt. If temperatures at the surface of the ice sheet are warm and snow is allowed to age in the absence of fresh accumulation, it will coarsen, and this changing morphology can lower the albedo by up to 14% (14). In addition, black carbon (BC) from incomplete fossil fuel combustion or biomass burning can reach the top of the sheet and reduce the albedo by up to 7% (15, 16). We measured BC on a Summit shallow core and also on a near-surface sample of firm containing the 2012 melt layer with an

Significance

Through an examination of shallow ice cores covering a wide area of the Greenland Ice Sheet (GIS), we show that the same mechanism drove two widespread melt events that occurred over 100 years apart, in 1889 and 2012. We found that black carbon from forest fires and rising temperatures combined to cause both of these events, and that continued climate change may result in nearly annual melting of the surface of the GIS by the year 2100. In addition, a positive feedback mechanism may be set in motion whereby melt water is retained as refrozen ice layers within the snow pack, causing lower albedo and leaving the ice sheet surface even more susceptible to future melting.


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inductively coupled plasma mass spectrometry (ICP-MS) continuous flow analysis system. This included measurements of total BC concentrations, as well as ammonium, $\delta^{18}O$, and a suite of other measurements that were used to establish the depth–age scale. The presence of a high concentration of ammonium concurrent with the BC indicates that the source of the BC was large boreal forest fires (17). Since BC coeval with high concentrations of non-sea-salt sulfur (nss-S) indicates a fossil fuel source of BC (18), we consider low nss-S concentrations concurrent with BC another indication of a forest fire source.

The total BC concentration varies inter- and intra-annually in the firn. The annual average BC concentration was 1.7 ng g$^{-1}$ during the preindustrial period (1750–1850 A.D.), 4.0 ng g$^{-1}$ during the early industrial period (1851–1950), and 2.3 ng g$^{-1}$ from 1951 to present (Fig. 2). Over our 262-y record, there are four large spikes in the annual average BC concentration of 10.4, 10.4, 14.3, and 14.4 ng g$^{-1}$ in 1868, 1889, 1908, and 2012, respectively (Fig. 2). Visual inspection of the firn core revealed the presence of an ice layer in 1889, and surface samples contained a melt layer formed on July 12, 2012. Figure courtesy of Dorothy Hall, NASA Goddard Space Flight Center.

![Fig. 1. Melt extent over the GIS determined from Oceansat-2 satellite scatterometer, Special Sensor Microwave Image/Sounder, and Moderate-resolution Imaging Spectroradiometer satellite data for (A) July 8, 2012, and (B) July 12, 2012. Red areas indicate melt detected by the satellites, white areas indicate no melt, and blue represents ocean. The surface of almost the entire ice sheet, including the dry snow region, experienced melt on July 12, 2012. Figure courtesy of Dorothy Hall, NASA Goddard Space Flight Center.](#)

![Fig. 2. (Lower) The annual average BC concentrations (ng g$^{-1}$) from 1750 to 2010 of the Summit-2010 firn core and the 2012 surface section. (Upper) Sections of the BC record along with $\delta^{18}O$ and ammonium records, plotted on a relative scale normalized to the maximum and minimum values in each record, for the time intervals (A) 1783–1788, (B) 1865–1870, (C) 1887–1892, and (D) 1905–1910, as well as (E) the normalized average value of BC and ammonium concentrations from the July 2012 surface sample, and approximate $\delta^{18}O$. These time intervals demonstrate extreme scenarios in the center of the GIS with (B–E) depicting the highest concentrations of BC, and (A) the warmest temperature since 1750, but widespread melt events only occurred in 1889 and 2012. In C and E, melt occurred because of the deposition of high concentrations of BC and ammonium, indicating an albedo reduction due to BC from summer forest fires. Importantly, these deposition events occurred during warm summers. In B, a high concentration of BC and presence of ammonium during a cooler summer suggest that the surface was below the energy threshold for melt. In D, the highest concentrations of BC and ammonium in the record were recorded during an average summer, suggesting that the BC was deposited at a time of the year when the available surface energy was well below the threshold for melt. The warmest temperature recorded in the core occurred in 1785, but widespread melting did not occur due to low BC concentration.](#)
Unlike other years where BC peaks occurred but melting of the ice sheet was minimal, widespread melting of the entire GIS surface did occur in 1889 and 2012. Why did it occur in these years and not the others? Our data suggest that during these two years, abnormally warm summer temperatures combined with BC deposited on the ice sheet to reduce albedo below a critical threshold. In 1889, the high BC (10.4 ng g\(^{-1}\)) and ammonium (7.0 μM) concentrations observed in firn cores indicate that the source of BC was forest fires (Fig. 2C and Fig. S3). Indeed, biomass burning due to large forest fires was documented in 1889 (19, 20). Further, analysis of the BC particles is necessary to fingerprint a specific geographic source of these deposits, and is beyond the scope of this study. However, air mass back-trajectories before the 2012 melting event show sources ranging from Siberia to North America (Fig. S4). Moreover, measurements of δ\(^{18}O\) (−32.4 ‰) demonstrate that this BC deposition occurred during an anomalously warm summer (Fig. 2C and Fig. S3). Similarly, in July 2012, the surface of the ice sheet experienced unusually warm temperatures, enhanced by a thin layer of low-level liquid clouds (3), that averaged −1.9 °C, and exceeded 0 °C for −5 min, on July 12 (13). Concurrently, BC was deposited with an ammonium signature (6.1 μM) indicating a forest fire source (Fig. S3), originating from the significant forest fires in Siberia (21) and North America in late June and July 2012 (22, 23) that were transported to Summit in early July (Fig. S4). The albedo reduction (∼6%) caused by warming and BC presence (14, 16) on July 12, 2012, at Summit provided an extra 0.3 MJ m\(^{-2}\) of energy to the surface, pushing the surface energy balance over the threshold of 1.1 MJ m\(^{-2}\) for melting. Therefore, changes in albedo due to BC deposition from Northern Hemisphere forest fires together with unusually warm conditions were the driving factors in the extensive 1889 and 2012 melt events across the ice sheet.

Over the next century, climate change is predicted to raise both the average summer temperature and the frequency of forest fires. The Arctic mean summer temperature is predicted to increase 2–9 °C by the end of the century (24), and forest fire frequency is expected to at least double per 1 °C rise in temperature (25). Based on the frequency of high BC concentrations (>9 ng g\(^{-1}\)) reaching the summit of the GIS over the past three centuries (0.015) and the projected change in mean temperature over the next century, we modeled the potential for climate change to increase the surface energy balance over time and change the frequency of widespread melt events on the Greenland Ice Sheet (Fig. 3). Considering only an increase in the surface temperature through the end of the century (in the absence of BC deposition), we can expect the probability of widespread melt to rise to 0.17 or 0.70 for the 2 °C and 9 °C scenarios, respectively. Thus, in the most conservative scenario, by the year 2100, we predict that widespread melting of the surface of the GIS should occur every 6 y on average. When the expected increase in forest fire frequency due to climate change is also included in projections, our simulations suggest that for the 2 °C and 9 °C scenarios, the probability of widespread melt will increase to 0.22 or 0.94, respectively, by the year 2100. Thus, in the least conservative scenario, our projections suggest that the GIS should experience almost annual widespread melting by the year 2100.

This widespread surface melting is likely to alter the surface mass balance of the ice sheet (26). In the dry snow region, retained ice layers generated from the refrozen melt water will further lower the albedo until covered by the next snowfall, and also warm the snowpack upon refreezing. Also, melting can concentrate particles such as BC, which would also lower the surface albedo (27). Therefore, these frequent widespread surface-melting events may initiate a positive feedback mechanism whereby the Greenland Ice Sheet melts even more often than our projections indicate.

In the past, surface melting of the Greenland Ice Sheet over a wide extent including the high, cold, dry snow region occurred only during rare synchronous periods of warm conditions accompanied by high snow surface concentrations of BC. In the future, warmer temperatures and more frequent Northern Hemisphere forest fires due to climate change may increase the frequency of these widespread melt events, contributing to the further demise of the Greenland Ice Sheet.

Materials and Methods

In each firn core, we used an SP2 intercavity laser-based instrument for measuring BC (with a detection limit of −0.02 ng g\(^{-1}\)) (18), a Picarro L2130-i cavity-ring-down-based laser instrument to measure stable water isotopes (δ\(^{18}O\)) (28), and fluorimetry to measure ammonium concentrations (29). For comparison, these records were normalized by the minimum and maximum values in their respective profiles (0.023 ng g\(^{-1}\) and 14.4 ng g\(^{-1}\) of BC, 0.003 μM and 8.86 μM of ammonium, and −37.6‰ and −30.6‰ δ\(^{18}O\)).

The amount of energy needed to generate surface melt and melt water percolation was calculated from the summation of Dingman’s equations for the warming phase (Q\(_{w}\)), ripening phase (Q\(_{rip}\)), and output phase (Q\(_{m}\)) of melt (30). The amount of energy available at the surface was calculated from a surface energy budget of the top 1 cm of depth by 1 m\(^2\) of the snow surface. The projected energy available at the surface was calculated for four possible scenarios over the next 90 y: a 2 °C and 9 °C rise in temperature both with and without an associated rise in forest fire frequency. We calculated the temperature effect on the incoming and outgoing longwave radiation, as well as on the albedo, which decreases from changing microstructure of the warming snowpack (up to −7.5%) (13). We then applied this increase in available energy to a 10 y dataset of daily summer (June 1 to August 31) net radiation. We calculated the effect of BC on the albedo, a 1% reduction when present, by conservatively assuming a doubling in frequency of forest fires per 1 °C rise in temperature (25). In other words, with every 1 °C rise in temperature, we assumed that the probability of a 1% albedo reduction would double. We applied the increase in fire frequency to the 262-y BC record at Summit. At 5-y time steps, we sampled the two energy datasets (temperature effect and BC effect) 1,000 times with replacement using the bootstrapping method; the summation of these two outputs at each time step yielded the projected total surface energy available.

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Supporting Information

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S1 Materials and Methods

Firn Core Analysis. We examined the stratigraphy of six firn cores, identified as Summit-2007, Summit-2010, D4, ACT 3, NEEM 2008 S1, and NEEM 2009 S2 in this paper, which cover a broad spatial extent over the Greenland Ice Sheet (GIS) (Fig. S1). A distinct refrozen melt layer was visible in each core, denoting a melt event in each core. Using depth–age scales for each site, we identified the ages of the melt layers (Table S1). The melt layer in both NEEM firn cores date to mid-1888, which seems reasonable, because the melt water would need to percolate to a firn layer cold enough to allow refreezing. Since the NEEM site is lower in elevation and warmer than Summit, it is reasonable to expect the melt water to percolate into the 1888 firn layers before refreezing. Annual snow accumulation rates at D4 (413 kg m$^{-2}$ yr$^{-1}$) and ACT 3 (662 kg m$^{-2}$ yr$^{-1}$) are much higher than the modern rates at Summit (223 kg m$^{-2}$ yr$^{-1}$) and NEEM (207 kg m$^{-2}$ yr$^{-1}$), so melt penetration at these sites is limited to a few months of record at most. Lastly, a vertical profile sample of the near-surface snow from Summit Greenland, which incorporated the 2012 melt layer, was also analyzed.

In this paper, we focus on the chemical and elemental analysis of the Summit-2010 firn core and Summit-2012 near-surface sample, because Summit is the least likely site to experience melt during a widespread melt event due to its position at the summit of the GIS. We measured black carbon (BC), ammonium as a proxy of forest fires (1), and δ$^{18}$O as a proxy for temperature (2) (Fig. S2). We use an SP2 intercavity laser-based instrument for measuring BC with a detection limit of ~0.02 ng g$^{-1}$ (3), a Picarro L2130-i cavity-ring-down-based laser instrument to measure the stable water isotopes (δ$^{18}$O) (4), and fluorimetry to measure the ammonium concentrations (5). In Fig. S2, we confirm that the temperature and δ$^{18}$O trends are related at Summit for the recent past.

To clearly demonstrate the effect of high concentrations of BC and ammonium, as well as the ratio of δ$^{18}$O on the same scale, we plotted the three variables together on a normalized scale. The records were normalized by the maximum and minimum values of the full record (1750–2010, 2012 A.D.) for each variable (Fig. S3), which were 0.023 ng g$^{-1}$ and 14.4 ng g$^{-1}$ of BC, 0.003 µM and 8.86 µM of ammonium, and ~37.6‰ and ~30.6‰ δ$^{18}$O, respectively. These normalized values were then plotted on a scale of 0–1 for each time series shown in Fig. 2 A–D. The average concentrations of BC and ammonium from the 2012 surface sample were also normalized with the maximum and minimum values from their respective full records (1750–2010, 2012 A.D.), and plotted in the bar graph in Fig. 2E. Stable water isotopes were not measured on the 2012 surface sample, so the average temperature measured at Summit from June–August of 2012 (~12.4 °C) was compared with average summer temperatures in the recent history and found to be similar to that of 2010 (~12.2 °C) (6). Therefore, we assigned the 2012 surface sample a normalized δ$^{18}$O value equivalent to that of 2010 (0.86).

To investigate the source of the BC found at Summit, we found data showing large-scale forest fires burning in North America (7, 8) and Siberia (9) in late June and early July of 2012, respectively. We identified potential source regions of BC using a HYSPPLIT 5-d back-trajectory model of particles arriving at Summit during the time period of June 28 through July 12, 2012 (Fig. S4). In Fig. S4, the blue dots indicate the origin of particles arriving at Summit June 28 through July 3 and red dots indicate the origin of those arriving July 7–12. Therefore, Fig. S4 demonstrates that air masses from both North America and Siberia reached Summit during this time, which could have carried BC particles.

Melt Calculations. The amount of energy needed to generate surface melt and percolation was calculated from Dingman’s equations (10) for the warming phase ($Q_{cc}$), ripening phase ($Q_{m2}$), and output phase ($Q_{m3}$)

$$Q_{cc} = -c_i \rho_w h_m(T_s - T_m)$$ \[S1\]

$$Q_{m2} = \theta_{ret} h_s \rho_w \lambda f$$ \[S2\]

$$Q_{m3} = (h_m - h_{wret}) \rho_w \lambda f$$ \[S3\]

where $c_i$ is the specific heat capacity of ice (2.102 kJ kg$^{-1}$ K$^{-1}$), $\rho_w$ is the density of water (1000 kg m$^{-3}$), $h_m$ is the snowpack water equivalent (meters), $T_s$ is the average temperature of the snow, $T_m$ is the melting temperature (273 K), $\theta_{ret}$ is the maximum volumetric water content, $h_s$ is the snow depth (meters), $\lambda f$ is the latent heat of fusion (334 kJ kg$^{-1}$), and $h_{wret}$ is the liquid water retaining capacity of the snowpack (meters). The $\theta_{ret}$ can be calculated from the empirical relationship

$$\theta_{ret} = -0.0745 \left( \frac{\rho_s}{\rho_w} \right) + 0.000267 \left( \frac{\rho_s}{\rho_w} \right)^2.$$ \[S4\]

where $\rho_s$ is the snowpack density. The $h_{wret}$ can be calculated from the relationship

$$h_{wret} = \theta_{ret} h_S.$$ \[S5\]

The snowpack water equivalent depth, $h_m$, can be determined from

$$h_m = \frac{\rho_s h_S}{\rho_w}.$$ \[S6\]

Substituting Equations S4–S6 into Equations S1–S3, allows us to solve Equations S1–S3 for the energy required. The summation of $Q_{cc}$, $Q_{m2}$, and $Q_{m3}$ gives the theoretical amount of energy required for surface melt and percolation (10).

To determine the amount of energy available at the surface, we calculated the surface energy balance from

$$S = (1 - \alpha) K_{in} + L_{in} - L_{out} + H + LE,$$ \[S7\]

where $S$ is the energy flux at the snow surface (W m$^{-2}$), $\alpha$ is the albedo, $K_{in}$ is the incoming shortwave radiation, $L_{in}$ is the incoming longwave radiation, $L_{out}$ is the outgoing longwave radiation, $H$ is the turbulent sensible heat flux, and $LE$ is the turbulent latent heat flux (11).

Given an average $T_S$ of ~15 °C, $h_s$ of 1 cm, and $\rho_s$ of 300 kg m$^{-3}$, it requires 1.1 MJ m$^{-2}$ of energy to cause surface melt and percolation. Since the melt layers are ~1–2 cm thick, it is reasonable to expect surface snow on the order of 1 cm to melt ($h_S$).

The amount of energy available at the surface for melting in July 2012 was calculated using the average daily incoming radiation, 409 W m$^{-2}$, an albedo that includes an average amount of coarsening, 0.77, longwave radiation of 205 W m$^{-2}$, outgoing radiation of 317 W m$^{-2}$, sensible heat flux of 4 W m$^{-2}$, and latent heat flux of 3 W m$^{-2}$ (11), and, over 7 h, gives 0.94 MJ m$^{-2}$ of...

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energy. The concentration of BC at the surface in 2012 provided at least 1% reduction in albedo (12), and surface snow metamorphism due to warmer temperatures provided a 5% reduction in albedo (13). These two effects combined (∼6% reduction in albedo) provided at least 0.3 MJ m⁻² of energy, tipping the surface energy available over the 1.1 MJ m⁻² threshold to 1.24 MJ m⁻² and thus allowing melting to occur.

**Modeling Future Widespread Melt.** To model how the probability of widespread melt events will change in the future due to climate change, the forecasted rise in temperature (14) and increase in forest fire frequency (15) were used. To incorporate a warming average summer temperature in the Arctic, the average energy required to cause melt and percolation was calculated with the Intergovernmental Panel on Climate Change ARS scenarios predicting Arctic summer temperature increases of 2 °C and 9 °C by 2100 (14).

The projected energy available at the surface was calculated for four scenarios: 2 °C end-of-century temperature rise, 2 °C rise with increasing forest fire frequency, 9 °C end-of-century temperature rise, and 9 °C rise with increasing forest fire frequency. We calculated the temperature effect on the incoming and outgoing longwave radiation, as well as on the albedo for 5-y time steps in each of the four scenarios. The albedo will decrease by ∼3% per °C rise in average temperature due to snow metamorphism at the surface, and will reach a maximum of ∼7.5% reduction (13). The increase in available energy was applied to a 10-y dataset of daily summer (June 1 through Aug. 31) net radiation, and the annual average was generated from bootstrapping 1,000 random samples from the dataset with replacement.

BC present at the surface of the ice sheet increases the amount of energy available at the surface by decreasing the albedo by ∼1% when present (12, 16). With climate change, it is projected that forest fires frequency will increase twofold to fourfold per °C in warming. We generated a dataset for the effect of BC on the surface energy available by taking the 262-y BC record at Summit and giving a reduction of albedo by 1% to large BC spikes (> 9 ng g⁻¹). Using a doubling in frequency of forest fire BC reaching the GIS per 1 °C, we randomly added large BC events based on the warming to the dataset. For each time step, the datasets were also sampled at random 1,000 times with replacement using the bootstrapping method.

To calculate the probability of widespread melt events occurring through 2100, we combined the 1,000 energy values due to warming with those due to BC and generated a probability that the energy available was greater than the 1.1MJ m⁻² needed to cause widespread melting events. We performed these calculations for each time step and plotted the probabilities in Fig. 3.

Fig. S1. A map of the four Greenland firn core site locations, which are denoted by black crosses. ACT3 is located at (67.0° N, 43.6° W), D4 is located at (71.4° N, 44.0° W), Summit is located at (72.6° N, 38.5° W), and NEEM is located at (79.0° N, 50.0° W).

Fig. S2. Annual average temperature vs. annual average $\delta^{18}$O at Summit, Greenland, from 2000 to 2010. The linear regression line shows a moderate correlation between temperature and $\delta^{18}$O ($r^2 = 0.29$).
Fig. S3. The non-sea-salt sulfur (nss-S), stable water isotopes ($\delta^{18}O$), BC, and ammonium records from the summit-2010 firn core, as well as BC and ammonium concentrations from the 2012 Summit near-surface sample.
Table S1. Site, depth, and age of the Greenland 1889 melt layer

<table>
<thead>
<tr>
<th>Core</th>
<th>Melt depth, meters</th>
<th>Age, A.D.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summit-2007</td>
<td>47.1</td>
<td>1889.2</td>
</tr>
<tr>
<td>Summit-2010</td>
<td>47.0</td>
<td>1889.1</td>
</tr>
<tr>
<td>NEEM 2009 S2</td>
<td>43.1</td>
<td>1888.5</td>
</tr>
<tr>
<td>NEEM 2008 S1</td>
<td>42.6</td>
<td>1888.5</td>
</tr>
<tr>
<td>ACT 3</td>
<td>108.7</td>
<td>1889.3</td>
</tr>
<tr>
<td>D4</td>
<td>72.4</td>
<td>1889.4</td>
</tr>
</tbody>
</table>

Fig. S4. A plot of particle origin to Summit, Greenland, from June 28 through July 12, 2012, as calculated by HYSPLIT with a 5-d back-trajectory. The shade of the dots indicate the date of landing at Summit, with blue shades showing the origin of particles landing on June 28 through July 3 and red shades on July 7–12.