

Editor Decision: Reconsider after major revisions (11 Aug 2017) by Philip Marsh

Comments to the Author:

Dr. Palm. The second round of review comments have been returned to me, with both reviewers thanking you for the efforts made to improve the paper after the first round of reviews. However, both reviewers still have significant concerns that need to be satisfactorily addressed prior to publication. Although this is the second round of reviews, I am willing to allow another round of revisions if you wish. I would suggest that you address all of the reviewers ongoing concerns, including the following.

Reviewer #1:

The main concern is that you need to substantiate the claim that MERRA-2 is too moist/cold. One possible approach is to conduct a more thorough comparison of MERRA-2 RH with observations, and make sure consistent RH parameters are used in the different data sets. In addition, please consider the following as well.

We have added supplemental figures S1-S6 which show additional AWS comparisons with the MERRA-2 data. All data relative humidity data presented in these plots is with respect to ice.

Figure 2: what is the source of the AWS data? What is their name and elevation? Why are those 'randomly'?

The source of the AWS data is noted in the acknowledgement section of the paper. They were obtained from the University of Wisconsin site: <ftp://amrc.ssec.wisc.edu/pub/aws/>.

The elevation and site names are now on the AWS-MERRA-2 comparison supplemental figures

Figure 3 a and 3b require significant improvement to address the following concerns. They basically show a big cloud of data and do not support any of the results. The authors should average the data in time and show e.g. monthly averages in comparison to stations. Also, what is the source of the station data? Are they quality controlled? Are they corrected for RH at low temperatures? Drawing conclusions from a comparison at PE (which is not a really typical 'Antarctic' station, since it is sheltered by topography, and blowing snow does not occur frequently) only does not provide evidence of any Antarctic-wide signal. If the authors want to justify this claim, it will require the inclusion of more weather stations.

We have taken the data shown in old figure 3 (Princess Elisabeth Station) and averaged it by month. This was then compared with the similarly averaged MERRA-2 data and the result is included in the revised paper as supplemental figure S7 and S8 and described in section 2.1 of the revised paper. The source of these data is noted in the acknowledgements. We have not as yet been able to determine how the data

was quality controlled or corrected for low temperature. Other AWS sites have been included for comparison with MERRA-2

Also, even more importantly, does MERRA-2 show RH with respect to ice or to water? I suspect it's with respect to ice, and I know ERA-Interim shows RH with respect to water (even at freezing temperatures). This will largely explain the difference between the two reanalyses. Please check thoroughly and make sure they are consistent. If they are inconsistent, I would argue that the claim that MERRA-2 is too moist is no longer justifiable.

Note that old figure 3 has been removed from the revised paper and what used to be old figure 3b is now supplemental figure S9. Referring to that figure and your comment above, all 3 models are showing relative humidity with respect to water. MERRA-2 reports humidity with respect to water and when we use it in our computations of sublimation, we convert it to relative humidity with respect to ice.

Table I: please include uncertainties for each year.

Done

Line 499: How is 5% justified? The authors should carefully compare MERRA-2 to many AWS stations in Antarctica, and derive an uncertainty/error from that – see above.

We have added surface observation and MERRA-2 comparisons and determined that on average MERRA-2 relative humidity is 7% higher than observations.

Reviewer #2:

The major concern is that procedures to estimates the sublimation rates involve considerable errors and its evaluation is not sufficient to date.

Please address the following concerns:

Once the sublimation is generated in the specific grid, the properties of air, such as the temperature and the humidity, change and then flow into the next downstream grid. Obviously, amount of sublimation decreases along the flow except for the case of a drastic temperature rise. Please confirm whether this process is taken into account in the procedure described in the paper. If it is not, it seems likely that you overestimate sublimation. Could this be the cause of the difference with previous research.

With all due respect, what the reviewer is asking us to do is beyond the scope or intent of this paper. We are not employing a model that has blowing snow physics or trying to model how the air is modified from the sublimation. We understand that sublimation occurring within a closed parcel of air will eventually saturate the layer. However, nature does not close its parcels of air in an air tight box. There are other processes that occur and for which models do not have a good handle on. These processes will act to keep the layer from reaching saturation. Processes such as entrainment of dry and warmer air from above the layer, adiabatic warming in a katabatic flow, and warming of the layer due to trapping of longwave radiation. Keep in mind, that the only observations that show relative humidity reaching saturation during blowing snow events are made at the surface. This paper deals with blowing snow layers on average 120 m thick, and often much higher than that. How do these layers grow? They grow by entraining the air from above which is almost always warmer and dryer.

There is a concern that the procedure to evaluate the total error is no relevant. Probably more careful analysis is needed. At a minimum, estimated errors of +/-50 % should be declared in the abstract as well.

We respectfully disagree with the reviewer's assessment that our analysis of error is not relevant. We have carefully looked at error contributions from every source. We feel we have done a thorough job with that. The 50% error is stated in the abstract.

Thank you ,

Philip Marsh

Line 212: there is a need to better explain why the same algorithm for cirrus clouds is applicable for the larger blowing snow particles.

As we explained in our previous response to the reviewer, cirrus cloud particles are in the same size range or larger than blowing snow particles. Blowing snow particles, especially above 30 m are generally smaller than cirrus cloud particles. Cirrus cloud particle sizes range from 10 to 100 um and even much larger (see Heymsfield et al., 2002 for instance)

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

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11

12 **ABSTRACT**

13 Blowing snow processes commonly occur over the earth's ice sheets when the 10 m wind speed
14 exceeds a threshold value. These processes play a key role in the sublimation and re-
15 distribution of snow thereby influencing the surface mass balance. Prior field studies and
16 modeling results have shown the importance of blowing snow sublimation and transport on the
17 surface mass budget and hydrological cycle of high latitude regions. For the first time, we
18 present continent-wide estimates of blowing snow sublimation and transport over Antarctica
19 for the period 2006 - 2016 based on direct observation of blowing snow events. We use an
20 improved version of the blowing snow detection algorithm developed for previous work that
21 uses atmospheric backscatter measurements obtained from the CALIOP (Cloud-Aerosol Lidar
22 with Orthogonal Polarization) lidar aboard the CALIPSO (Cloud-Aerosol Lidar and Infrared
23 Pathfinder Satellite Observation) satellite. The blowing snow events identified by CALIPSO and
24 meteorological fields from MERRA-2 are used to compute the blowing snow sublimation and
25 transport rates. Our results show that maximum sublimation occurs along and slightly inland of
26 the coastline. This is contrary to the observed maximum blowing snow frequency which occurs
27 over the interior. The associated temperature and moisture re-analysis fields likely contribute
28 to the spatial distribution of the maximum sublimation values. However, the spatial pattern of
29 the sublimation rate over Antarctica is consistent with modeling studies and precipitation

30 estimates. Overall, our results show that the 2006 – 2016 Antarctica average integrated
31 blowing snow sublimation is about $393 \pm 196 \text{ Gt yr}^{-1}$ which is considerably larger than previous
32 model-derived estimates. We find maximum blowing snow transport amount of 5 Megatons
33 $\text{km}^{-1} \text{ yr}^{-1}$ over parts of East Antarctica and estimate that the average snow transport from
34 continent to ocean is about 3.7 Gt yr^{-1} . These continent-wide estimates are the first of their
35 kind and can be used to help model and constrain the surface-mass budget over Antarctica.

36

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

38

39 **1 Introduction**

40 The surface mass balance of the earth's great ice sheets that cover Antarctica and Greenland is
41 one of today's most important topics in climate science. The processes that contribute to the
42 mass balance of a snow or ice-covered surface are precipitation (P), surface evaporation and
43 sublimation (E), surface melt and runoff (M), blowing snow sublimation (Q_s) and snow transport
44 (Q_t). Sublimation of snow can occur at the surface but is greatly enhanced within the
45 atmospheric column of the blowing snow layer. The contributions of these processes to the
46 mass balance vary greatly spatially, and can be highly localized and very difficult to quantify.

$$47 \quad S = \int_{\text{year}} (P - E - M - Q_t - Q_s) dt \quad (1)$$

48 It is well known that the Arctic is experiencing rapid warming and loss of sea ice cover and
49 thickness. In the past few decades, the Arctic has seen an increase in average surface air
50 temperature by $2 \text{ }^\circ\text{C}$ (Przybylak, 2007). Modeling studies suggests an increase in annual mean
51 temperatures over the Arctic by $8.5 \pm 4.1 \text{ }^\circ\text{C}$ over the current century that could lead to a
52 decrease in sea ice cover by $49 \pm 18 \%$ (Bintanja and Kriken, 2016). While the Antarctic has
53 experienced an increase in average surface temperature, most of the warming is observed over
54 West Antarctica at a rate of $0.17 \text{ }^\circ\text{C}$ per decade from 1957 to 2006 (Steig et al., 2009; Bromwich
55 et al., 2013). Such surface warming undoubtedly has implications for ice sheet mass balance
56 and sea level rise mainly through the melting term of the mass balance equation. However, the
57 other processes affecting the mass balance of ice sheets may also be experiencing changes that

58 are difficult to identify and quantify. For instance, models have shown that in a warming
59 climate, precipitation should increase over Antarctica and most of it will fall as snow (Church et
60 al., 2013). If snowfall is increasing, perhaps the frequency of blowing snow and subsequently
61 the magnitude of transport and sublimation will increase as well. Thus, understanding how
62 these processes affect the overall mass balance of the ice sheets and how they may be
63 responding to a changing climate is of growing concern.

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

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

82 For the current work, we focus on blowing snow processes over the Antarctic region. Due to the
83 uninhabited expanse of Antarctica and the lack of observations, prior, continent-wide studies of
84 blowing snow sublimation over Antarctica had to rely on parameterized methods that use
85 model re-analysis of wind speed and low level moisture. The presence of blowing snow is
86 inferred from surface temperature, wind speed and snow age (if known). In a series of papers

87 on the modeling of blowing snow, Dery and Yau (1998, 1999, 2001) develop and test a
88 parameterization of blowing snow sublimation. Dery and Yau (2002) utilize the model with the
89 ECMWF re-analysis covering 1979 to 1993 and show that most blowing snow sublimation
90 occurs along the coasts and over sea ice with maximums in some coastal areas of 150 mm snow
91 water equivalent (swe) yr^{-1} . Lenearts et al., (2012a) utilized a high resolution regional climate
92 model (RACMO2) to simulate the surface mass balance of the Antarctic ice sheet. They found
93 drifting and blowing snow sublimation to be the most significant ablation term reaching values
94 as high as 200 mm yr^{-1} swe along the coast. Average monthly rates of blowing snow sublimation
95 calculated for Halley Station, Antarctica for the years 1995 and 1996, varied between 0.04
96 (winter) to 0.44 (summer) mm day^{-1} (14.6 and 160 mm yr^{-1} respectively) (King et al. 2001).
97 There has been some recent work done on blowing snow sublimation and transport from field
98 measurements (see for instance Barral et al., 2014 and Trouvilliez et. al., 2014) , but the data are
99 sparse and measurements are only available within the surface layer (< 10 m).

100 While transport of blowing snow is considered to be less important than sublimation in terms
101 of mass balance of the Antarctic ice sheet, erosion and transport of snow by wind can be
102 considerable in certain regions. Das et al., (2013) have shown that blue ice areas are frequently
103 seen in Antarctica. These regions exhibit a negative mass balance as all precipitation that falls is
104 either blown off or sublimated away. Along the coastal regions it has been argued that
105 considerable mass is transported off the coast via blowing snow in preferential areas dictated
106 by topography (Scarchilli et al., 2010). In the Tera Nova Bay region of East Antarctica, manned
107 surface observations show that drifting and blowing snow occurred 80 % of the time in fall and
108 winter and cumulative snow transport was 4 orders about of magnitude higher than snow
109 precipitation. Much of this airborne snow is transported off the continent producing areas of
110 blue ice. Such observations raise questions as to how often and to what magnitude continent to
111 ocean transport occurs. This is important, particularly for Antarctica where the coastline
112 stretches over 17,000 km in length (<https://en.wikipedia.org/wiki/Antarctica>) and where
113 prevailing strong winds occur through most of the year. Due to the sparsity of observations, the
114 only way to estimate the mass of snow being blown off the coast of Antarctica is by using

115 model parameterizations. Now, for the first time, satellite observations of blowing snow can
116 help better ascertain the magnitude of this elusive quantity.

117 Considering that the accuracy of model data is questionable over Antarctica, and the
118 complicated factors that govern the onset of blowing snow, it is difficult to assess the accuracy
119 of the parameterization of blowing snow sublimation and transport. Recently, methods have
120 been developed to detect the occurrence of blowing snow from direct satellite observations.
121 Palm et al., (2011) show that blowing snow is widespread over much of Antarctica and, in all
122 but the summer months, occurs over 50 % of the time over large areas of East Antarctica. In
123 this paper, we present a technique that uses direct measurements of blowing snow from the
124 CALIPSO satellite lidar combined with The Modern-Era Retrospective analysis for Research and
125 Applications, Version 2 (MERRA-2) re-analysis fields of moisture, temperature and wind to
126 quantify the magnitude of sublimation and mass transport occurring over most of Antarctica
127 (north of 82 south). Section 2 discusses the method used to compute blowing snow sublimation
128 from CALIPSO and MERRA-2 data. In Sect. 3 we show results and compare with previous
129 estimates of sublimation. In Sect. 4 we examine sources of error, their approximate magnitudes
130 and perform a study on the sensitivity of the calculated sublimation to error in the estimated
131 relative humidity of the layer. Summary and discussion follow in Sect. 5.

132 **2 Method**

133 The method developed for detection of blowing snow using satellite lidar data (both ICESat and
134 CALIPSO) was presented in Palm et al., (2011). That work showed examples of blowing snow
135 layers as seen by the calibrated, attenuated backscatter data measured by the CALIOP
136 instrument on the CALIPSO satellite. CALIOP (Cloud-Aerosol Lidar with Orthogonal Polarization)
137 is a two wavelength (532 and 1064 nm) backscatter lidar with depolarization at 532 nm and has
138 been operating continuously since June of 2006 (Winker et. al., 2009). In the lower 5 km of the
139 atmosphere, the vertical resolution of the CALIOP backscatter profile is 30 m. The CALIOP
140 backscatter profiles are produced at 20 Hz, which is about a horizontal resolution of 330 m
141 along track. The relatively strong backscattering produced by the earth's surface is used to
142 identify the ground bin in each profile. After the ground signal is detected, each 20 Hz profile is

143 examined for an elevated backscatter signal (above a pre-defined threshold) in the first bin
144 above the ground. If found and the surface wind speed is greater than 4 m s^{-1} , successive bins
145 above that are searched for a 80 % decrease in signal value, which is then the top of the layer.
146 Limited by the vertical resolution of the signal, our approach has the ability to identify blowing
147 snow layers that are roughly 20-30 m or more in thickness. Thus, drifting snow which is
148 confined to 10 m or less and occurs frequently over Antarctica would not be reliably detected.
149 The signal from these layers is likely inseparable from the strong ground return. More
150 information on the blowing snow detection algorithm can be found in Palm et al., (2011).

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

169 Typically, the blowing snow layers are 100–200 m thick, but can range from the minimum
170 detectable height (20 - 30 m) to over 400 m in depth (Mahesh et al., 2003). Often they are seen

171 to be associated with blowing snow storms that cover vast areas of Antarctica and can persist
172 for days. Blowing snow can occur as frequently as 50 % of the time over large regions of East
173 Antarctica in all months but December–February and as frequently as 75 % April through
174 October (Palm et al., 2011). An example of a typical blowing snow layer as seen from the
175 CALIOP backscatter data is shown in Fig. 1.

176 **2.1 MERRA-2 Reanalysis Data**

177 In order to compute blowing snow sublimation, the temperature and relative humidity of the
178 layer must be known. Here we use the MERRA-2 reanalysis (Gelaro, 2017). MERRA-2 is
179 produced with version 5.12.4 of the GEOS atmospheric data assimilation system and contains
180 72 vertical levels from the surface to 0.01 hPa on an approximately 0.5° x 0.625° global grid. The
181 re-analyses are available every 3 hours. To obtain the temperature and relative humidity at a
182 given location, height and time, we use the data from the MERRA-2 grid box which is closest in
183 space and time to the observation. Then we linearly interpolate the temperature, moisture and
184 wind to the height of the CALIPSO observation.

185 ~~We understand that~~ MERRA-2 does not include the effects of blowing snow sublimation on
186 atmospheric moisture and thus may have a dry (and possibly warm) bias. MERRA-2
187 temperature and moisture has not been evaluated over Antarctica but in this section we
188 present a comparison of MERRA-2 temperature and moisture at 2 m height with a manned
189 surface station (princess Elisabeth Station) and ~~two six~~ AWS sites. In supplemental Figs. S1 – S6
190 ~~2~~ are data from ~~two the~~ AWS sites (~~chosen at random~~) comparing MERRA-2 and AWS 2 m
191 temperature and relative humidity with respect to ice (RH_{ice}). In ~~both all but one~~ cases MERRA-
192 2 is, on average, slightly colder ~~and moister~~ than the observations (about ~~31~~ °C ~~and 2%~~,
193 respectively). For all six comparisons, the average MERRA-2 moisture is greater than the AWS
194 observation (roughly 7% higher).

195 Supplemental Figs. ~~3a~~ S7 and ~~3b~~ S8 show MERRA-2 data compared to the surface station at
196 Princess Elisabeth (PE) for data taken over 2009-2015. Princess Elisabeth Station is located in
197 East Antarctica at 71.95S, 23.35E at an elevation of 1322m. The PE surface observations are
198 made year round at 3 hourly intervals. MERRA-2 data is then extracted at the time closest to

199 the PE observation. Both the MERRA-2 and the PE data are then averaged over the month. The
200 result shown in Figs. S7 and S8 indicate that MERRA-2 is consistently colder and moister than
201 the observations. Here MERRA-2 is considerably colder and moister (about 6.14 °C and 6-8.4%,
202 respectively). Note also from Fig. S8 that MERRA-2 is much colder than the observations in
203 winter. It is somewhat closer to observations in the summer. The bias shown in Figs. S1 - S7 is
204 calculated as the average of the MERRA-2 data minus the average of the station data. Also
205 shown in Figs. ~~3e, d and e~~ S9 are the annual mean relative humidity at 2 meters above the
206 surface over Antarctica in 2015 estimated by MERRA-2, ERA-Interim, and AMPS-Polar WRF
207 showing that MERRA-2 is considerably moister than ERA-Interim or AMPS. Note that the model
208 humidity fields shown in Fig. S9 are with respect to water.

209 From these comparisons it is likely that MERRA-2 does not exhibit a dry or warm bias and is
210 rather slightly cold and moist compared to surface observations and other models.

211 **2.2 Sublimation**

212 Sublimation of snow occurs at the surface but is greatly enhanced when the snow becomes
213 airborne by the action of wind and turbulence. Once snow particles become airborne, their
214 total surface area is exposed to the air. If the relative humidity of the ambient air is less than
215 100 %, then sublimation will occur. The amount of sublimation is dictated by the number of
216 snow particles in suspension and the relative humidity and temperature of the air. Thus, to
217 estimate sublimation of blowing snow, we must be able to derive an estimate of the number
218 density of blowing snow particles and have knowledge of atmospheric temperature and
219 moisture within the blowing snow layer. The only source of the latter, continent wide at least, is
220 from global or regional models or re-analysis fields. The number density of blowing snow
221 particles can be estimated directly from the CALIOP calibrated, attenuated backscatter data if
222 we can estimate the extinction within the blowing snow layer and have a rough idea of the
223 blowing snow particle radius. The extinction can be estimated from the backscatter through an
224 assumed extinction to backscatter ratio (lidar ratio) for the layer. The lidar ratio, though
225 unknown, would theoretically be similar to that of cirrus clouds, which has been extensively
226 studied. Work done by Josset et al., (2012) and Chen et al., (2002) shows that the extinction to

227 backscatter ratio for cirrus clouds typically ranges between 25 and 30 with an average value of
228 29. However, the ice particles that make up blowing snow are more rounded than the ice
229 particles that comprise cirrus clouds and are on average somewhat smaller (Walden et al.,
230 2003). For this paper, we use a value of 25 for the extinction to backscatter ratio.

231 Measurements of blowing snow particle size have been made by a number of investigators
232 [Schmidt, 1982; Mann et al., 2000; Nishimura and Nemoto, 2005; Walden et al. 2003; Lawson et
233 al., 2006; Gordon and Taylor, 2009], but they were generally made within the first few meters
234 of the surface and may not be applicable to blowing snow layers as deep as those studied here.
235 Most observations have shown a height dependence of particle size ranging from 100 to 200
236 μm in the lower tens of centimeters above the surface to 50–60 μm near 10 m height
237 (Nishimura and Nemoto, 2005). A notable exception is the result of Harder et al., (1996) at the
238 South Pole, who measured the size of blowing snow particles during a blizzard by collecting
239 them on a microscope slide. They report nearly spherical particles with an average effective
240 radius of 15 μm , but the height at which the measurements were made is not reported. From
241 surface observations made at the South Pole, Walden et al., (2003) and Lawson et al., (2006)
242 report an average effective radius for blowing snow particles of 19 and 17 μm , respectively.

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

$$249 \quad r(z) = 40 - \frac{z}{20} \quad (1)$$

250 Thus, for the lowest level of CALIPSO retrieved backscatter (taken to be 15 m – the center of
251 the first bin above the surface), $r(15) = 39.25 \mu\text{m}$ and at the highest level (500 m), $r(500) = 15$
252 μm .

253 The blowing snow particle number density $N(z)$ (particles per cubic meter) can be estimated
 254 from the extinction. Note that the extinction is the numerator in equation 2:

$$255 \quad N(z) = \frac{(\beta(z) - \beta_m(z)) S}{2\pi r^2(z)} \quad (2)$$

256 Where $\beta(z)$ is the CALIPSO measured attenuated calibrated backscatter at height z (30 m
 257 resolution), $\beta_m(z)$ is the molecular backscatter at height z and S is the extinction to backscatter
 258 ratio (25). Here $\beta(z)$ represents the atmospheric backscatter profile through the blowing snow
 259 layer. Both $\beta_m(z)$ and $\beta(z)$ have units of $\text{m}^{-1} \text{sr}^{-1}$. We found that the values of $N(z)$ obtained
 260 from Eq. (2) for the typical blowing snow layer range from about 5.0×10^4 to 1.0×10^6 particles
 261 per cubic meter. This is consistent with the blowing snow model results of Dery and Yau (2002)
 262 and the field observations of Mann et al., (2000). A plot of the average particle density for the
 263 blowing snow layer in Fig. 1 is shown in Fig. 24. Note that the decrease in particle number
 264 density below about 75 m is most likely due to attenuation of the lidar signal as it propagates
 265 through the layer. We did not attempt to correct for this and the overall effect is an under
 266 estimation of the particle density in this region (which would lead to lower calculated blowing
 267 snow sublimation).

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

$$272 \quad q_b(z) = \frac{4\pi \rho_{ice} r^3(z) N(z)}{3 \rho_{air}} \quad (3)$$

273 Or substituting for $N(z)$ (Eq. 2):

$$274 \quad q_b(z) = \frac{2 \rho_{ice} r(z) [\beta(z) - \beta_m(z)] S}{3 \rho_{air}} \quad (4)$$

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

277
$$S_b(z) = \frac{q_b(z) Nu [q_v(z)/q_{is}(z) - 1]}{2 \rho_{ice} r^2(z) [F_k(z) + F_d(z)]} \quad (5)$$

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

279
$$S_b(z) = \frac{\alpha(z) Nu [q_v(z)/q_{is}(z) - 1]}{3 \rho_{ice} r(z) [F_k(z) + F_d(z)]} \quad (6)$$

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

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

282 where v_b is the snow particle fall speed (assumed here to be 0.1 ms^{-1}) and v the kinematic
 283 viscosity of air ($1.512 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$). q_v is the water vapor mixing ratio of the air (obtained from
 284 model data), q_{is} is the saturation mixing ratio with respect to ice, and F_k and F_d are the heat
 285 conduction and diffusion terms (m s kg^{-1}):

286
$$F_k = \left(\frac{L_s}{R_v T} - 1 \right) \frac{L_s}{KT} \quad (7)$$

287

288
$$F_d = \frac{R_v T}{D e_i(T)} \quad (8)$$

289 Where L_s is the latent heat of sublimation ($2.839 \times 10^6 \text{ J/Kg}$), R_v is the individual gas constant for
 290 water vapor ($461.5 \text{ J kg}^{-1} \text{ K}^{-1}$), T is temperature (K), K is the thermal conductivity of air, and D the
 291 coefficient of diffusion of water vapor in air (both D and K are functions of temperature (see
 292 Rogers and Yau, 1989). S_b has units of $\text{kg kg}^{-1} \text{ s}^{-1}$. This can be interpreted as the mass of snow
 293 sublimated per mass of air per second.

294 Then the column integrated blowing snow sublimation is:

295
$$Q_s = \rho_{air} \int_{z=0}^{z_{top}} S_b(z) dz \quad (9)$$

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

$$299 \quad \rho' = 10^3 N_s / \rho_{ice} (1) \quad (10)$$

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

$$302 \quad Q' = \rho' Q_s \quad (11)$$

303 This computation is performed for every blowing snow detection along the CALIPSO track over
304 Antarctica. A 1 x 1 degree grid is then established over the Antarctic continent and each
305 sublimation calculation (Q') is added to its corresponding grid box over the length of time being
306 considered (i.e. a year or month). This value is then normalized by the total number of CALIPSO
307 observations that occurred for that grid box over the time span. The total number of
308 observations includes all CALIPSO shots within the grid box for which a ground return was
309 detected, regardless of whether blowing snow was detected for that shot or not. Thus, the
310 normalization factor is the total number of shots with ground return detected for that box and
311 is always greater than the number of blowing snow detections (which equals the number of
312 sublimation retrievals). In order for the blowing snow detection algorithm to function, it must
313 first detect the position of the ground return in the backscatter profile. If it cannot do so, it is
314 not considered an observation. Over the interior of Antarctica, failure to detect the surface
315 does not occur often as cloudiness is less than 10 % and most clouds are optically thin. Near the
316 coasts, optically thick clouds become more prevalent. This approach will result in higher
317 sublimation values for those grid boxes that contain a lot of blowing snow detections and vice
318 versa (as opposed to just taking the average of the sublimation values for a grid box).

319 **2.3 Transport**

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

$$323 \quad Q_t = \rho_{air} \int_{z=0}^{z_{top}} q_b(z) u(z) dz \quad (12)$$

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

332 **3 Results**

333 **3.1 Sublimation**

334 | Fig. 35 shows the average blowing snow frequency and corresponding total annual blowing
335 snow sublimation over Antarctica for the period 2007–2015. The highest values of sublimation
336 are along and slightly inland of the coast. Notice that this is not necessarily where the highest
337 blowing snow frequencies are located. Sublimation is highly dependent on the air temperature
338 and relative humidity. For a given value of the blowing snow mixing ratio (q_b), the warmer and
339 drier the air, the greater the sublimation. In Antarctica, it is considerably warmer along the
340 coast but one would not necessarily conclude that it is drier there. However, other authors
341 have noted that the katabatic winds, flowing essentially downslope, will warm and dry the air
342 as they descend (Gallee, 1998, and others). We have examined the MERRA-2 relative humidity
343 (with respect to ice) and indeed, according to the model, it is usually drier along the coast. The
344 model data often shows 90 to 100 % (or even higher) relative humidity for interior portions of
345 Antarctica, while along the coast it is often 70 % or less. It should be noted, however, that this

346 model prediction has never been validated through observations. The combination of warmer
347 and drier air makes a big difference in the sublimation as shown in Fig. 46. For a given relative
348 humidity the sublimation can increase by almost a factor of 100 as temperature increases from
349 -50 to -10 °C. For temperatures greater than -20 °C, sublimation is very dependent on relative
350 humidity, but this dependence lessens somewhat at colder temperatures. Continental interior
351 areas with very high blowing snow frequency that approach 75 % (like the Mega Dune region in
352 East Antarctica) exhibit fairly low values of sublimation because it is very cold and the model
353 relative humidity is high.

354 Fig. 57 shows the annual total sublimation for years 2007–2015. It is evident that the
355 sublimation pattern or magnitude does not change much from year to year. The overall spatial
356 pattern of sublimation is similar to the model prediction of Dery and Yau, (2002) with our
357 results showing noticeably greater amounts in the Antarctic interior and generally larger values
358 near the coast. As previously noted, most sublimation occurs near the coast due mainly to the
359 warmer temperatures. The areas of sublimation maximums near the coast are consistently in
360 the same location year to year, indicating that these areas may experience more blowing snow
361 episodes and possibly more precipitation (availability of snow to become airborne). It is
362 interesting to compare the sublimation pattern with current estimates of Antarctic
363 precipitation. Precipitation is notoriously difficult to quantify over Antarctica due to the scarcity
364 of observations and strong winds producing drifting and blowing snow which can be
365 misidentified as precipitation. Precipitation is often measured by looking at ice cores or is
366 estimated by models. But perhaps the most complete (non-model) measure of Antarctic
367 precipitation come from the CloudSat mission. Palerme et al., (2014) used CloudSat data to
368 construct a map of Antarctic precipitation over the entire continent (north of 82 S). They
369 showed that along the East Antarctic coast and slightly inland, precipitation ranges from 500 to
370 700 mm swe yr⁻¹ and decreases rapidly inland to less than 50 mm yr⁻¹ in most areas south of 75
371 S. Their precipitation pattern is in general agreement with the spatial pattern of our
372 sublimation results and the magnitude of our sublimation estimates is in general less than the
373 precipitation amount, with a few exceptions. These occur mostly inland in regions of high
374 blowing snow frequency such as the Megadune region and in the general area of the Lambert

375 glacier. In these regions, our sublimation estimates exceed the CloudSat yearly precipitation
376 estimates. When this occurs, it is likely that either the precipitation estimate is low or the
377 sublimation estimate is too high. Otherwise it would indicate a net negative mass balance for
378 the area unless transport of snow into the region accounted for the difference.

379 Table 1 shows the average sublimation over all grid cells in snow water equivalent and the
380 integrated sublimation amount over the Antarctic continent (north of 82S) for the CALIPSO
381 period in Gt yr^{-1} . Note that the 2006 data include only months June–December (CALIOP began
382 operating in June, 2006) and the 2016 data are only up through October, and do not include the
383 month of February (CALIOP was not operating). To obtain the integrated amount, we take the
384 year average swe (column 1) multiplied by the surface area of Antarctica north of 82S and the
385 density of ice. The average integrated value for the 9 year period 2007–2015 of 393 Gt yr^{-1} is
386 significantly greater than (about twice) values in the literature obtained from model
387 parameterizations (Lenaerts 2012b). Note also that this amount does not include the area
388 poleward of 82S, the southern limit of CALIPSO observations. If included, and the average
389 sublimation rate over this area were just 4 mm swe per year, this would increase the
390 sublimation total by 10 Gt yr^{-1} . Palerme et al., (2014) has shown that the mean snowfall rate
391 over Antarctica (north of 82 S) from August 2006 to April 2011 is 171 mm yr^{-1} . The average
392 yearly snow water equivalent sublimation from Table I is the average sublimation over the
393 continent (and grounded ice shelves) north of 82 S. For the same time period, our computed
394 CALIPSO-based average blowing snow sublimation is about 50 mm yr^{-1} . This means that on
395 average, over one third of the snow that falls over Antarctica is lost to sublimation through the
396 blowing snow process. In comparison surface sublimation (sublimation of snow on the surface)
397 is considered to be relatively small (about a tenth of airborne sublimation) except in summer
398 (Lenaerts 2012a, 2012b).

399 **3.2 Transport**

400 Transport of snow via the wind is generally important locally and does not constitute a large
401 part of the ice sheet mass balance in Antarctica. There are areas where the wind scours away all

402 snow that falls producing a net negative mass balance (i.e. blue ice areas), but in general, the
403 snow is simply moved from place to place over most of the continent. At the coastline,
404 however, this is not the case. There, persistent southerly winds can carry airborne snow off the
405 continent. This can be seen very plainly in Fig. 68 which is a MODIS false color (RGB = 2.1, 2.1,
406 .85 μm) image of a large area of blowing snow covering an area about the size of Texas (16,662
407 km^2) in East Antarctica. We have found this false color technique to be the best way to visualize
408 blowing snow from passive sensors. The one drawback is that sunlight is required. In Fig. 68,
409 blowing snow shows up as a dirty white, the ice/snow surface (in clear areas) is blue and clouds
410 are generally a brighter white. Also shown in Fig. 68 are two CALIPSO tracks (yellow lines) and
411 their associated retrieved blowing snow backscatter (upper and lower images of CALIOP
412 backscatter). Note that the yellow track lines are drawn only where blowing snow was detected
413 by CALIOP and that not all the CALIOP blowing snow detections are shown. The green dots
414 denote the coastline. Plainly seen along the coast near longitude 145–150E is blowing snow
415 being carried off the continent. In this case, topography might have played a role to funnel the
416 wind in those specific areas. Fig. 79 shows a zoomed in image of this area with the red lines
417 indicating the approximate position of the coastline. Also note that, as evidenced by the times
418 of the MODIS images, this transport began on or before October 13 at 23:00 UTC and continued
419 for at least 7 hours. This region is very close to the area of maximum sublimation seen in Fig. 35
420 and shown to be quite stable from year to year in Fig. 57. Undoubtedly, this continent to ocean
421 transport also occurs in other coastal areas of Antarctica and most often during the dark winter
422 (when MODIS could not see it).

423 In an attempt to better understand the magnitude of this phenomena, we have computed the
424 amount of snow mass being blown off the continent by computing the transport at 342 points
425 evenly spaced (about 60 km apart) along the Antarctic coast using only the v component of the
426 wind. If the v component is positive, then the wind is from south to north. The transport (Eq.
427 (12) using only the v wind component) is computed at each coastal location and then summed
428 over time at that location. The resulting transport is then summed over each coastal location to
429 arrive at a continent-wide value of transport from continent to ocean. Of course this assumes
430 that the coastline is oriented east-west everywhere. This is true of a large portion of Antarctica

431 but there are regional exceptions. Thus we view the results shown in Table II to be an upper
432 limit of the actual continent to ocean transport. Evident from Table II is that most of the
433 transport for East Antarctica occurs in a relatively narrow corridor, with on average over half
434 (51 %) of the transport occurring between 135E and 160E. This is obviously due to the very
435 strong and persistent southerly winds (see Fig. S10 and S11) and high blowing snow frequency
436 in this region and is consistent with the conclusions of Scarchilli et al., (2010). In West
437 Antarctica, an even greater fraction (60 %) of the transport off the coast occurs between 80W
438 and 120W.

439 In Fig. [810](#) we show the magnitude of blowing snow transport for the 2007–2015 timeframe in
440 $\text{Mt km}^{-1} \text{yr}^{-1}$ as computed from Eq. (12). The magnitude of snow transport, as expected, closely
441 resembles the overall blowing snow frequency pattern as shown in Fig. [35](#). The maximum
442 values (white areas in Fig. [810](#)) exceed about 3×10^6 tons of snow per km per year. In the
443 supplemental Figs. [S10](#) and [S112](#) we display the MERRA-2 average 10 m wind speed and
444 direction for the years 2007–2015. By inspection of Figs. [S10](#) and [S112](#) it is seen that the overall
445 transport in East Antarctica is generally from south to north and obviously dominated by the
446 katabatic wind regime. It is immediately apparent that the average wind speed and direction
447 does not change much from year to year, with the former helping to explain why the average
448 continent-wide blowing snow frequency is also nearly constant from year to year (not shown).

449 **4. Error Analysis**

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

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

458 7) Limited spatial sampling

459 The magnitude of some of these can be estimated, others are hard to quantify. For instance, 1),
460 2) and 6) are directly involved in the calculation of sublimation (Eq. 6). The error in extinction,
461 particle radius, temperature and moisture can be estimated. The error associated with the
462 attenuation of the lidar signal above the blowing snow layer (3) is probably very small over the
463 interior of Antarctica, but could be appreciable nearer the coastline. In the interior, clouds are a
464 rare occurrence and when present are usually optically thin. Cloudiness increases dramatically
465 near the coast both in terms of frequency and optical depth. Here the effect of overlying
466 attenuating layers could be appreciable in that it would reduce the backscatter of the blowing
467 snow layer and the derived extinction. This in turn would lead to a lower blowing snow mixing
468 ratio and thus lower sublimation and transport. The effect of attenuation within the layer is
469 unaccounted for here and will also reduce the amount of calculated blowing snow sublimation.

470 With regard to 5) above, the method presented here cannot reliably detect blowing snow layers
471 less than 30 m thick. Therefore, sublimation associated with these layers is not accounted for.
472 Other studies have shown that drifting snow sublimation within the salutation layer can be very
473 significant (Huang et al., 2016). There is a further point to be made with respect to clouds that
474 relates to 5) above. The method we use to detect blowing snow will not work in the presence of
475 overlying, fully attenuating clouds. It is reasonable to suspect that cyclonic storms which
476 impinge upon the Antarctic coast and travel some distance inland would be associated with
477 optically thick clouds and contain both precipitating and blowing snow. Our method would not
478 be able to detect blowing snow during these storms, but we would not count such cases as
479 “observations”, since the ground would not be detected. The point is, blowing snow probably
480 occurs often in wintertime cyclones, but we are not able to detect it. This could lead to an
481 under prediction of blowing snow occurrence, especially near the coast. Also, blowing snow
482 layers less than 20 - 30 m thick would also likely be missed. It is not clear how often these layers
483 occur, but they are known to exist and missing them will produce an underestimate of blowing
484 snow sublimation and transport amounts. With regard to spatial sampling (7 above), unlike
485 most passive sensors, CALIPSO obtains only point measurements along the spacecraft track at

486 or near nadir. On a given day, sampling is poor. CALIPSO can potentially miss a large portion of
487 blowing snow storms such as is evidenced from inspection of Fig. 68. We have seen many
488 examples of such storms in both the MODIS and CALIPSO record. Quantifying the effect of poor
489 sampling on sublimation estimates would be difficult but should be pursued in future work.

490 **4.1 Sensitivity Analysis**

491 A major limitation of this work is the uncertainty inherent in the meteorological data used for
492 obtaining the temperature and moisture within the blowing snow layer. Re-analyses like
493 MERRA-2 do not have the vertical or horizontal resolution to enable an accurate description of
494 the temperature and moisture profile through the blowing snow layer. Also, as mentioned in
495 section 2.1, MERRA-2, or more accurately the GEOS-5 model on which it is based, does not
496 incorporate the effects of blowing snow sublimation on the moisture within the layer. Even so,
497 we have already shown that MERRA-2 is moist compared to surface observations and to other
498 models. Thus we do not feel that using the MERRA-2 moisture will cause a large overestimation
499 of blowing snow sublimation. However, it is important to examine the effects of moisture on
500 the calculated sublimation. To demonstrate this we have taken one CALIPSO track with blowing
501 snow (shown in Fig. 911a) and plotted the MERRA-2 humidity (wrt ice) and the calculated
502 blowing snow sublimation along the track. We then increased the moisture amount by 5 and
503 10% to see the effect on the calculated sublimation. The temperature was not changed. In Figs.
504 911b – 911d the MERRA-2 relative humidity is the dark solid line, MERRA-2 temperature is the
505 dotted line and the calculated blowing snow sublimation is the thin black line. The temperature
506 and moisture shown are the MERRA-2 averages through the blowing snow layer. Figure 911b
507 shows the unperturbed MERRA-2 moisture and the resulting blowing snow sublimation
508 (integrated through the layer). In Fig 911c and 911d we have increased the MERRA-2 relative
509 humidity by 5 and 10%, respectively. The effect on the average blowing snow sublimation is
510 marked. A 10% increase in relative humidity produces about a 30% reduction in the calculated
511 blowing snow sublimation. This exercise demonstrates the non-linear effect of the moisture
512 level on the calculated sublimation.

513 If we assume then that the error in moisture is 10%, we must accept that the resulting blowing
514 snow sublimation could be 30% too high. But is that realistic, given the fact that the MERRA-2
515 data were shown to be moist compared to observation and other models (moister on average
516 by 7%)? We do not think so. Rather we take the error in MERRA-2 moisture to be 5%. This
517 produces an 18% over estimation of sublimation (Fig. 911b compared to Fig. 911c). This error
518 must be combined with other errors such as extinction, particle radius and temperature. Here
519 we assume the extinction error to be 20 %, the particle radius error 10 % and the temperature
520 error 5%. In Eq. (6) these terms are multiplicative. The total error in sublimation is then:

$$521 \quad \pm 1 - (0.8 * 0.9 * 0.95) + 0.18 = \pm 0.50$$

522 This indicates that the sublimation values derived in this work should be considered to have an
523 error bar of ± 50 %. The error in computed transport involves error in wind speed and the
524 blowing snow mixing ratio, the latter being dependent on extinction and particle size. If we
525 assume wind speed has an error of 20 %, extinction 20 % and particle size 10 %, the total error
526 in transport is:

$$527 \quad \pm 1 - (0.8 * 0.8 * 0.9) = \pm 0.42$$

528 **5. Summary and Discussion**

529 This paper presents the first estimates of blowing snow sublimation and transport over
530 Antarctica that are based on actual observations of blowing snow layers from the CALIOP space
531 borne lidar onboard the CALIPSO satellite. We have used the CALIOP blowing snow retrievals
532 combined with MERRA-2 model re-analyses of temperature and moisture to compute the
533 temporal and spatial distribution of blowing snow sublimation and transport over Antarctica for
534 the first time. The results show that the maximum sublimation, with annual values exceeding
535 250 (± 125) mm swe, occurs within roughly 200 km of the coast even though the maximum
536 frequency of blowing snow most often occurs considerably further inland. This is a result of the
537 warmer and drier air near the coast which substantially increases the sublimation. In the
538 interior, extremely cold temperatures and high model relative humidity lead to greatly reduced

539 sublimation. However, the values obtained in parts of the interior (notably the Megadune
540 region of East Antarctica - roughly 75 to 82S and 120 to 160E) are considerably higher than
541 prior model estimates of Dery and Yau (2002) or Lenaerts et al., 2012a). This is most likely due
542 to the very high frequency of occurrence of blowing snow as detected from CALIOP data in this
543 region which is not necessarily captured in models (Lenaerts et al., 2012b).

544 The spatial pattern of the transport of blowing snow follows closely the pattern of blowing
545 snow frequency. The maximum transport values are about 5 Megatons per km per year and
546 occur in the Megadune region of East Antarctica with other locally high values at various
547 regions near the coast that generally correspond to the maximums in sublimation. We
548 attempted to quantify the amount of snow being blown off the Antarctic continent by
549 computing the transport along the coast using only the v component of the wind. While this
550 may produce an overestimate of the transport (since the Antarctic coast is not oriented east-
551 west everywhere), we find the amount of snow blown off the continent to be significant and
552 fairly constant from year to year. The average off-continent transport for the 9 year period
553 2007–2015 was 3.68 Gt yr⁻¹ with about two thirds of that coming from East Antarctica and over
554 one third from a relatively small area between longitudes 135E and 160E.

555 Over the nearly 11 years of data, the inter-annual variability of continent wide sublimation
556 (Table 1) can be fairly large – 10 to 15 % - and likely the result of precipitation variability and or
557 changes in the MERRA-2 temperature and moisture data. There seems to be a weak trend to
558 the sublimation data with earlier years having greater sublimation than more recent years.
559 However, based on the short length of the time series and the likely magnitude of error in the
560 sublimation estimates, the trend cannot be considered statistically significant.

561 The overall spatial pattern of blowing snow sublimation is consistent with previous modelling
562 studies (Dery and Yau, 2002 and Lenaerts et al., 2012a). However, we find the Antarctic
563 continent-wide integrated blowing snow sublimation to be larger than previous studies such as
564 Lenaerts et al., (2012a) (393 ±196 vs roughly 190 Gt yr⁻¹), even though the observations include
565 only the area north of 82° S. The maximum in sublimation is about 250 (±125) mm swe per year

566 near the coast between longitudes 140E and 150E and seems to occur regularly throughout the
567 11 year data record. There are a number of reasons for the higher sublimation values in this
568 study compared to prior estimates. 1) The depth of the layer: the average blowing snow layer
569 depth as determined from the CALIOP measurements is 120 m. Layers as high as 200 - 300 m
570 are not uncommon. It is likely that models such as those cited above do not always capture the
571 full depth of blowing snow layers, thus producing a smaller column-integrated sublimation
572 amount. 2) We only compute sublimation from blowing snow layers that are known to exist
573 (meaning they have been detected from actual backscatter measurements). Models, on the
574 other hand, must infer the presence of blowing snow from pertinent variables within the
575 model. The existence of blowing snow is not easy to predict. It is a complicated function of the
576 properties of the snowpack, surface temperature, relative humidity and wind speed. Snowpack
577 properties include the dendricity, sphericity, grain size and cohesion, all of which can change
578 with the age of the snow. In short, it is very difficult for models to predict exactly when and
579 where blowing snow will occur, much less the depth that blowing snow layers will attain. 3) The
580 lack of blowing snow physics within the MERRA-2 reanalysis. This produces perhaps the largest
581 uncertainty in the derived results. It was shown that MERRA-2 is slightly colder and moister
582 than some surface measurements and moister compared to other re-analyses. However, given
583 the limited number of comparisons, a definitive conclusion on the accuracy of MERRA-2 data
584 cannot be drawn. Since the model on which MERRA-2 re-analysis is based (GEOS-5) does not
585 include blowing snow (and thus blowing snow feed backs on moisture and temperature), it is
586 likely that our estimates of blowing snow sublimation are probably too high. However, the fact
587 that we do not include blowing snow layers less than 30 m in depth and are not able to detect
588 blowing snow beneath thick clouds layers means that we are missing potentially important
589 contributions to sublimation. An addition, the retrieved blowing snow number density below
590 about 80 m is probably too low for layers greater than 120 m in depth because of lidar signal
591 attenuation. This will act to erroneously reduce the calculated sublimation. While we estimate
592 an upper limit on the error of our blowing snow sublimation results as 50%, we believe that the
593 error is considerably less than that.

594 Future work should involve coupling the CALIPSO blowing snow observations with a regional
595 model that contains blowing snow physics. This could increase the accuracy of the calculated
596 blowing snow sublimation by incorporating the moisture feedback processes within the layer
597 that have been neglected here.

598 **Data Availability**

599 The CALIPSO calibrated attenuated backscatter data used in this study can be obtained from
600 the NASA Langley Atmospheric Data Center at: [https://earthdata.nasa.gov/about/daacs/daac-
601 asdc](https://earthdata.nasa.gov/about/daacs/daac-
601 asdc)

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

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

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741 *Oceanic Technol.*, 26, 2310-2323, doi: 10.1175/2009jtecha1281.1, 2009.

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

| Year | Average Sublimation (mm swe) | Integrated Sublimation (Gt yr ⁻¹) |
|------------|---------------------------------|--|
| 2006* | 28.3 ± 14.1 | 255 ± 128 |
| 2007 | 56.8 ± 28.4 | 514 ± 207 |
| 2008 | 49.2 ± 24.6 | 446 ± 223 |
| 2009 | 45.3 ± 22.6 | 409 ± 204 |
| 2010 | 42.9 ± 21.4 | 388 ± 194 |
| 2011 | 47.6 ± 23.8 | 431 ± 215 |
| 2012 | 44.4 ± 22.2 | 402 ± 201 |
| 2013 | 47.7 ± 23.8 | 432 ± 216 |
| 2014 | 41.5 ± 20.7 | 376 ± 188 |
| 2015 | 41.3 ± 20.6 | 374 ± 187 |
| 2016* | 33.2 ± 16.6 | 301 ± 150 |
| AVG | 43.5[*] ± 21.7 | 393.4[*] ± 197 |

745
 746 Table II. The total transport (Gt yr⁻¹) from continent to ocean for various regions in Antarctica
 747 for 2007–2015.

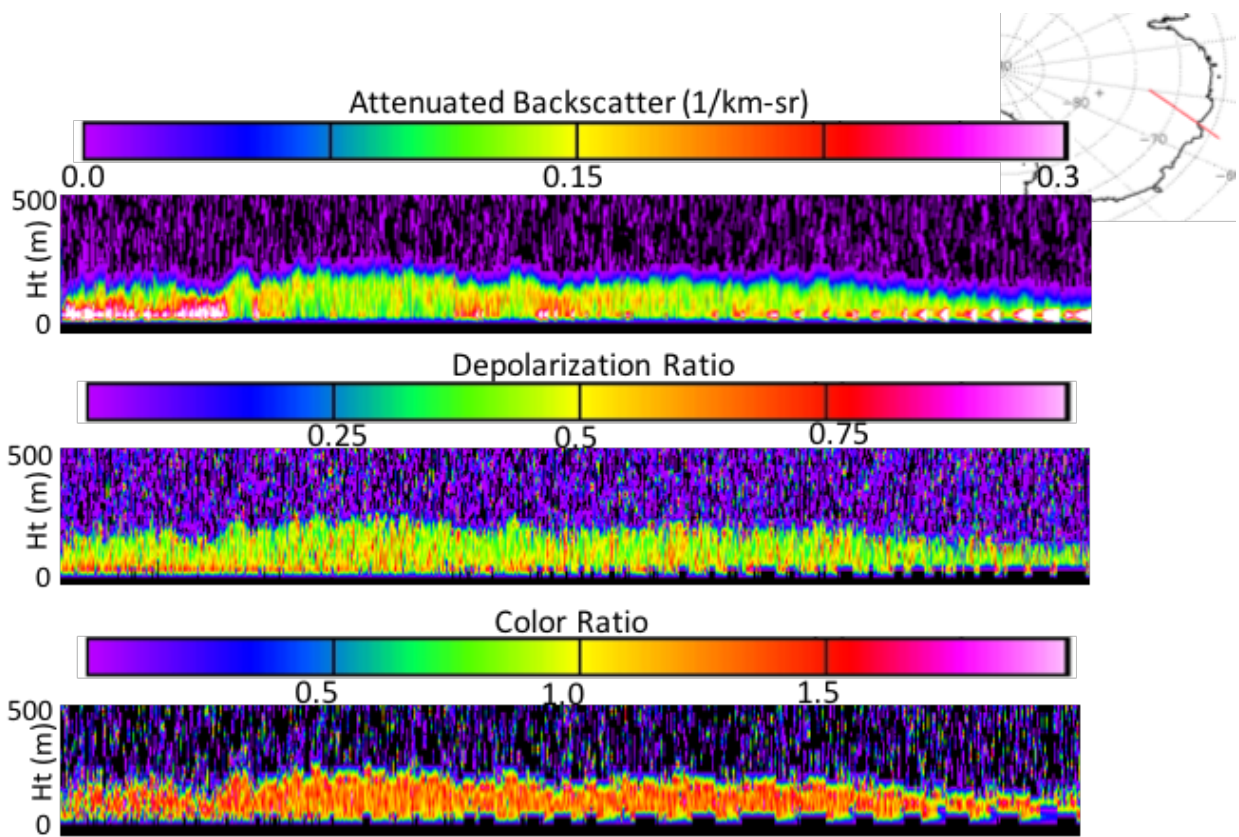
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| Year | East Antarctica | West Antarctica | 135E – 160E | 80W – 120W |
|------------|-----------------|-----------------|-------------|-------------|
| 2007 | 2.52 | 1.29 | 1.72 | 0.82 |
| 2008 | 2.20 | 1.43 | 1.21 | 0.90 |
| 2009 | 2.63 | 1.27 | 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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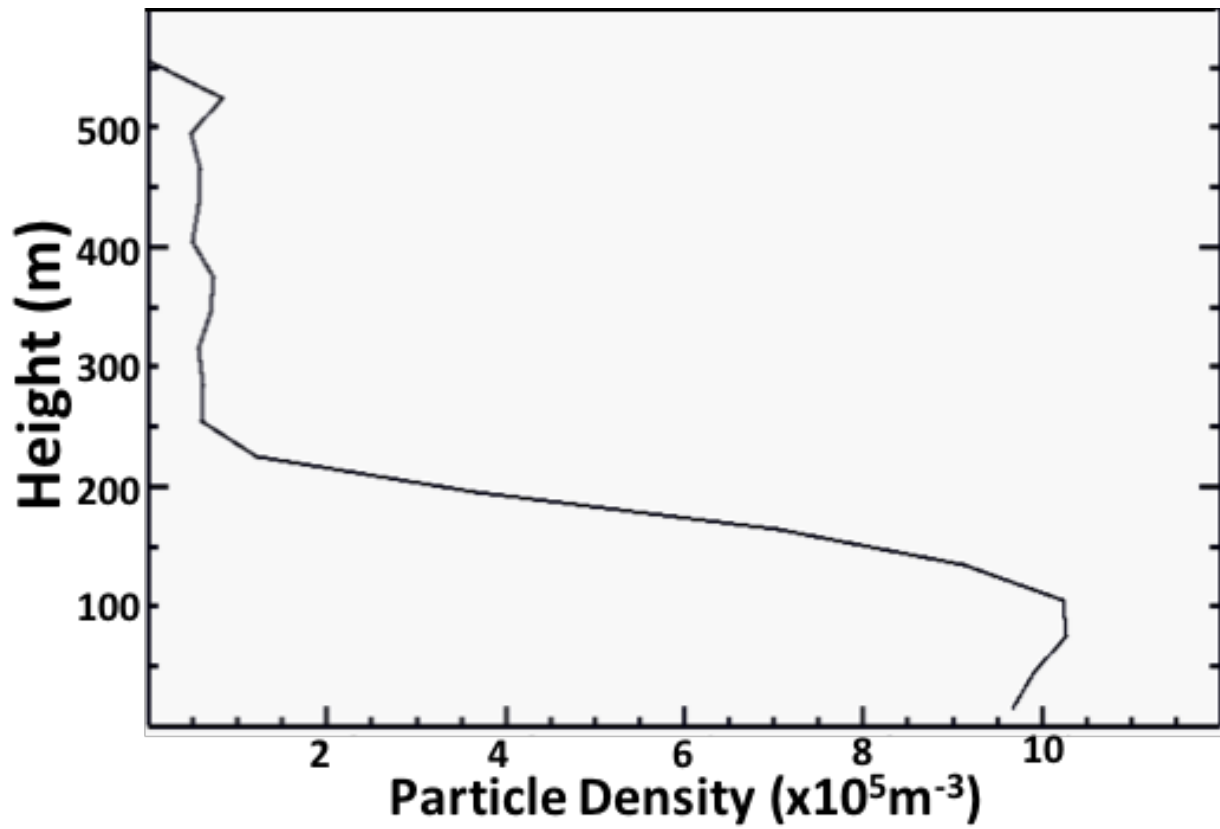


755

756 Figure 1. A typical Antarctic blowing snow layer as measured by CALIPSO on May 28, 2015 at
 757 17:08:41 – 17:11:33 UTC. Displayed (from top to bottom) are the 532 nm calibrated, attenuated
 758 backscatter, the depolarization ratio at 532 nm, and the color ratio (1064 nm / 532 nm).

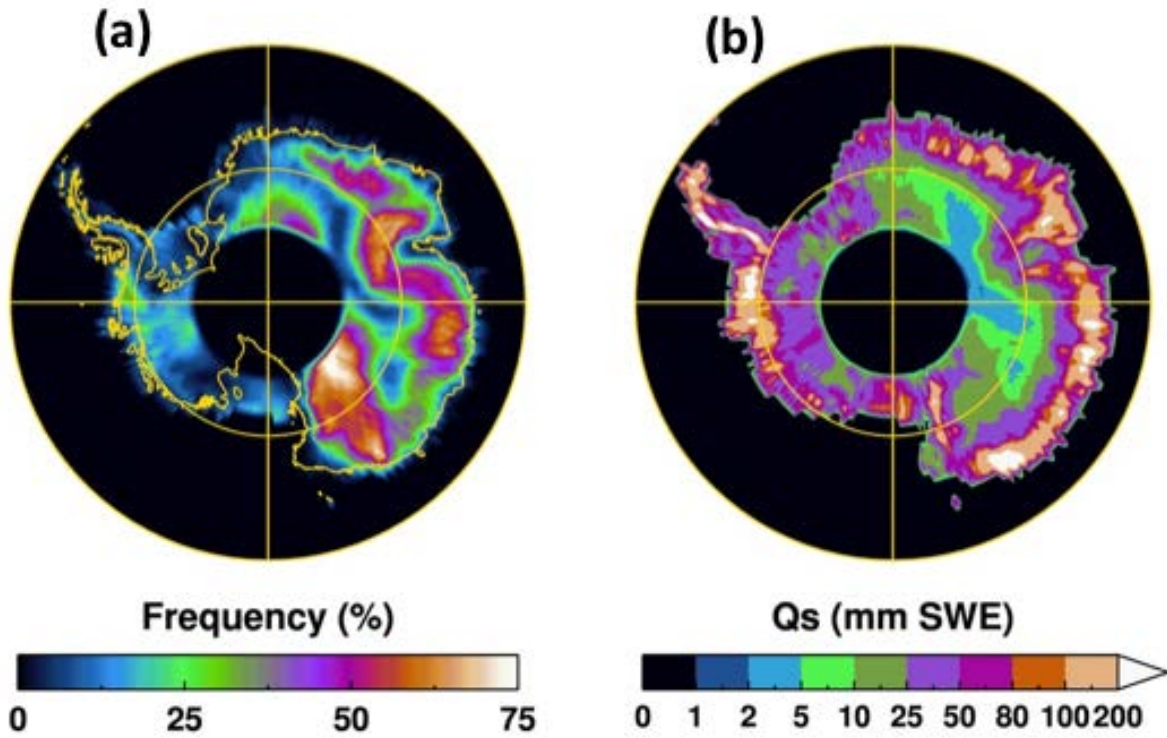
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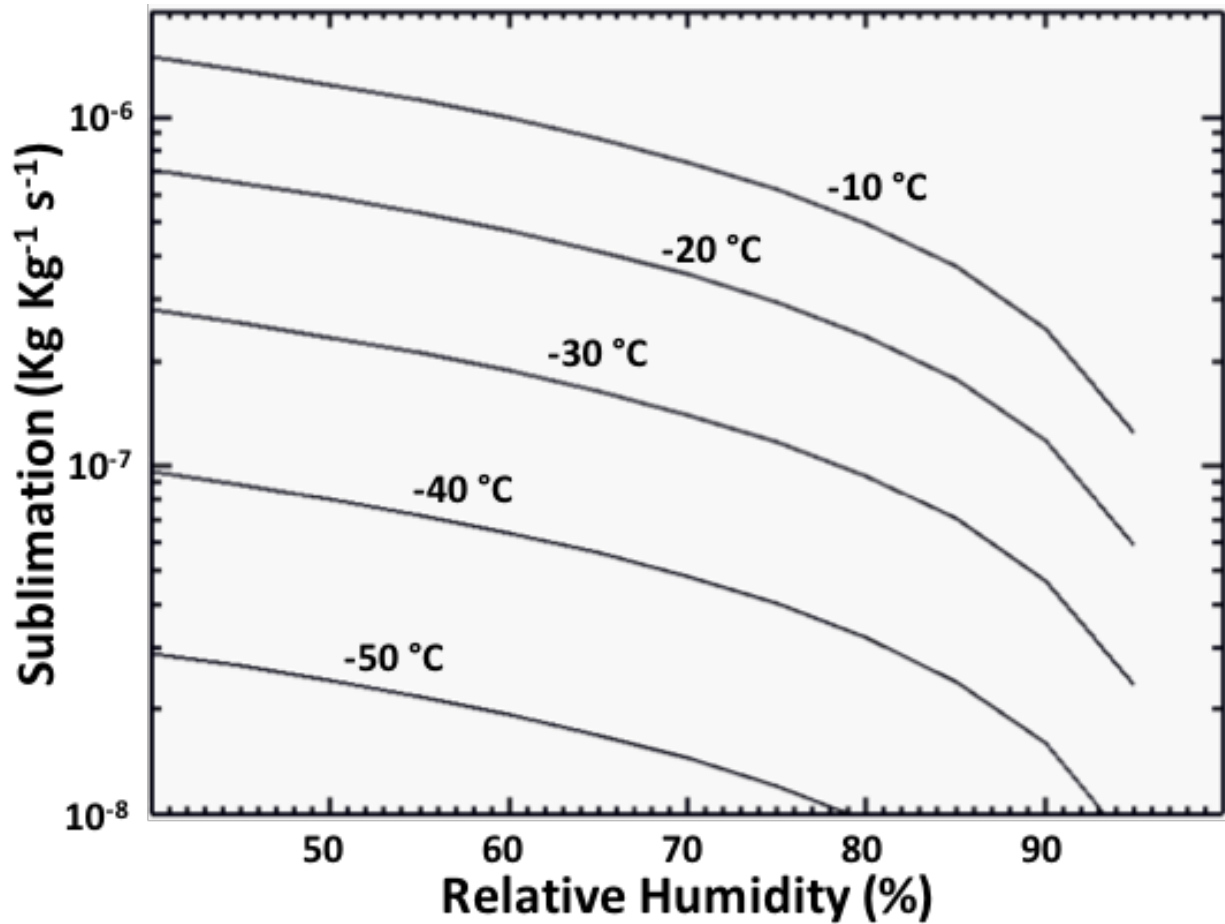
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762 Figure 2. Average particle density profile (Eq. 2) through the blowing snow layer shown in Fig. 1.



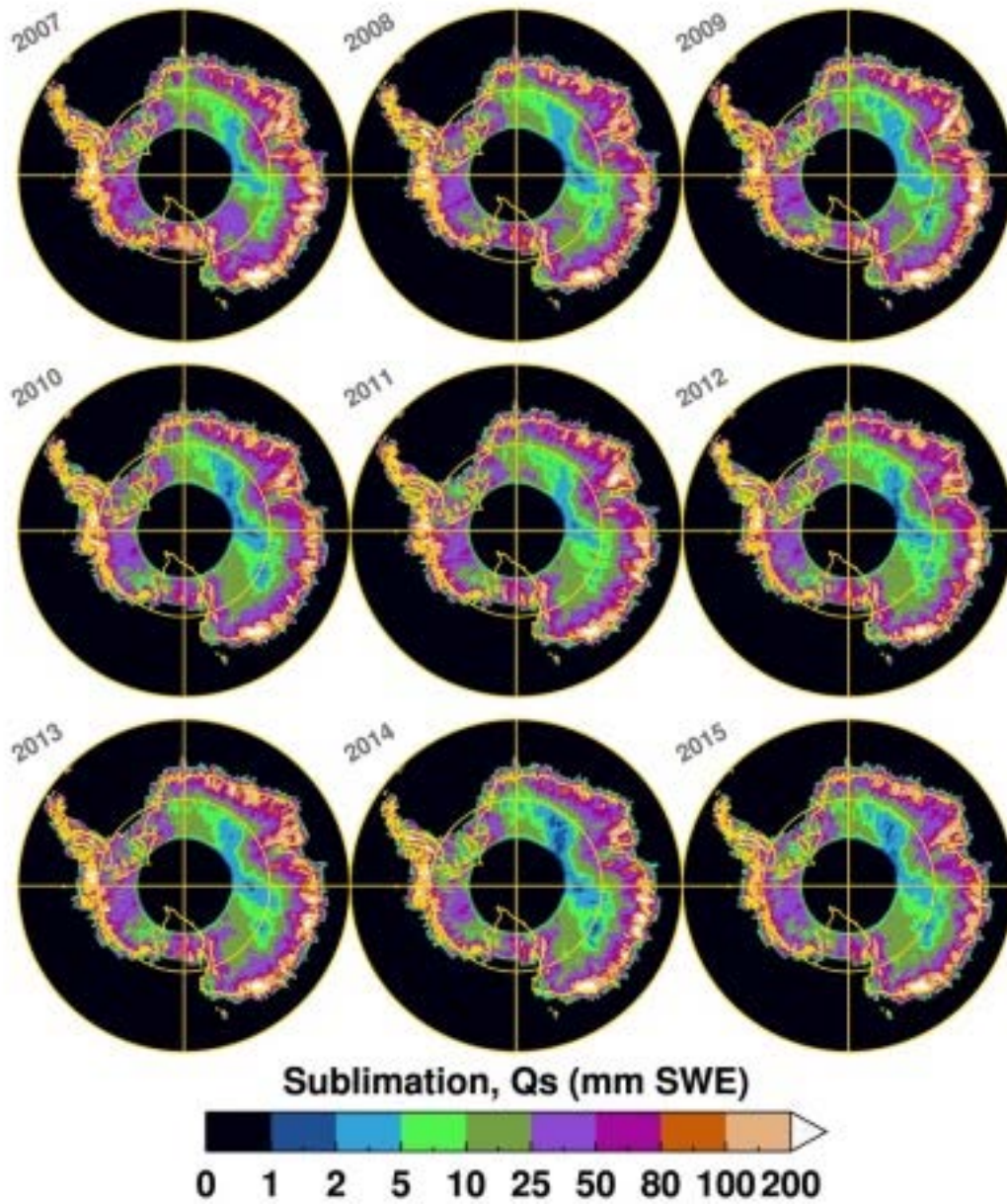
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764 Figure 3. (a) The average April through October blowing snow frequency for the period 2007–
765 2015. (b) The average annual blowing snow sublimation for the same period as in (a).



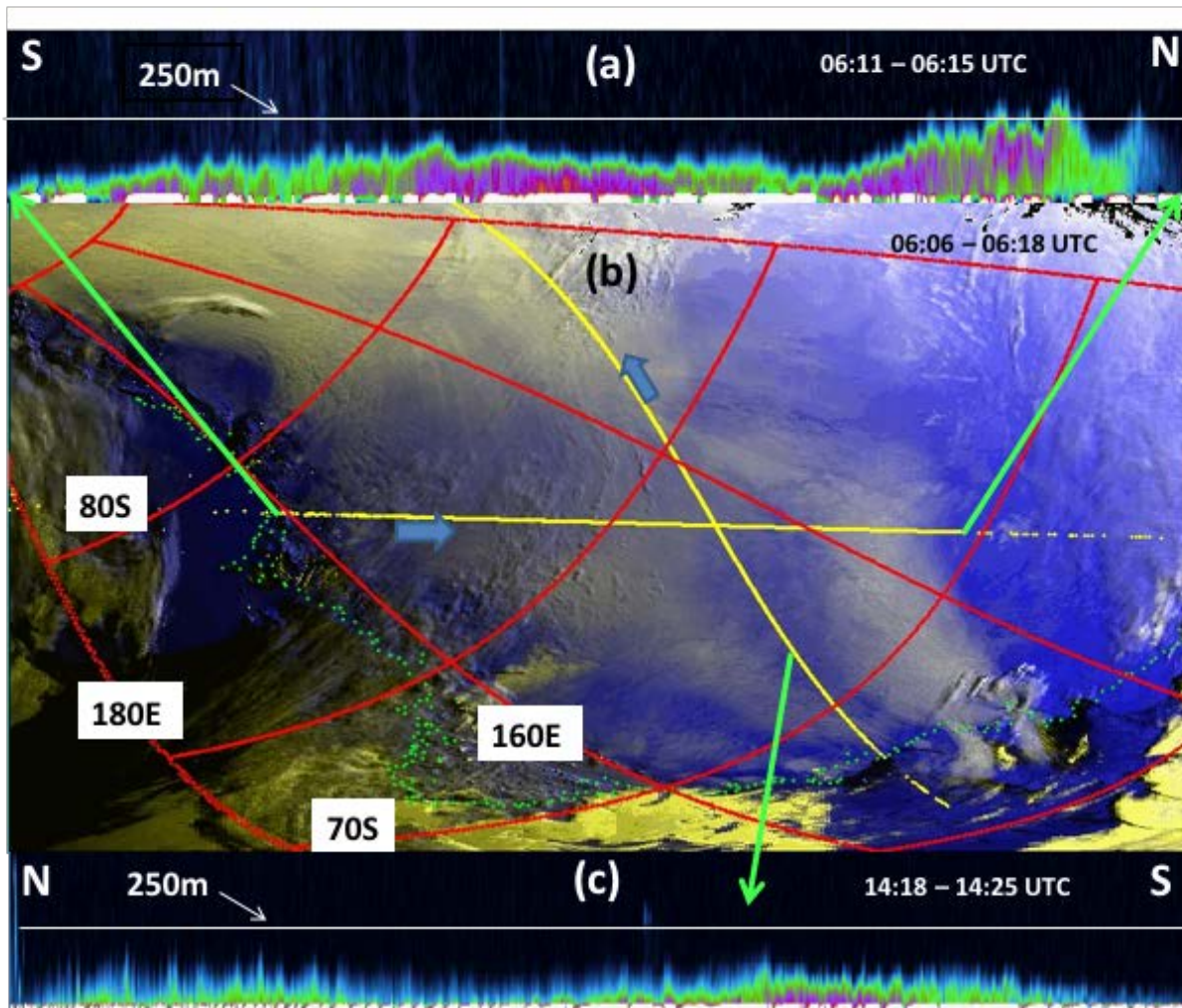
766

767 Figure 4. Computed blowing snow sublimation rate using Eqs. (3) and (4) as function of relative
 768 humidity for varying air temperatures. The particle density value used in Eq. (3) was 10^6 m^{-3}
 769 which corresponds to a blowing snow mixing ratio (q_b) of $4.7 \times 10^{-5} \text{ kg kg}^{-1}$



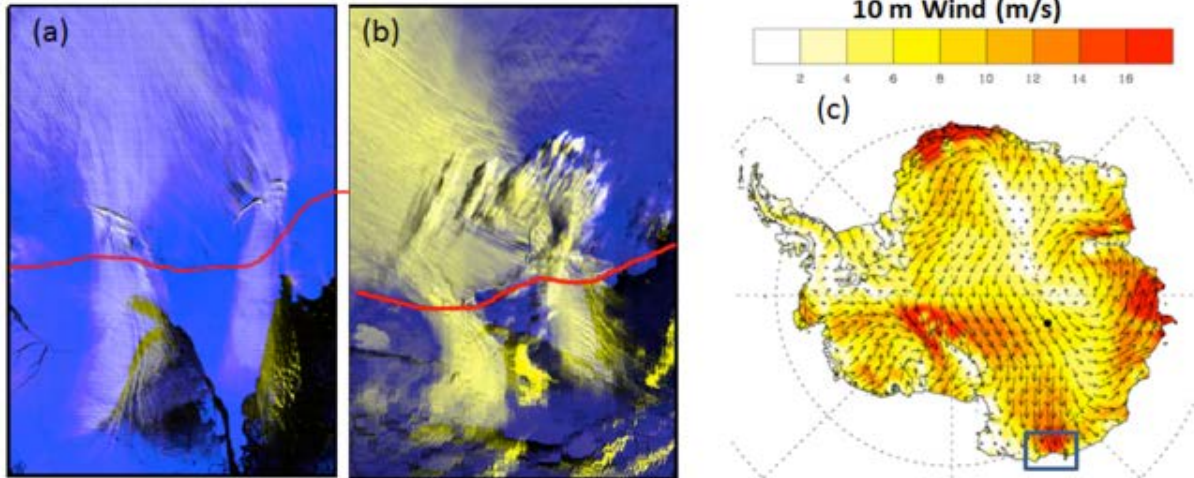
770

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



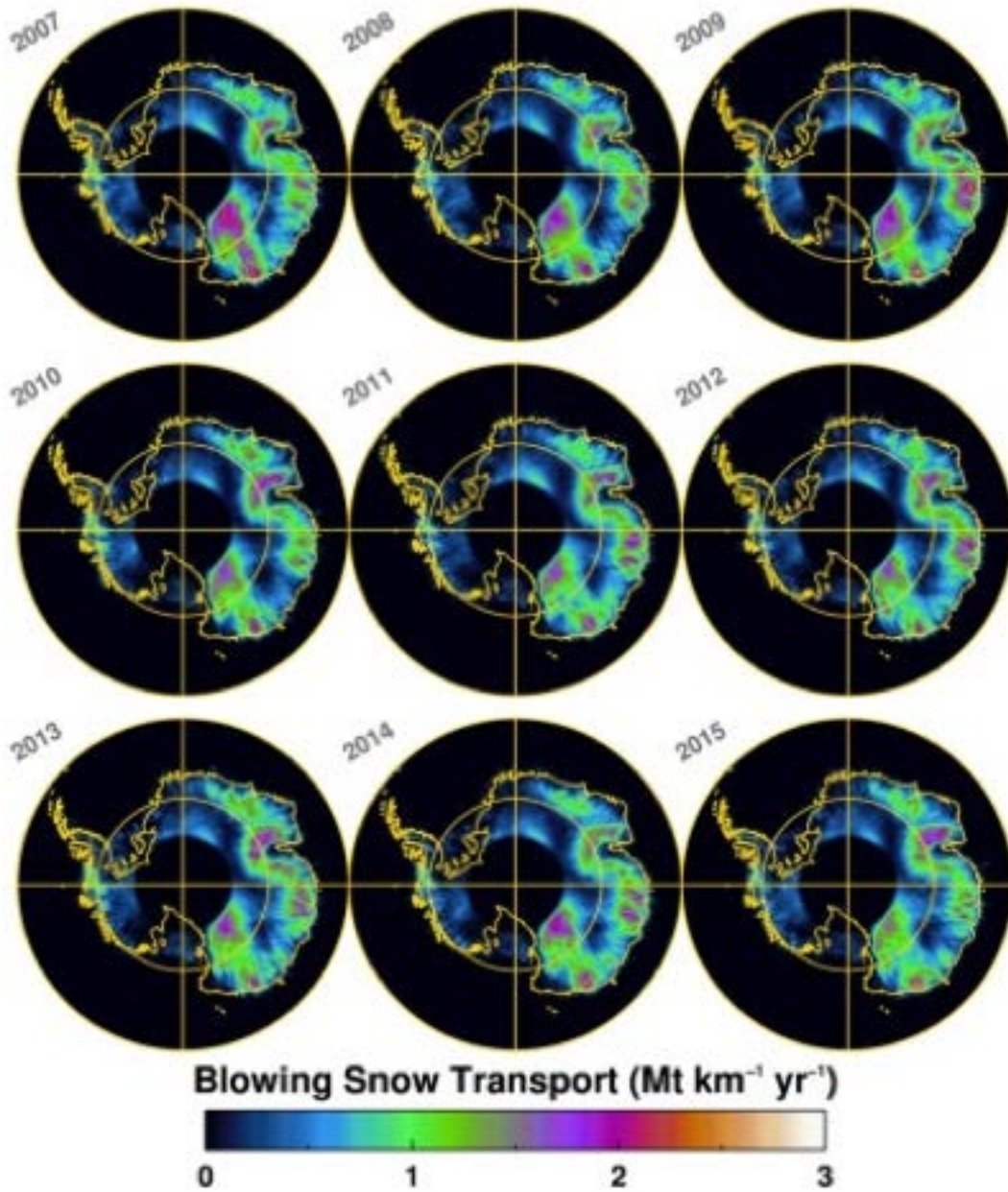
772

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



780

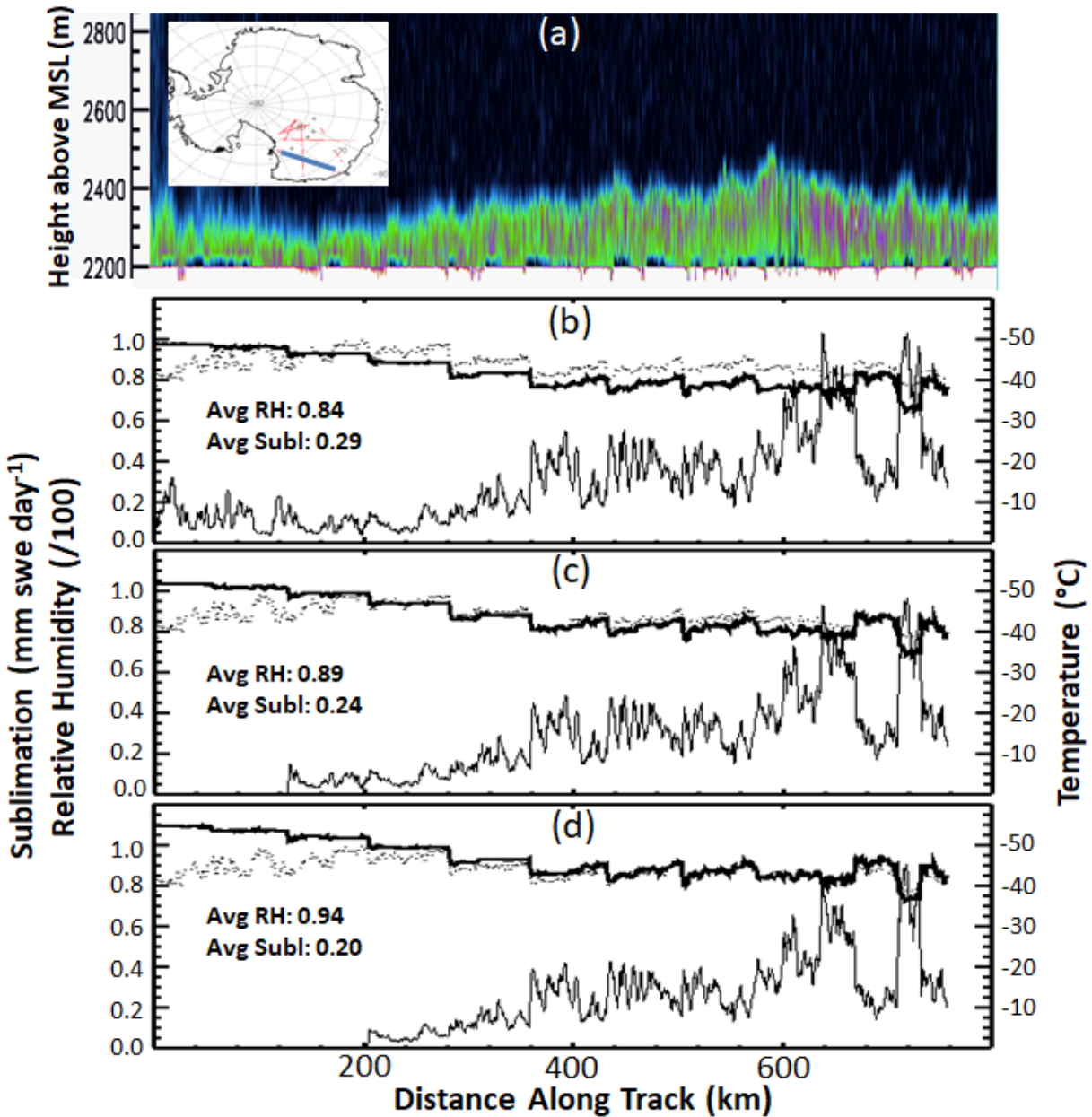
781 Figure 7. (a) MODIS false color image on October 13, 2009, 23:00 UTC and (b) October 14, 2009,
 782 06:16 UTC. The red line is the approximate position of the coastline. (c) The 10 m wind speed
 783 from the AMPS model (Antarctic Mesoscale Prediction System) for October 14, 2009. The area
 784 covered by the MODIS images is roughly that indicated by the blue box in (c).



785

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

788



789

790 Figure 9. (a) CALIPSO backscatter showing blowing snow layer along the blue line in the map
 791 inset on 10/12/2010 at 05:51 UTC. (b) Average MERRA-2 moisture (dark black line),
 792 temperature (dotted line) and calculated sublimation through the blowing snow layer along the
 793 CALIPSO track. (c and d) Same as in (b) but increasing MERRA-2 humidity by 5 and 10%,
 794 respectively.

795