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Blowing Snow Sublimation and Transport over Antarctica from 11 Years of CALIPSO Observations

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ABSTRACT

Blowing snow processes commonly occur over the earth's ice sheets when near surface wind speed exceeds a threshold value. These processes play a key role in the sublimation and re-distribution of snow thereby influencing the surface mass balance. Prior field studies and modeling results have shown the importance of blowing snow sublimation and transport on the surface mass budget and hydrological cycle of high latitude regions. For the first time, we present continent-wide estimates of blowing snow sublimation and transport over Antarctica based on direct observation of blowing snow events. We use an improved version of the blowing snow detection algorithm developed for previous work that uses atmospheric backscatter measurements obtained from the CALIOP lidar aboard the CALIPSO satellite. The blowing snow events identified by CALIPSO and meteorological fields from MERRA-2 are used to compute the sublimation and transport rates. Our results show that maximum sublimation occurs along and slightly inland of the coastline. This is contrary to the observed maximum blowing snow frequency which occurs over the interior. The associated temperature and moisture re-analysis fields likely contribute to the spatial distribution of the maximum sublimation values. However, the spatial pattern of the sublimation rate over Antarctica is consistent with modeling studies and precipitation estimates. Overall, our results show that Antarctica average integrated blowing snow sublimation is about 393.4 ± 138 Gt yr^{-1} which is

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considerably larger than previous model-derived estimates. We find maximum blowing snow transport amount of 5 Megatons km^{-1} yr^{-1} over parts of East Antarctica and estimate that the average snow transport from continent to ocean is about 3.68 Gt yr^{-1}. These continent-wide estimates are the first of their kind and can be used to help model and constrain the surface-mass budget over Antarctica.

Keywords: Blowing snow, sublimation, transport, CALIPSO, Antarctica, surface mass balance

1 Introduction

The surface mass balance of the earth's great ice sheets that cover Antarctica and Greenland is one of today's most important topics in climate science. The processes that contribute to the mass balance of a snow or ice-covered surface are precipitation, surface runoff (melting), removal/loss (erosion/sublimation) and deposition of snow mass (wind transport). Sublimation of snow can occur at the surface but is greatly enhanced within the atmospheric column of the blowing snow layer. The contributions of these processes to the mass balance vary greatly spatially, and can be highly localized and very difficult to quantify.

It is well known that the Arctic is experiencing rapid warming and loss of sea ice cover and thickness. In the past few decades, the Arctic has seen an increase in average surface air temperature by 2 °C (Przybylak, 2007). Modeling studies suggests an increase in annual mean temperatures over the Arctic by 8.5 ± 4.1 °C over the current century that could lead to a decrease in sea ice cover by 49 ± 18 % (Bintanja and Krikken, 2016). While the Antarctic has experienced an increase in average surface temperature, most of the warming is observed over West Antarctica at a rate of 0.17 °C per decade from 1957 to 2006 (Steig et al., 2009; Bromwich et al., 2013). Such surface warming undoubtedly has implications for ice sheet mass balance and sea level rise mainly through the melting term of the mass balance equation. However, the other processes affecting the mass balance of ice sheets may also be experiencing changes that are difficult to identify and quantify. For instance, models have shown that in a warming climate, precipitation should increase over Antarctica and most of it will fall as snow (Church et

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Discussion started: 3 April 2017

al., 2013). If snowfall is increasing, perhaps the frequency of blowing snow and subsequently the magnitude of transport and sublimation will increase as well. Thus, understanding how these processes affect the overall mass balance of the ice sheets and how they may be responding to a changing climate, is of growing concern.

In addition to ice sheet mass balance, sublimation of blowing snow is also important for the atmospheric moisture budget in high latitudes. For instance, in the Canadian Prairies and parts of Alaska sublimation of blowing snow was shown to be equal to 30 % of annual snowfall (Pomeroy et al., 1997). About 50 % of the wind-transported snow sublimates in the high plains of southeastern Wyoming (Tabler et al., 1990). Adequate model representation of sublimation processes are important to obtain reliable prediction of spring runoff and determine the spatial distribution/variability of energy and water fluxes and their subsequent influence on atmospheric circulation in high latitude regions (Bowling et al., 2004).

Over Antarctica, blowing snow occurs more frequently than anywhere else on earth. Models driven by long-term surface observations over the Neumayer station (East Antarctica), estimate that blowing snow sublimation removes up to 19 % of the solid precipitation (Van den Broeke et al., 2010). Over certain parts of the Antarctica, where persistent katabatic winds prevail, blowing snow sublimation is found to remove up to 85 % of the solid precipitation (Frezzotti et al., 2002). Over coastal areas up to 35% of the precipitation may be removed by wind through transport and sublimation (Bromwich 1988). Das et al., (2013) concluded that ~ 2.7–6.6 % of the surface area of Antarctica has persistent negative net accumulation due to wind scour (erosion and sublimation of snow). These studies show the potential role of the blowing snow sublimation process in the surface mass balance of the earth's ice sheets.

For the current work, we focus on blowing snow processes over the Antarctic region. Due to the uninhabited expanse of Antarctica and the lack of observations, prior, continent-wide studies of blowing snow sublimation over Antarctica had to rely on parameterized methods that use model re-analysis of wind speed and low level moisture. The presence of blowing snow is inferred from surface temperature, wind speed and snow age (if known). In a series of papers on the modeling of blowing snow, Dery and Yau (1998, 1999, 2001) develop and test a parameterization of blowing snow sublimation. Their results show that most sublimation occurs

87 along the coasts and over sea ice with maximums in some coastal areas of 150 mm snow water
88 equivalent (swe). Lenearts et al., (2012a) utilized a high resolution regional climate model
89 (RACMO2) to simulate the surface mass balance of the Antarctic ice sheet. They found drifting
90 and blowing snow sublimation to be the most significant ablation term reaching values as high
91 as 200 mm yr^{-1} swe along the coast. There has been some work done on blowing snow
92 sublimation from field measurements, but the data are sparse and measurements are only
93 available within the surface layer (< 10 m). For instance, average monthly rates of sublimation
94 calculated for Halley Station, Antarctica, varied between 0.13 to 0.44 mm day−1 (King et al.
95 2001).

96 While transport of blowing snow is considered to be less important than sublimation in terms
97 of mass balance of the Antarctic ice sheet, erosion and transport of snow by wind can be
98 considerable in certain regions. Das et al., (2013) have shown that blue ice areas are frequently
99 seen in Antarctica. These regions exhibit a negative mass balance as all precipitation that falls is
100 either blown off or sublimated away. Along the coastal regions it has been argued that
101 considerable mass is transported off the coast via blowing snow in preferential areas dictated
102 by topography (Scarchilli et al., 2010). In the Tera Nova Bay region of East Antarctica, manned
103 surface observations show that drifting and blowing snow occurred 80 % of the time in fall and
104 winter and cumulative snow transport was 4 orders about of magnitude higher than snow
105 precipitation. Much of this airborne snow is transported off the continent producing areas of
106 blue ice. Such observations raise questions as to how often and to what magnitude continent to
107 ocean transport occurs. This is important, particularly for Antarctica where the coastline
108 stretches over 20,000 km in length and where prevailing strong winds through most of the year.
109 Due to the sparsity of observations, the only way to estimate the mass of snow being blown off
110 the coast of Antarctica is by using model parameterizations. Now, for the first time, satellite
111 observations of blowing snow can help better ascertain the magnitude of this elusive quantity.

112 Considering that the accuracy of model data is questionable over Antarctica, and the
113 complicated factors that govern the onset of blowing snow, it is difficult to assess the accuracy
114 of the parameterization of blowing snow sublimation and transport. Recently, methods have

been developed to detect the occurrence of blowing snow from direct satellite observations. Palm et al., (2011) show that blowing snow is widespread over much of Antarctica and, in all but the summer months, occurs over 50 % of the time over large areas of East Antarctica. In this paper, we present a technique that uses direct measurements of blowing snow from the CALIPSO satellite lidar combined with The Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2) re-analysis fields of moisture, temperature and wind to quantify the magnitude of sublimation and mass transport occurring over most of Antarctica (north of 82 south). Section 2 discusses the method used to compute blowing snow sublimation from CALIPSO and MERRA-2 data. In Sect. 3 we show results and compare with previous estimates of sublimation. In Sect. 4 we examine sources of error and their approximate magnitudes. Summary and discussion follow in Sect. 5.

2 Method

The method developed for detection of blowing snow using satellite lidar data (both ICESat and CALIPSO) was presented in Palm et al., (2011). That work showed examples of blowing snow layers as seen by the calibrated, attenuated backscatter data measured by the CALIOP instrument on the CALIPSO satellite. CALIOP (Cloud-Aerosol Lidar with Orthogonal Polarization) is a two wavelength (532 and 1064 nm) backscatter lidar with depolarization at 532 nm and has been operating continuously since June of 2006 (Winker et. al., 2009). In the lower 5 km of the atmosphere, the vertical resolution of the CALIOP backscatter profile is 30 m. The CALIOP backscatter profiles are produced at 20 Hz, which is about a horizontal resolution of 330 m along track. The relatively strong backscattering produced by the earth's surface is used to identify the ground bin in each profile. After the ground signal is detected, each 20 Hz profile is examined for an elevated backscatter signal (above a pre-defined threshold) in the first bin above the ground. If found and the surface wind speed is greater than 4 m s^{-1}, successive bins above that are searched for a 80 % decrease in signal value, which is then the top of the layer. Limited by the vertical resolution of the signal, our approach has the ability to identify blowing snow layers that are roughly 15-20 m or more in thickness. Thus, drifting snow which is

143 confined to 10 m or less and occurs frequently over Antarctica would not be reliably detected.
144 The signal from these layers is likely inseparable from the strong ground return. More
145 information on the blowing snow detection algorithm can be found in Palm et al., (2011).

146 For the work done in this paper we have created a new version of the blowing snow detection
147 algorithm which strives to reduce the occurrence of false positive blowing snow detections. This
148 is done by looking at both the layer average depolarization ratio and color ratio and limiting the
149 top height of the layer to 500 m. If a layer is detected, but the top of the layer is above 500 m, it
150 is not included as blowing snow. This height limit helped screen out diamond dust which often
151 stretches for a few kilometers vertically and frequently reaches the ground. It was found that
152 for most blowing snow layers, the depolarization and color ratio (1064/532) averaged about 0.4
153 and 1.3, respectively (see Fig. 1). If the layer average color or depolarization ratios were out of
154 pre-defined threshold limits, the layer was rejected. The layer average color ratio had to be
155 greater than 1.0 and the depolarization ratio greater than 0.25. The large color ratio is
156 consistent with model simulations for spherical ice particles (Bi et al., 2009). Further, logic was
157 included to reduce misidentification of low cloud as blowing snow by limiting both the
158 magnitude and height of the maximum backscatter signal in the layer. If the maximum signal
159 were greater than 2.0×10^{-1} km^{-1} sr^{-1}, the layer was assumed cloud and not blowing snow. In
160 addition, if the maximum backscatter, regardless of its value, occurs above 300 m, the layer is
161 rejected. These changes to the blowing snow detection algorithm slightly decreased (few
162 percent) the overall frequency of blowing snow detections, but we believe we have reduced the
163 occurrence of false positives and the resulting retrievals are now more accurate.

164 Typically, the blowing snow layers are 100–200 m thick, but can range from the minimum
165 detectable height (20 - 30 m) to over 400 m in depth (Mahesh et al., 2003). Often they are seen
166 to be associated with blowing snow storms that cover vast areas of Antarctica and can persist
167 for days. Blowing snow can occur as frequently as 50 % of the time over large regions of East
168 Antarctica in all months but December–February and as frequently as 75 % April through
169 October (Palm et al., 2011). An example of a typical blowing snow layer as seen from the
170 CALIOP backscatter data is shown in Fig. 1.

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

171 **2.1 Sublimation**

172 Sublimation of snow occurs at the surface but is greatly enhanced when the snow becomes
173 airborne by the action of wind and turbulence. Once snow particles become airborne, their
174 total surface area is exposed to the air. If the relative humidity of the ambient air is less than
175 100 %, then sublimation will occur. The amount of sublimation is dictated by the number of
176 snow particles in suspension and the relative humidity and temperature of the air. Thus, to
177 estimate sublimation of blowing snow, we must be able to derive an estimate of the number
178 density of blowing snow particles and have knowledge of atmospheric temperature and
179 moisture within the blowing snow layer. The only source of the latter, continent wide at least, is
180 from global or regional models or re-analysis fields. The number density of blowing snow
181 particles can be estimated directly from the CALIOP calibrated, attenuated backscatter data if
182 we can estimate the extinction within the blowing snow layer and have a rough idea of the
183 blowing snow particle radius. The extinction can be estimated from the backscatter through an
184 assumed extinction to backscatter ratio (lidar ratio) for the layer. The lidar ratio, though
185 unknown, would theoretically be similar to that of cirrus clouds, which has been extensively
186 studied. Work done by Josset et al., (2012) and Chen et al., (2002) shows that the extinction to
187 backscatter ratio for cirrus clouds typically ranges between 25 and 30 with an average value of
188 29. However, the ice particles that make up blowing snow are more rounded than the ice
189 particles that comprise cirrus clouds and are on average somewhat smaller Walden et al.,
190 (2003). For this paper, we use a value of 25 for the extinction to backscatter ratio.

191 Measurements of blowing snow particle size have been made by a number of investigators
192 [Schmidt, 1982; Mann et al., 2000; Nishimura and Nemoto, 2005; Walden et al. 2003; Lawson et
193 al., 2006; Gordon and Taylor, 2009], but they were generally made within the first few meters
194 of the surface and may not be applicable to blowing snow layers as deep as those studied here.
195 Most observations have shown a height dependence of particle size ranging from 100 to 200
196 μm in the lower tens of centimeters above the surface to 50–60 μm near 10 m height
197 (Nishimura and Nemoto, 2005). A notable exception is the result of Harder et al., (1996) at the
198 South Pole, who measured the size of blowing snow particles during a blizzard by collecting

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199 them on a microscope slide. They report nearly spherical particles with an average effective
200 radius of 15 µm, but the height at which the measurements were made is not reported. From
201 surface observations made at the South Pole, Walden et al., (2003) and Lawson et al., (2006)
202 report an average effective radius for blowing snow particles of 19 and 17 µm, respectively.

203 While no field-measured values for particle radii above roughly 10 m height are available,
204 modeling work indicates that they approach an asymptotic value of about 10-20 µm at heights
205 of 200 m or more (Dery and Yau, 1998). It is also reasonable to assume that snow particles that
206 are high up in the layer are smaller since they have spent more time aloft and have had a
207 greater time to sublimate. Based on the available data, we have defined particle radius (*r(z)*,
208 µm) as a linear function of height:

209 $$r(z) = 40 - \frac{z}{20} \quad (1)$$

210 Thus, for the lowest level of CALIPSO retrieved backscatter (taken to be 15 m – the center of
211 the first bin above the surface), r(15) = 39.25 µm and at the highest level (500 m), r(500) = 15
212 µm.

213 The blowing snow particle number density *N(z)* (particles per cubic meter) is:

214 $$N(z) = \frac{(\beta(z) - \beta_m(z))\, S}{2\pi\, r^2(z)} \quad (2)$$

215 Where $\beta(z)$ is the CALIPSO measured attenuated calibrated backscatter at height *z* (30 m
216 resolution), $\beta_m(z)$ is the molecular backscatter at height *z* and *S* is the extinction to backscatter
217 ratio (25). Here $\beta(z)$ represents the atmospheric backscatter profile through the blowing snow
218 layer. Both $\beta_m(z)$ and $\beta(z)$ have units of m^{-1} sr^{-1}. We found that the values of N(z) obtained
219 from Eq. (2) for the typical blowing snow layer range from about 5.0 x 10^4 to 1.0 x 10^6 particles
220 per cubic meter. This is consistent with the blowing snow model results of Dery and Yau (2002)
221 and the field observations of Mann et al., (2000). A plot of the average particle density for the
222 blowing snow layer in Fig. 1 is shown in Fig. 2.

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

Once an estimate of blowing snow particle number density and radii are obtained, the sublimation rate of the particles can be computed based on the theoretical knowledge of the process. Following Dery and Yau, (2002), the blowing snow mixing ratio q_b (Kg ice / Kg air) is given by:

$$q_b(z) = \frac{4\pi\, \rho_{ice}\, r^3(z)\, N(z)}{3\, \rho_{air}} \quad (3)$$

Or substituting for *N(z)* (Eq. 2):

$$q_b(z) = \frac{2\, \rho_{ice}\, r(z)\, [\beta(z) - \beta_m(z)]\, S}{3\, \rho_{air}} \quad (4)$$

Where ρ_{ice} is the density of ice (917 kg m^{-3}), and ρ_{air} the density of air. Again following Dery and Yau (2002) and others, the sublimation S_b at height *z* is computed from:

$$S_b(z) = \frac{q_b(z)\, N_u\, [q_v(z)/q_{is}(z) - 1]}{2\, \rho_{ice}\, r^2(z)\, [F_k(z) + F_d(z)]} \quad (5)$$

Or, letting $\alpha(z)$ be the extinction and substituting for $q_b(z)$:

$$S_b(z) = \frac{\alpha(z)\, N_u\, [q_v(z)/q_{is}(z) - 1]}{3\, \rho_{ice}\, r(z)\, [F_k(z) + F_d(z)]} \quad (6)$$

Where Nu is the Nusslet number defined as: $Nu = 1.79 + 0.606\, Re^{0.5}$

with the Reynolds number being: $Re = 2r(z)\ v_b/v$

where v_b is the snow particle fall speed (assumed here to be 0.1 ms^{-1}) and v the kinematic viscosity of air (1.512x10^{-5} m^{2}s^{-1}). q_v is the water vapor mixing ratio of the air (obtained from model data), q_{is} is the saturation mixing ratio with respect to ice, and F_k and F_d are the heat conduction and diffusion terms (m s Kg^{-1}):

$$F_k = \left(\frac{L_s}{R_v T} - 1\right) \frac{L_s}{KT}$$

$$F_d = \frac{R_v T}{D\, e_i(T)}$$

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Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

Where L_s is the latent heat of sublimation (2.839x10^6 J/Kg), R_v is the individual gas constant for water vapor (461.5 J Kg^{-1} K^{-1}), T is temperature (K), *K* is the thermal conductivity of air, and *D* the coefficient of diffusion of water vapor in air (both *D* and *K* are functions of temperature (see Rogers and Yau, 1989). S_b has units of Kg Kg^{-1} s^{-1}. This can be interpreted as the mass of snow sublimated per mass of air per second.

Then the column integrated blowing snow sublimation is:

$$Q_s = \rho_{air} \int_{z=0}^{Z_{top}} S_b(z)\, dz \quad (7)$$

Where Z_{top} is the top of the blowing snow layer and *dz* is 30 meters. Q_s has units of Kg m^{-2} s^{-1}). Conversion to mm snow water equivalent (swe) per day is performed by multiplying by a conversion factor:

$$\rho' = 10^3 \; N_s/\rho_{ice} \quad (8)$$

Where N_s is the number of seconds in a day (86,400). The total sublimation amount in mm swe per day is then:

$$Q' = \rho' \, Q_s \quad (9)$$

This computation is performed for every blowing snow detection along the CALIPSO track over Antarctica. A 1 x 1 degree grid is then established over the Antarctic continent and each sublimation calculation (*Q'*) is added to its corresponding grid box over the length of time being considered (i.e. a year or month). This value is then normalized by the total number of CALIPSO observations that occurred for that grid box over the time span. The total number of observations includes all CALIPSO shots within the grid box for which a ground return was detected, regardless of whether blowing snow was detected for that shot or not. Thus, the normalization factor is the total number of shots with ground return detected for that box and is always greater than the number of blowing snow detections (which equals the number of sublimation retrievals). In order for the blowing snow detection algorithm to function, it must first detect the position of the ground return in the backscatter profile. If it cannot do so, it is

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not considered an observation. Over the interior of Antarctica, failure to detect the surface does not occur often as cloudiness is less than 10 % and most clouds are optically thin. Near the coasts, optically thick clouds become more prevalent. This approach will result in higher sublimation values for those grid boxes that contain a lot of blowing snow detections and vice versa (as opposed to just taking the average of the sublimation values for a grid box).

2.2 Transport

The transport of blowing snow is computed using the CALIPSO retrievals of blowing snow mixing ratio and the MERRA-2 winds. A transport value is computed at each 30 m bin level and integrated through the depth of the blowing snow layer:

$$Q_t = \rho_{air} \int_{z=0}^{Z_{top}} q_b(z)\ u(z)\ dz \quad (10)$$

Where $q_b(z)$ is the blowing snow mixing ratio from Eq. (3) and $u(z)$ is the MERRA-2 wind speed at height z and Q_t has units of kg m^{-1} s^{-1}. The wind speed is linearly interpolated from the nearest two model levels. As with the sublimation, these values are gridded and normalized by the total number of observations. The transport values are computed for each month of the year by summing daily values and then multiplying by the number of seconds in the month (resulting units of kg m^{-1}). The monthly values are then summed to obtain a yearly amount. A further conversion is performed to produce units of Gt m^{-1} yr^{-1} by dividing by 10^{12} (1000 kg per metric ton and 10^9 tons per Gt).

3 Results

3.1 Sublimation

Figure 3 shows the average blowing snow frequency and corresponding total annual blowing snow sublimation over Antarctica for the period 2007–2015. The highest values of sublimation are along and slightly inland of the coast. Notice that this is not necessarily where the highest

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Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

291 blowing snow frequencies are located. Sublimation is highly dependent on the air temperature
292 and relative humidity. For a given value of the blowing snow mixing ratio (q_b), the warmer and
293 drier the air, the greater the sublimation. In Antarctica, it is considerably warmer along the
294 coast but one would not necessarily conclude that it is drier there. However, other authors
295 have noted that the katabatic winds, flowing essentially downslope, will warm and dry the air
296 as they descend (Gallee, 1998, and others). We have examined the MERRA-2 relative humidity
297 (with respect to ice) and indeed, according to the model, it is usually drier along the coast. The
298 model data often shows 90 to 100 % (or even higher) relative humidity for interior portions of
299 Antarctica, while along the coast it is often 60 % or less. It should be noted, however, that this
300 model prediction has never been validated through observations. The combination of warmer
301 and drier air makes a big difference in the sublimation as shown in Fig. 4. For a given relative
302 humidity the sublimation can increase by almost a factor of 100 as temperature increases from
303 -50 to -10 °C. For temperatures greater than -20 °C, sublimation is very dependent on relative
304 humidity, but this dependence lessens somewhat at colder temperatures. Continental interior
305 areas with very high blowing snow frequency that approach 75 % (like the Mega Dune region in
306 East Antarctica) exhibit fairly low values of sublimation because it is very cold and the model
307 relative humidity is high.

308 Figure 5 shows the annual total sublimation for years 2007–2015. It is evident that the
309 sublimation pattern or magnitude does not change much from year to year. The overall spatial
310 pattern of sublimation is similar to the model prediction of Dery and Yau, (2002) with our
311 results showing noticeably greater amounts in the Antarctic interior and generally larger values
312 near the coast. As previously noted, most sublimation occurs near the coast due mainly to the
313 warmer temperatures. The areas of sublimation maximums near the coast are consistently in
314 the same location year to year, indicating that these areas may experience more blowing snow
315 episodes and possibly more precipitation (availability of snow to become airborne). It is
316 interesting to compare the sublimation pattern with current estimates of Antarctic
317 precipitation. Precipitation is notoriously difficult to quantify over Antarctica due to the scarcity
318 of observations and strong winds producing drifting and blowing snow which can be
319 misidentified as precipitation. Precipitation is often measured by looking at ice cores or is

320 estimated by models. But perhaps the most complete (non-model) measure of Antarctic
321 precipitation come from the CloudSat mission. Palerme et al., (2014) used CloudSat data to
322 construct a map of Antarctic precipitation over the entire continent (north of 82 S). They
323 showed that along the East Antarctic coast and slightly inland, precipitation ranges from 500 to
324 700 mm swe yr^{-1} and decreases rapidly inland to less than 50 mm yr^{-1} in most areas south of 75
325 S. Their precipitation pattern is in general agreement with the spatial pattern of our
326 sublimation results and the magnitude of our sublimation estimates is in general less than the
327 precipitation amount, with a few exceptions. These occur mostly inland in regions of high
328 blowing snow frequency such as the Megadune region and in the general area of the Lambert
329 glacier. In these regions, our sublimation estimates exceed the CloudSat yearly precipitation
330 estimates. When this occurs, it is likely that either the precipitation estimate is low or the
331 sublimation estimate is too high. Otherwise it would indicate a net negative mass balance for
332 the area unless transport of snow into the region accounted for the difference.

333 Table 1 shows the average sublimation over all grid cells in snow water equivalent and the
334 integrated sublimation amount over the Antarctic continent (north of 82S) for the CALIPSO
335 period in Gt yr^{-1}. Note that the 2006 data include only months June–December (CALIOP began
336 operating in June, 2006) and the 2016 data are only up through October, and do not include the
337 month of February (CALIOP was not operating). To obtain the integrated amount, we take the
338 year average swe (column 1) multiplied by the surface area of Antarctica north of 82S and the
339 density of ice. The average integrated value for the 9 year period 2007–2015 of 419 Gt yr^{-1} is
340 significantly greater than (about twice) values in the literature obtained from model
341 parameterizations (Lenaerts 2012b). Note also that these figures do not include the area
342 poleward of 82S, the southern limit of CALIPSO observations. If included, and the average
343 sublimation rate over this area were just 4 mm swe per year, this would increase the
344 sublimation total by 10 Gt yr^{-1}. Gallee (1998) used a model to compute blowing snow
345 sublimation over Antarctica and obtained a continent-wide average value of 0.087 mm per day.
346 Further he postulated that If this were constant over a year, it would amount to about 32 mm
347 swe and if all of it were deposited in the ocean it would account for 1.3 mm yr^{-1} of sea level rise.
348 The same logic applied to our results would yield a 1.7 mm yr^{-1} seal level rise.

Our results (Table 1) show somewhat of a decreasing trend to the data with the years 2007 and 2008 having the highest sublimation values while 2014 and 2015 have the lowest. If this trend is real, it is not clear what is causing it. The frequency of blowing snow is relatively constant over this period when integrated over the continent (not shown). Other factors that could cause a decrease in sublimation are colder temperatures, increased humidity, or decreased blowing snow particle concentration (which would equate to a decreased average layer integrated backscatter).

Palerme et al., (2014) has shown that the mean snowfall rate over Antarctica (north of 82 S) from August 2006 to April 2011 is 171 mm yr^{-1}. The average yearly snow water equivalent sublimation from Table I is the average sublimation over the continent (also north of 82 S). For the same time period, our computed CALIPSO-based average blowing snow sublimation is about 50 mm yr^{-1}. This means that on average, over one third of the snow that falls over Antarctica is lost to sublimation through the blowing snow process. In comparison surface sublimation (sublimation of snow on the surface) is considered to be relatively small (about a tenth of airborne sublimation) except in summer (Lenearts 2012a, 2012b).

3.2 Transport

Transport of snow via the wind is generally important locally and does not constitute a large part of the ice sheet mass balance in Antarctica. There are areas where the wind scours away all snow that falls producing a net negative mass balance (i.e. blue ice areas), but in general, the snow is simply moved from place to place over most of the continent. At the coastline, however, this is not the case. There, persistent southerly winds can carry airborne snow off the continent. This can be seen very plainly in Fig. 6 which is a MODIS false color (RGB = 2.1, 2.1, .85 µm) image of a large area of blowing snow covering an area about the size of Texas in East Antarctica. We have found this false color technique to be the best way to visualize blowing snow from passive sensors. The one drawback is that sunlight is required. In the figure, blowing snow shows up as a dirty white, the ice/snow surface (in clear areas) is blue and clouds are generally a brighter white. Also shown in Fig. 6 are two CALIPSO tracks (yellow lines) and their

376 associated retrieved blowing snow backscatter (upper and lower images of CALIOP
377 backscatter). Note that the yellow track lines are drawn only where blowing snow was detected
378 by CALIOP and that not all the CALIOP blowing snow detections are shown. The green dots
379 denote the coastline. Plainly seen along the coast near longitude 145–150E is blowing snow
380 being carried off the continent. In this case, topography might have played a role to funnel the
381 wind in those specific areas. Figure 7 shows a zoomed in image of this area with the red lines
382 indicating the approximate position of the coastline. Also note that, as evidenced by the times
383 of the MODIS images, this transport began on or before October 13 at 23:00 UTC and continued
384 for at least 7 hours. This region is very close to the area of maximum sublimation seen in Fig. 3
385 and shown to be quite stable from year to year in Fig. 5. Undoubtedly, this continent to ocean
386 transport also occurs in other coastal areas of Antarctica and most often during the dark winter
387 (when MODIS could not see it).

388 In an attempt to better understand the magnitude of this phenomena, we have computed the
389 amount of snow mass being blown off the continent by computing the transport at 342 points
390 evenly spaced (about 70 km apart) along the Antarctic coast using only the v component of the
391 wind. If the v component is positive, then the wind is from south to north. The transport (Eq.
392 (10) using only the v wind component) is computed at each coastal location and then summed
393 over time at that location. The resulting transport is then summed over each coastal location to
394 arrive at a continent-wide value of transport from continent to ocean. Of course this assumes
395 that the coastline is oriented east-west everywhere. This is true of a large portion of Antarctica
396 but there are regional exceptions. Thus we view the results shown in Table II to be an upper
397 limit of the actual continent to ocean transport. Evident from Table 2 is that most of the
398 transport for East Antarctica occurs in a relatively narrow corridor, with on average over half
399 (51 %) of the transport occurring between 135E and 160E. This is obviously due to the very
400 strong and persistent southerly winds (see Fig. 10) and high blowing snow frequency in this
401 region and is consistent with the conclusions of Scarchilli et al., (2010). In West Antarctica, an
402 even greater fraction (60 %) of the transport off the coast occurs between 80W and 120W.

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

In Fig. 8 we show the magnitude of blowing snow transport for the 2007–2015 timeframe in Mt km^{-1} yr^{-1} as computed from Eq. (10). The magnitude of snow transport, as expected, closely resembles the overall blowing snow frequency pattern as shown in Fig. 3. The maximum values (white areas in Fig. 8) exceed about 3 x 10^6 tons of snow per km per year. In Figs. 9 and 10 we display the MERRA-2 average 10 m wind speed and direction for the years 2007–2015. By inspection of Figs. 8 and 10 it is seen that the overall transport in East Antarctica is generally from south to north and obviously dominated by the katabatic wind regime. It is immediately apparent that the average wind speed and direction does not change much from year to year, with the former helping to explain why the average continent-wide blowing snow frequency is also nearly constant from year to year (not shown).

4. Error Analysis

There are a number of factors that can affect the accuracy of the results presented in this work. These include:

1) Error in the calibrated backscatter and conversion to extinction
2) Errors in the assumed size of blowing snow particles
3) Not correcting for possible attenuation above the blowing snow layer
4) Misidentification of some layers as blowing snow when in fact they were not (false positives)
5) Failure to detect some layers (false negatives)
6) Errors in the MERRA-2 temperature and moisture data
7) Limited spatial sampling

The magnitude of some of these can be estimated, others are hard to quantify. For instance, 1), 2) and 6) are directly involved in the calculation of sublimation (Eq. 6). The error in extinction, particle radius, temperature and moisture can be estimated. The error associated with 3) is probably very small over the interior of Antarctica, but could be appreciable nearer the coastline. In the interior, clouds are a rare occurrence and when present are usually optically thin. Cloudiness increases dramatically near the coast both in terms of frequency and optical depth. Here the effect of overlying attenuating layers could be appreciable in that it would

reduce the backscatter of the blowing snow layer and the derived extinction. This in turn would lead to a lower blowing snow mixing ratio and thus lower sublimation and transport.

There is a further point to be made with respect to clouds that relates to 5) above. The method we use to detect blowing snow will not work in the presence of overlying, fully attenuating clouds. It is reasonable to suspect that cyclonic storms which impinge upon the Antarctic coast and travel some distance inland would be associated with optically thick clouds and contain both precipitating and blowing snow. Our method would not be able to detect blowing snow during these storms, but we would not count such cases as "observations", since the ground would not be detected. The point is, blowing snow probably occurs often in wintertime cyclones, but we are not able to detect it. This could lead to an under prediction of blowing snow occurrence, especially near the coast. Also, blowing snow layers less than 20 - 30 m thick would also likely be missed. It is not clear how often these layers occur, but they are known to exist and missing them will produce an underestimate of blowing snow sublimation and transport amounts. With regard to spatial sampling (7 above), unlike most passive sensors, CALIPSO obtains only point measurements along the spacecraft track at or near nadir. On a given day, sampling is poor. CALIPSO can potentially miss a large portion of blowing snow storms such as is evidenced from inspection of Fig. 6. We have seen many examples of such storms in both the MODIS and CALIPSO record. Quantifying the effect of poor sampling on sublimation estimates would be difficult but should be pursued in future work.

In an effort to quantify the error in our sublimation estimates, here we assume the extinction error to be 20 %, the particle radius error 10 % and the temperature and moisture error 5 % each. In Eq. (6) these terms are multiplicative. The total error in sublimation is then:

$$\pm 1 - (0.8 * 0.9 * 0.95 * 0.95) = \pm 0.35$$

This indicates that the sublimation values derived in this work should be considered to have an error bar of ±35 %. The error in computed transport involves error in wind speed and the blowing snow mixing ratio, the latter being dependent on extinction and particle size. If we

456 assume wind speed has an error of 20 %, extinction 20 % and particle size 10 %, the total error
457 in transport is:

458 ±1 – (0.8 * 0.8 * 0.9) = ±0.42

459 **5. Summary and Discussion**

460 This paper presents the first estimates of blowing snow sublimation and transport over
461 Antarctica that are based on actual observations of blowing snow layers from the CALIOP space
462 borne lidar. We have used the CALIOP blowing snow retrievals combined with MERRA-2 model
463 re-analyses of temperature and moisture to compute the temporal and spatial distribution of
464 blowing snow sublimation and transport over Antarctica for the first time. The results show that
465 the maximum sublimation, with annual values exceeding 250 mm swe, occurs within roughly
466 200 km of the coast even though the maximum frequency of blowing snow most often occurs
467 considerably further inland. This is a result of the warmer and drier air near the coast (at least
468 in the MERRA-2 re-analysis data) which substantially increases the sublimation. In the interior,
469 extremely cold temperatures and high model relative humidity lead to greatly reduced
470 sublimation. However, the values obtained in parts of the interior (notably the Megadune
471 region of East Antarctica - roughly 75 to 82S and 120 to 160E) are considerably higher than
472 prior model estimates of Dery and Yau (2002) or Lenaerts et al., 2012a). This is most likely due
473 to the very high frequency of occurrence of blowing snow as detected from CALIOP data in this
474 region which is not necessarily captured in models (Lenaerts et al., 2012b).

475 The overall spatial pattern of blowing snow sublimation is consistent with previous modelling
476 studies (Dery and Yau, 2002 and Lenearts et al., 2012a). However, we find the Antarctic
477 continent-wide integrated blowing snow sublimation to be about twice the amount of previous
478 studies such as Lenaerts et al., (2012a) (393 ±138 vs roughly 190 Gt yr^{-1}), even though the
479 observations include only the area north of 82° S. The maximum in sublimation is about 250
480 mm swe per year near the coast between longitudes 140E and 150E and seems to occur
481 regularly throughout the 11 year data record. One reason for the higher sublimation values in
482 this study is that the average blowing snow layer depth as determined from the CALIOP

483 measurements is 120 m. Layers as high as 200 - 300 m are not uncommon. It is likely that
484 models such as those cited above do not always capture the full depth of blowing snow layers,
485 thus producing a smaller column-integrated sublimation amount. Another reason for
486 differences between our result and previous model estimates lies in the fact that we only
487 compute sublimation from blowing snow layers that are known to exist (meaning they have
488 been detected from actual backscatter measurements). Models, on the other hand, must infer
489 the presence of blowing snow from pertinent variables within the model. The existence of
490 blowing snow is not easy to predict. It is a complicated function of the properties of the
491 snowpack, surface temperature, relative humidity and wind speed. Snowpack properties
492 include the dendricity, sphericity, grain size and cohesion, all of which can change with the age
493 of the snow. In short, it is very difficult for models to predict exactly when and where blowing
494 snow will occur, much less the depth that blowing snow layer will attain.

495 The spatial pattern of the transport of blowing snow follows closely the pattern of blowing
496 snow frequency. The maximum transport values are about 5 Megatons per km per year and
497 occur in the Megadune region of East Antarctica with other locally high values at various
498 regions near the coast that generally correspond to the maximums in sublimation. We
499 attempted to quantify the amount of snow being blown off the Antarctic continent by
500 computing the transport along the coast using only the v component of the wind. While this
501 may produce an overestimate of the transport (since the Antarctic coast is not oriented east-
502 west everywhere), we find the amount of snow blown off the continent to be significant and
503 fairly constant from year to year. The average off-continent transport for the 9 year period
504 2007–2015 was 3.68 Gt yr^{-1} with about two thirds of that coming from East Antarctica and over
505 one third from a relatively small area between longitudes 135E and 160E.

506 Over the nearly 11 years of data, the inter-annual variability of continent wide sublimation
507 (Table 1) can be fairly large – 10 to 15 % - and likely the result of precipitation variability. There
508 seems to be a weak trend to the sublimation data with earlier years having greater sublimation
509 than more recent years. However, based on the likely magnitude of error in the sublimation
510 estimates, the trend cannot be considered statistically significant.

Data Availability

The CALIPSO calibrated attenuated backscatter data used in this study can be obtained from the NASA Langley Atmospheric Data Center at: https://earthdata.nasa.gov/about/daacs/daac-asdc. CALIPSO Science Team (2015), CALIPSO/CALIOP Level 1B, Lidar Profile Data, version 3.30, Hampton, VA, USA: NASA Atmospheric Science Data Center (ASDC), Accessed at various times during 2016 at doi: 10.5067/CALIOP/CALIPSO/CAL_LID_L1-ValStage1-V3-30_L1B-003.30

The MERRA-2 data are available from the Goddard Earth Sciences Data and Information Services Center (GESDISC) at: https://disc.gsfc.nasa.gov/datareleases/merra_2_data_release.

The blowing snow data (layer backscatter, height, etc.) are available through the corresponding author and will be made publicly available through the NASA Langley Atmospheric Data Center in the near future.

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The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017

640 Table I. The year average sublimation per year (average off all grid boxes) and the integrated
641 sublimation over the Antarctic continent (north of 82S). *2006 and 2016 consist of only 7 and 9
642 months of observations, respectively.

Year	Average Sublimation (mm swe)	Integrated Sublimation (Gt yr^{-1})
2006*	28.3	255
2007	56.8	514
2008	49.2	446
2009	45.3	409
2010	42.9	388
2011	47.6	431
2012	44.4	402
2013	47.7	432
2014	41.5	376
2015	41.3	374
2016*	33.2	301
AVG	**43.5***	**393.4***

643

644 Table II. The total transport (Gt yr^{-1}) from continent to ocean for various regions in Antarctica
645 for 2007–2015.

Year	East Antarctica	West Antarctica	135E – 160E	80W – 120W
2007	2.52	1.29	1.72	0.82
2008	2.20	1.43	1.21	0.90
2009	2.63	1.27	1.51	0.78
2010	2.26	1.15	1.38	0.73
2011	2.04	1.04	1.13	0.64
2012	2.49	1.21	1.41	0.73
2013	2.54	1.41	1.26	0.83
2014	2.55	1.02	1.49	0.67
2015	2.76	1.38	1.58	0.69
Avg	**2.44**	**1.24**	**1.41**	**0.75**

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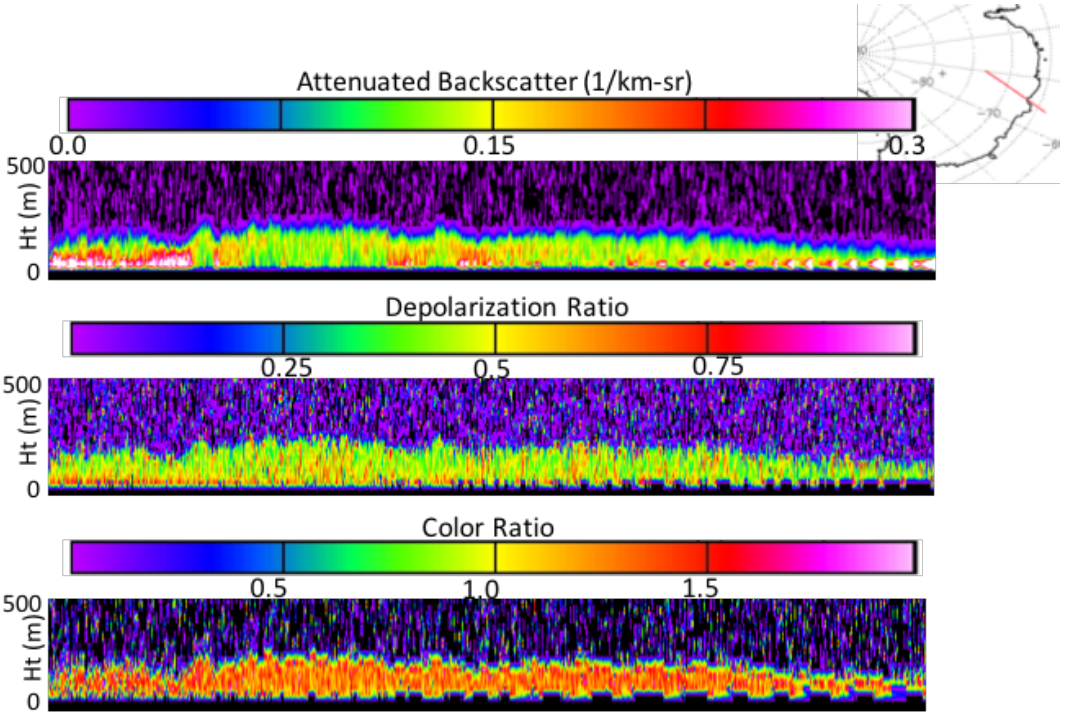
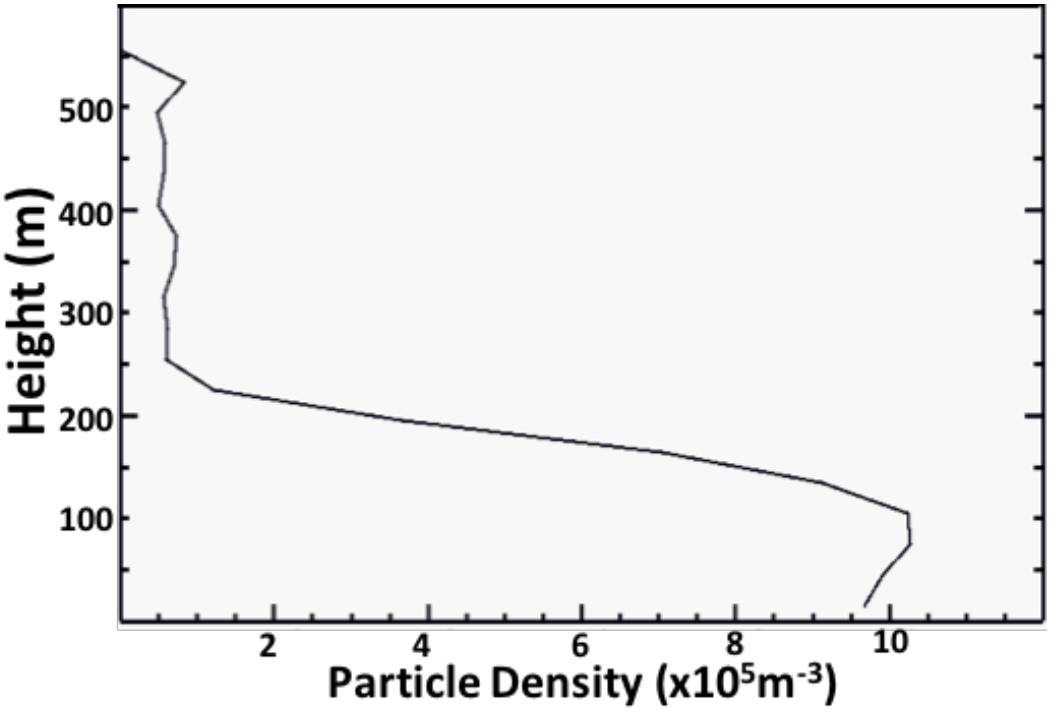
The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



647

648 Figure 1. A typical Antarctic blowing snow layer as measured by CALIPSO on May 28, 2015 at
649 17:08:41 – 17:11:33 UTC. Displayed (from top to bottom) are the 532 nm calibrated, attenuated
650 backscatter, the depolarization ratio at 532 nm, and the color ratio (1064 nm / 532 nm).

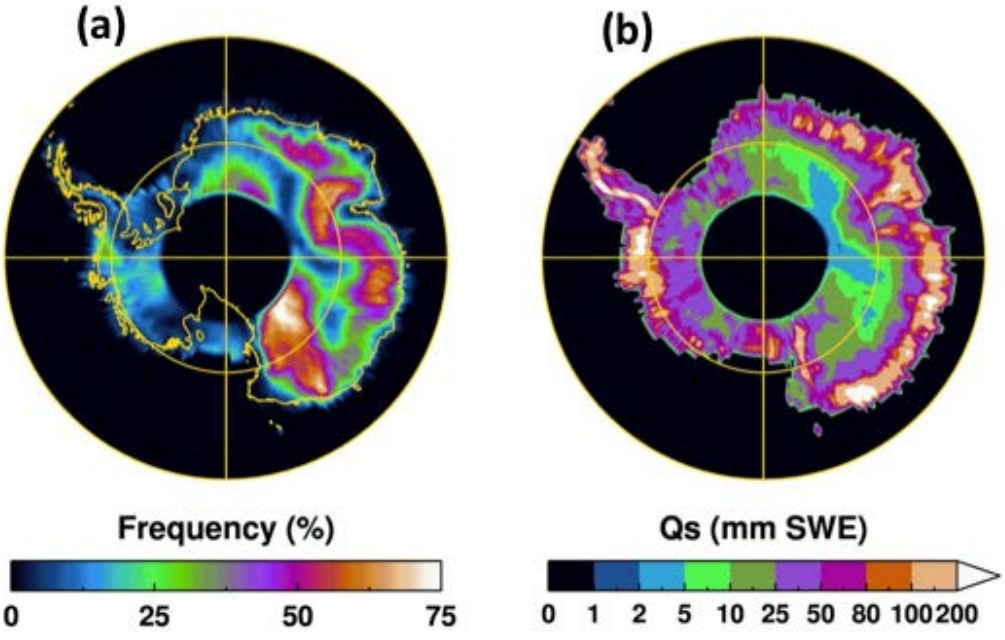
The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



651

652 Figure 2. Average particle density profile (Eq. 2)through the blowing snow layer shown in Fig. 1.

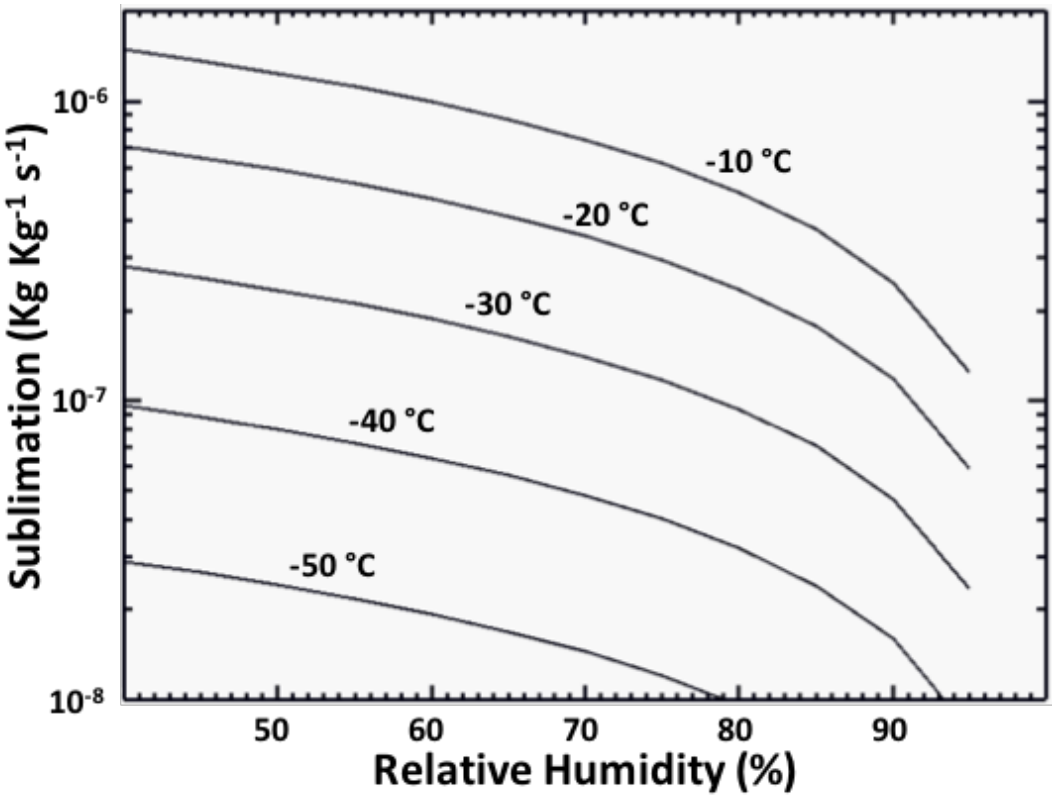
The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



653

654 Figure 3. (a) The average April through October blowing snow frequency for the period 2007–
655 2015. (b) The average annual blowing snow sublimation for the same period as in (a).

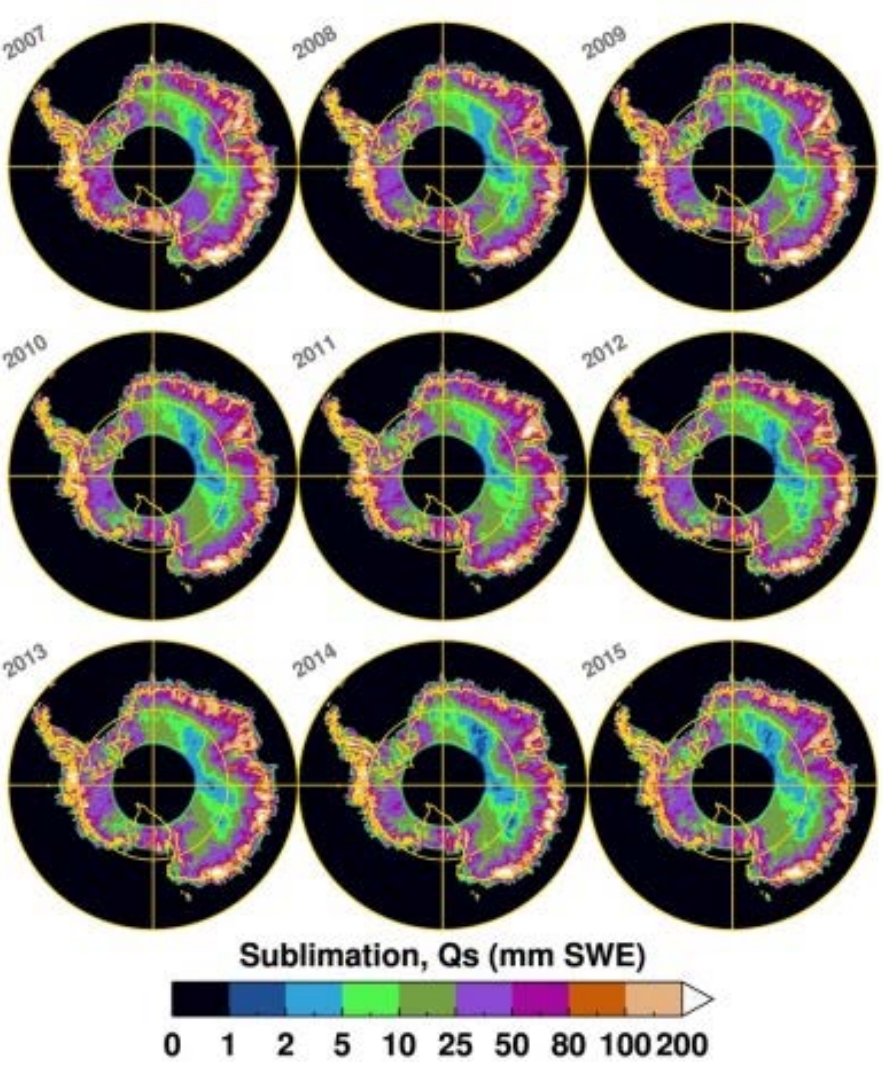
The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



656

657 Figure 4. Computed blowing snow sublimation rate using Eqs. (3) and (4) as function of relative
658 humidity for varying air temperatures. The particle density value used in Eq. (3) was 106 m-3
659 which corresponds to a blowing snow mixing ratio (qb) of 4.7x10-5 Kg Kg-1

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



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661 Figure 5. Blowing snow total sublimation over Antarctica by year for 2007–2015.

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
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Discussion started: 3 April 2017

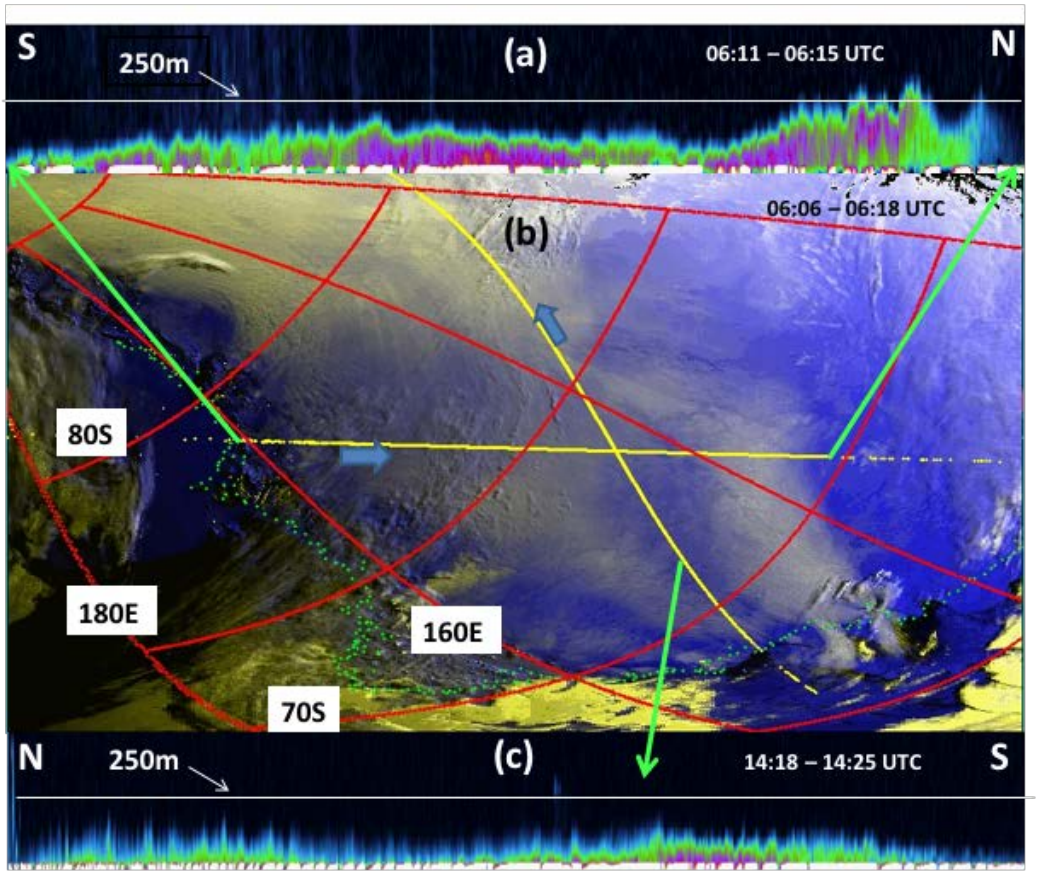

Figure 6. A large blowing snow storm over Antarctica with blowing snow transport from continent to ocean on October 14, 2009. (a) CALIOP 532 nm attenuated backscatter along the yellow (south to north) line bounded by the green arrows as shown in (b) at 06:11 – 06:15 UTC. (b) MODIS false color image at 06:06:14 – 06:17:31 UTC showing blowing snow as dirty white areas. The coastline is indicated by the green dots, and two CALIPSO tracks, where blowing snow was detected are indicated by the yellow lines. (c) CALIOP 532 nm attenuated backscatter along the yellow (north to south) line, 14:18 – 14:25 UTC.

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
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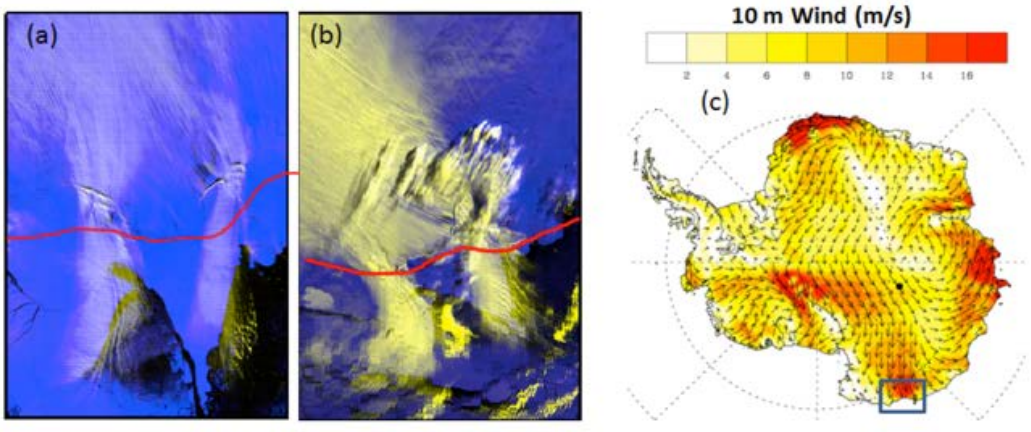
Discussion started: 3 April 2017



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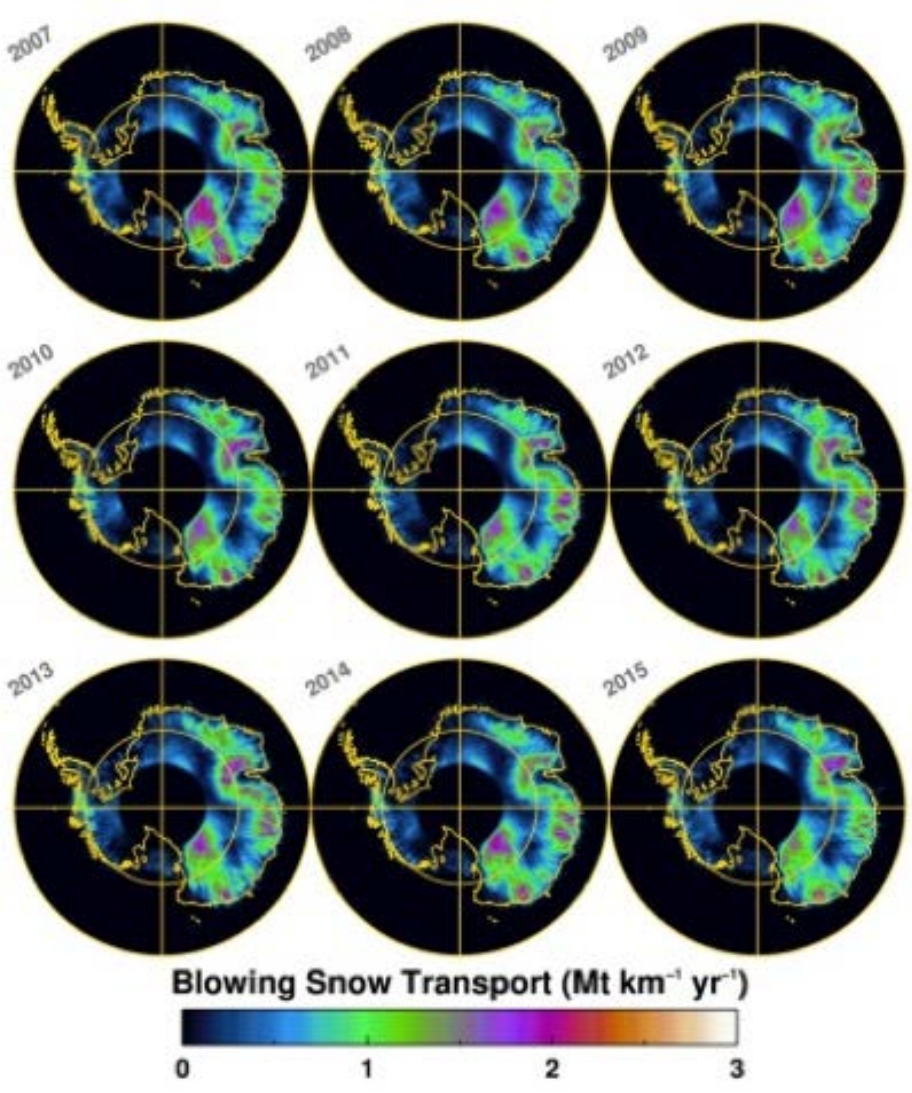
671 Figure 7. (a) MODIS false color image on October 13, 2009, 23:00 UTC and (b) October 14,
672 2009, 06:16 UTC. The red line is the approximate position of the coastline. (c) The 10 m wind
673 speed from the AMPS model (Antarctic Mesoscale Prediction System) for October 14, 2009. The
674 area covered by the MODIS images is roughly that indicated by the blue box in (c).

The Cryosphere Discuss., doi:10.5194/tc-2017-45, 2017
Manuscript under review for journal The Cryosphere
Discussion started: 3 April 2017



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676 Figure 8. The magnitude of blowing snow transport over Antarctica integrated over the year for
677 years 2007–2015.

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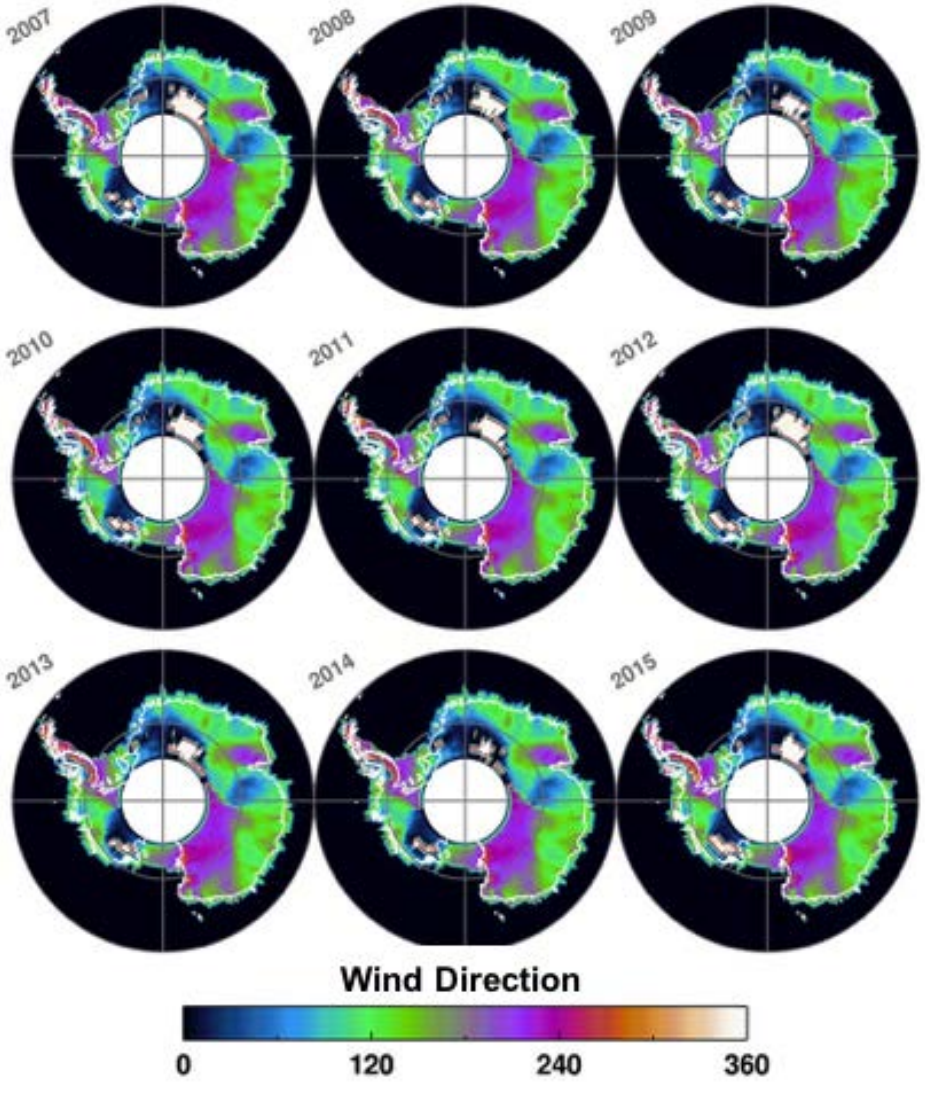
Discussion started: 3 April 2017



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679 Figure 9. The average wind direction over Antarctica for years 2007–2015

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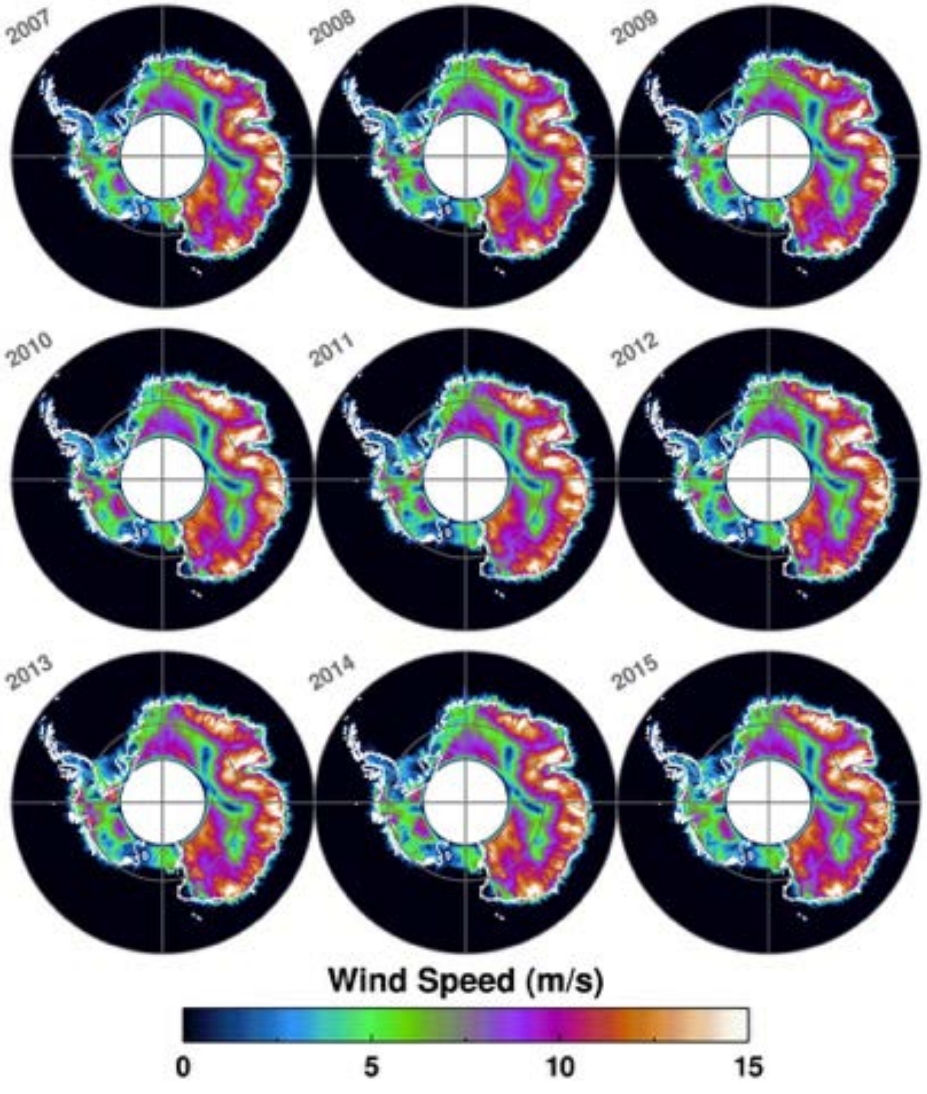

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681 Figure 10. The average wind speed over Antarctica for years 2007–2015