Reviewer 1, major comments (Authors responses in red)

I have strong reservations with regard to the use of simplified, steady-state parameterizations and meteorological forcing, which put a strong constraint on the resulting blowing snow sublimation. The authors use MERRA-2 temperature and RH data to derive sublimation rates, but these MERRA-2 data are (a) not at all evaluated over Antarctica, and – more importantly – (b) since MERRA-2 does not simulate blowing snow, do not imply the atmospheric effects of the well-documented self-limiting behavior of blowing snow sublimation, in which sublimation will lead to latent heat release to the atmosphere at the top of the blowing snow layer, in turn cooling and moistening the atmosphere and limiting subsequent sublimation. This effect is expected to have a first-order negative effect on the sublimation (while retaining the blowing snow layer transport active, so this is not observable from space), and should – in some way or another – be included in this approach. I realize that the authors do not (and do not want to) utilize a model that includes this behavior, nor include blowing snow processes in MERRA-2. One option is to perform multiple sensitivity tests with gradually higher RHice values and lower temperatures, based on and in combination with a MERRA-2 near-surface climate evaluation at select stations over Antarctica. These sensitivity tests should be combined with varying other important parameters to determine sensitivity. For instance, why did the authors choose a fall velocity of 0.1 m/s? I would strongly suggest to expand the Section 4 and include a detaileddescription of the sensitivity tests.

The authors understand the reviewers concerns regarding the use of MERRA-2 re-analysis as the source of meteorological data for the blowing snow sublimation computations. Indeed, the MERRA-2 data have not been evaluated over Antarctica and the reviewers concerns are well justified. However, we have compared MERRA-2 relative humidity amounts with a few surface stations, dropsondes and other models and determined that MERRA-2 has, in general, a cold and moist bias. We have added section 2.1 to the paper which describes the MERRA-2 data and also show comparisons of MERRA-2 relative humidity and temperature to three surface stations and other models. The comparisons are shown in new figures 2 and 3.

The reviewer also notes that since MERRA-2 is a re-analysis based on the GEOS-5 model, it does not contain blowing snow physics that would capture processes such as the cooling and moistening of the blowing snow layer as sublimation proceeds. While we understand that blowing snow sublimation will act to moisten the layer, we also know that other processes operate, especially in deep blowing snow layers, that can act to reduce the humidity within the layer. For instance, entrainment of drier and warmer air at the top of the inversion and adiabatic warming associated with the descending katabatic flow. Warming of the layer can also occur from trapping of longwave radiation. Most if not all of the observations showing how blowing snow will increase the moisture to or near saturation have been made at or below 10 m height. There are no observations through the depth of very deep blowing snow layers such as those presented here. Recently, we have been working with the Concordiasi dropsondes and have identified a number that fell through deep blowing snow layers. While this work is as yet too preliminary to present (or include in the paper), the observations suggest that typically the layer is not saturated.

Also noted by the reviewer, we do not necessarily want to use a model that contains blowing snow physics at this time. However, this may be an area that can be explored in the future.

The Reviewer suggests a sensitivity study to see how increasing moisture will affect the blowing snow sublimation values. We regard this as a valuable suggestion and have added a new section 4.1 which addresses the effect of increasing moisture on blowing snow sublimation. Based on the results we have modified the wording and the computation of the error beginning on line 499. We now conclude error could be as large as 50%

We have also modified the conclusion section starting on line 565 where we talk about error.

As to why we chose a fall velocity of 0.1 m/s, it is a value close to the average fall velocity found by Mann et al., 2000. Granted these are for the surface layer and for larger particles. The particles in deep blowing snow layers will be smaller and have a lower fall speed. We did a test and found that reducing the fall speed by an order of magnitude (to 0.01 m/s) reduces the calculated sublimation by about 10%. Just between you and me, I wonder whether fall speed is the right thing to use here, considering that in many of these blowing snow episodes, wind speeds approach 20 m/s and wind speed and directional shear is very large. This has to produce incredible turbulence which will act to ventilate the particles well beyond simply falling at a constant speed through the atmosphere.

Reviewer 1 Minor comments: (Authors response in red font)

The writing should be improved in places and caution is warranted to very clearly describe the process the authors are referring to. Also, some parts are clearly too speculative and should be revised (see below).

L13: near-surface

Changed

L15: define surface mass balance

Added Equation 1 on line 48 and text to explain on lines 41-45

L17 and beyond: clearly mention the time period considered in this study

Done on line 19

L23: blowing snow sublimation!

Done

L29: 2006-2015

Done

L94: it would be helpful to mention all sublimation rates (also those from earlier literature)in the same units to facilitate comparison. Which time period are these from?

Changed all references to amount to mm swe yr^{-1}. The Halley observations were form 1995 and 1996. This is added to text on line 95.

L108: 20,000 km – reference needed

Added

L152: 1064/532 – include units

Added

L189: (Walden et al., 2003)

Changed

L233: How to go from blowing snow mixing ratio to extinction? What are the units of this extinction, and why does they relate as alfa(z) = 3/2 qb(z)/r(z) ?

We do not go from blowing snow mixing ratio to extinction. Rather we first compute the extinction and from that the particle number density N(z). From N(z) we get blowing snow mixing ratio via equation 3. The units of extinction are 1/m

L240 and around: The use of MERRA-2 needs to be described here. How are T and RH incorporated here? How is temporal and vertical interpolation dealt with?

We have added section 2.1 that describes the MERRA-2 data, its characteristics and how we use it to obtain values at a given point in space and time. Note this begins on line 175 of revised text.

L241: Equations are not numbered

Fixed

L321: It would be very helpful to plot the CloudSAT precipitation numbers and plot theratio sublimation/precipitation to guide this discussion.

The Authors agree this would be a useful thing to do, but we do not have time to obtain the data and do the analysis.

L339: 419 Gt/yr – this is a different number than mentioned anywhere else.

Should have been 393. This is now corrected.

L344-348: This is extremely speculative and contains the wrong translation from Gt/yrto mm sea level rise (360 Gt = 1 mm SLE). Most of the sublimation is probably recycled on the ice sheet, and of course Antarctic SMB is positive and dominated by precipitation. Please remove.

Agreed. It has been removed.

L349-355: is the trend significant? Probably not, with significant inter-annual variability and only 10 years in the time series. If it is not significant, please remove. If it is significant, it would be useful to relate this to MERRA-2 T and RH averages.

Removed

L357: clarify if you consider the grounded or total (include ice shelves) ice sheet.

Clarified on line 380

L371: size of Texas – quantify.

We have added the exact area of Texas to the text

Table I: Is the average from 2006-2015 (the full years)? Clarify.

Table I now explains this in a footnote.

Figures 9 and 10 can be removed or moved to supplements. They do not contain any results that are necessary to be shown in a separate (main) figure.

Agreed. We have moved them to the supplement.

Reviewer 2 comments (note, Authors responses in red)

General comments This paper presents very interesting and unprecedented continent wide statistics of blowing snow over Antarctica from long-term satellite observations. These include estimations of blowing snow sublimation, a significant but poorly known component of the Antarctic surface mass balance. Such works are essential for evaluation of atmospheric models from which the total surface mass budget of the ice sheet can be estimated. However, there are some important missing aspects and information in the study that I would like to report here. Of particular concern is the method from which sublimation estimates are computed. One possibly very significant source of error is an underestimation of atmospheric moisture by MERRA-2: the method does not take into account the fact that moisture from blowing snow sublimation is retained while air flows further through blowing snow regions, strongly reducing (or cutting if saturation is reached) any further sublimation of blowing snow downstream. MERRA-2 does not account for blowing snow sublimation, thus the method constantly resets air moisture to values for which blowing snow sublimation has never occurred, and very likely overestimates total sublimation.

We have added section 2.1 to the paper to show that MERRA-2 is moist compared to surface observations. Please see our comments to Reviewer 1 about the effect of blowing snow sublimation on the humidity of the layer. In short, there are no observations supporting the assertion that blowing snow sublimation will lead to saturation of the layer when you are dealing with layers 100-400 m thick. The only observations to support this are below 10 m height. Entrainment of dry and warm air from above, descending air in the katabatic flow and warming of the layer through absorption of longwave radiation are all process that can act to keep the layer from saturating.

Specific comments Observational studies on blowing snow in Antarctica are very scarce, to the extent that continuous measurements extending beyond a few weeks or months barely exist. However, considerable efforts have been made in the recent years on that specific topic, that you might have missed in your bibliography. An observation campaign dedicated to blowing snow has been run in January 2010 by the Laboratoire de Glaciologie et Géophysique de l'Environnement (LGGE, France) in coastal Adélie Land. Some of the collected data have been presented, for instance, in Trouvilliez et al. (2014), Barral et al. (2014) and Amory et al. (2016, 2017), and used for evaluation of preliminary modelling results (Gallée et al. 2013, Amory et al. 2015). Ground measurements on the ice sheet have been performed using second-generation acoustic FlowCapt™ sensors. While these sensors have been shown to slightly underestimate the blowing snow flux compared to optical snow particle counters SPC-S7 in the French Alps, they remain excellent detectors of blowing snow occurrences (Trouvilliez et al. 2015). To date, up to 7 years (2010-2016) of continuous ground measurements of blowing snow frequency in coastal Adélie Land are available (for comparison with CALIPSO data). The dataset also includes (discontinuous) measurements of snow particle size performed since 2013 at 50-m height above the ground with a SPC at Dumont d'Urville station (see Palerme 2014). I'm part of the research team that has produced (and still does) these observations and I'm open to discuss it with the authors if they wish.

The Authors thank the Reviewer for this valuable information and references. We have added Barral et al. (2014) and Trouvilliez et al. (2014) to the paper. We would like to explore these observations with the Reviewer in the near future.

P8, L222: Figure 2 shows an increase in particle density with height for the first 100 meters above the surface. This is surprising since the density of blowing snow particle is supposed to decrease as distance from the ground (i.e., from the particle source) increases (see for instance the strong decrease within the first 10 meters above the ground in Fig. 4 of Mann et al. 2000). Have you an idea of what can cause thisfeature?

This is a very astute observation by the Reviewer. One of the things that one has to be concerned with when analyzing the CALIPSO data is contamination of the atmospheric measurement by the ground return signal. One has to be very careful to eliminate the ground return from the backscatter profile. If this is not done correctly, it will add (incorrectly) signal to the lowest bin of the profile (the bin directly above the ground). We have been very cautious about this and have probably erred on the conservative side. Thus some of the larger signals that may have been present in the bin directly above the ground may have been eliminated because of the possibility of ground signal contamination. But the calculated particle density profile in (revised) Fig 4 has a maximum 3 or 4 bins above the ground. We believe this is due to attenuation of the lidar signal as it passes through the blowing snow layer. We have not attempted to correct for this, but if it were corrected it would tend to increase the particle number density as one approached the ground. We have added text to describe this and its effect.

P17, L431: In addition, clouds may be associated with precipitation which contributes moistening the dry surface air layer (Grazioli et al. 2017; http://www.the-cryospherediscuss. net/tc-2017-18/) and thus correspondingly reduces blowing snow sublimation.

We cannot detect blowing snow occurring beneath precipitating clouds and such cases are not included here.

P17, L449 and onwards: Although this aspect is already partly discussed in the paper, estimating blowing snow sublimation by using MERRA-2 re-analysis fields of moisture could be misleading because i) re-analysed moisture near the surface could be underestimated and ii) no retro-action of sublimation on moisture is accounted for. Systematic dry biases in atmospheric models and meteorological (re-) analyses that do not account for blowing snow have been discussed in Barral et al. (2014). Using a 3-year dataset of ground measurements at a coastal location in Adélie Land, they showed (their Figure 6) for 3 modelling products that the moisture error in the near-surface layer for the continental grid point closest to the measurement location is much larger than 5% (as considered in the error analysis in Section 4), and that the 3 models fail to represent the observed increase of atmospheric moisture with wind speed. For instance, the moisture error almost averages 100% for the ECMWF operational analysis for wind speeds exceeding 12 m/s. It is likely that most meteorological and climate models ignoring blowing snow are affected by similar dry biases, at least over windy peripheral areas of East Antarctica where blowing snow is highly active. In addition, in the blowing snow layer the air quickly saturates as part of the blowing snow sublimates. This limits the total amount of blowing snow that can be sublimated and thus negatively feeds back on blowing snow sublimation. Following the method presented in the paper, forcing the blowing snow parameterization with an atmospheric model that ignore blowing snow and its sublimation neglects this negative feedback. In otherwords, this makes the atmosphere acting as an infinite sink for water vapor. Then, even though the method presented relies on satellite observations, using raw moisture fields from such models to compute blowing snow sublimation very likely leads to significant overestimation. This appears to be a major limitation to the quantitative aspect of this work. Together with the arguments claimed in the discussion part, this certainly accounts for the large differences with previous model-derived estimates of Déry and Yau (2002) and Lenaerts et al. (2012). The overestimation of blowing snow sublimation compared to RACMO2 also seems questionable since the model has been shown to overestimate considerably the blowing snow flux and the resulting horizontal snow mass transport (see Lenaerts et al. 2014, their Figures 6b and 7c – pay attention to the difference between the left and right scales), suggesting an overestimation of the modelled sublimation.

We have added section 2.1 to the paper which demonstrates that MERRA-2 is cold and moist with respect to surface observations and other models. Moreover, in ongoing research we have seen that this bias in MERRA-2 is not limited to the surface (2 m). Using the Concordiasi dropsondes, we have seen that MERRA-2 is cold and moist even at levels above the surface. Based on this we do not feel that MERRA-2 has a dry bias. Rather, it is likely too moist. We have also added section 4.1 – a sensitivity analysis to show the effect of moisture error on the calculated sublimation. We have modified our error estimates as well. Also please see the Author's comment to Reviewer 1 who had the same concern about not including blowing snow physics.

Reviewer 3 Comments (Author's response in red font):

First of all, previous works are well reviewed in general except for the field observation lately carried out, such as Trouvilliez et al., 2014.

This reference has been added to the text at line 98

Line 156: Here new logic to detect the blowing snow is introduced. I wonder how the new one gives effect on the blowing snow detections shown in the authors' previous publications (overestimate, misrecognize or negligible?).

We have compared results with the previous algorithm and have seen very little change. This is stated on line 165 of the revised paper.

Line 185: Particle sizes of cirrus clouds are around 5 micron and are much less than the blowing snow particles. Is the same algorithm still applicable?

The reviewer is mistaken. Cirrus cloud particle sizes range from 10 to 100 um and even much larger (see Heymefield et al., 2002 for instance)

Line 209: The definition of particle radius seems too rough, since the sublimation rates (vapor pressure) strongly rely on the particle radius. Although, latter part of the manuscript on line 450, authors estimated the error caused by the radius is 10 %, I believe the contribution cannot be assessed by such a simple product.

The particle sizes produced by Equation 1 give sizes ranging from 40 (15 m) to 15 µm (500 m). Observations of blowing snow particle size made in the field are generally at 10 m or less. Until we get measurements that extend high up into the layer, this is about the best that can be done.

Line 221: Observations by Mann et al. should be done much lower altitude than the one discussed here. Is it worth comparing?

Mann does have a plot that goes up to 11 m, and by extrapolation one can get an idea of the value at say 30 m. Also Mann has a theoretical plot of number density that goes up to 20m.

Fig. 2: Particle density profile shown here is very interesting, however, at the same time, it is very confusing. Please explain the physical mechanism why the particle density increases with height, as far as I know it should be opposite, and shows the maximum at 100 m.

This is a good catch. I believe this is due to attenuation of the lidar return (the measured backscatter) as it traverses downward through the blowing snow layer. This is producing an underestimate of the extinction for the bins closest to the surface (say below 100 m). And the number density is computed from the extinction via equation 2. Note also that the radius of the particle is in the denominator of equation 2 and r decreases with height. This would cause N(z) to increase with height. The bottom line is I think you are right and the particle densities are lower than they should be in the lower 50-100 m. Note that this would lead to an under estimation of sublimation.

Line 239: Fk and Fd should be expressed in italic.

I am not sure what you mean. In my document, they are in italic.

Line 230: Kg ! kg

Fixed

Line 256: 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 the drastic temperature rise. It looks this process is not taken into account in the procedure shown here. If this is the case, I am afraid it overestimates the sublimation amount largely and caused the difference with the previous research.. Further, the accuracy of the MERRA-2 data needs to be indicated because they are also the key for the following estimates. Since a number of AWSs exist over the Antarctica, comparisons with these measurements must be done, or at least be referred, and show the MERRA-2 data are precise enough as the input data.

This is a very good point and was also noted by other reviewers. There is really nothing we can do about the modification of the layer temperature and moisture by the blowing snow sublimation other than using a model that incorporates these processes. Even if we did so, there is no guarantee that when CALIPSO identifies a blowing snow layer, the model would know it is there and have correspondingly modified its temperature and moisture. Having said that, I believe this is an area in need of more research. Please note we have added sections 2.1 to describe the MERRA-2 data and compare with observations. We have also added section 4.1 which is a sensitivity study on the effect of increasing moisture on blowing snow sublimation. In section 4.1 (and section 5 – conclusions) we note this as a shortcoming and have modified our total error calculation to +/- 50%.

Line 347: Since the sublimated vapor does not always contribute on the sea level rise, I suppose this part is meaningless.

Agreed. This has been removed.

Line 351: If this really the case, it is of great interest. Since the authors have all the ingredients (factors) to deduce the sublimation amount, the reason which brought the annual change can be conjectured, I believe. More detailed considerations are recommended.

Another Reviewer suggested that this be removed as the time series is rather short

Line 413: When we discuss the amount of sublimation and transport quantitatively, the blowing snow from the surface to 30 m, where this satellite technique is not applicable, cannot be neglected. As far as I know, the flux increases with decreasing height on the contrary to Fig. 2. Although the particles may suspend as high as 300 m, most of the transport concentrate within 0 to 30 m layer. It should be also taken into account in the error analysis. Recently, Huang et al.(2016) published the manuscript on Atmospheric Chemistry and Physics and discussed the impacts of moisture transport on drifting snow sublimation in the saltation layer. Thickness of the saltation layer is less than 0.1 m in usual, nevertheless, they say that the blowing (drifting) snow sublimation is important on the distribution and mass-energy balance of snow cover.

Thank you for this information. We have added text to the revised paper and cited Huang et al., 2016 on line 458 of revised text.

Line 458: It the estimates of the error amounted to large as 40 %, the conclusion that 'the sublimation amount is about the twice the one of the previous studies' on line 478 is not always the case.

We have changed this wording

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 the 10 m 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 for the period 2006 - 2016 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 (Cloud-Aerosol Lidar with Orthogonal Polarization) lidar aboard the CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation) satellite. The blowing snow events identified by CALIPSO and meteorological fields from MERRA-2 are used to compute the blowing snow 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 the 2006 – 2016 Antarctica average integrated blowing snow sublimation is about 393 ± 196 Gt yr^{-1} which is 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 3.7 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 (P), surface evaporation and sublimation (E), surface melt and runoff (M), blowing snow sublimation (Q_s) and snow transport (Q_t). 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.

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

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

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

Considering that the accuracy of model data is questionable over Antarctica, and the complicated factors that govern the onset of blowing snow, it is difficult to assess the accuracy 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 20-30 m or more in thickness. Thus, drifting snow which is confined to 10 m or less and occurs frequently over Antarctica would not be reliably detected. The signal from these layers is likely inseparable from the strong ground return. More information on the blowing snow detection algorithm can be found in Palm et al., (2011).

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

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

2.1 MERRA-2 Reanalysis Data

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

We understand that MERRA-2 does not include the effects of blowing snow sublimation on atmospheric moisture and thus may have a dry (and possibly warm) bias. MERRA-2 temperature and moisture has not been evaluated over Antarctica but in this section we present a comparison of MERRA-2 temperature and moisture at 2 m height with a surface station and two AWS sites. In Fig. 2 are data from two AWS sites (chosen at random) comparing MERRA-2 and AWS 2 m temperature and relative humidity with respect to ice (RHice). In both cases MERRA-2 is, on average, colder and moister than the observations (about 1 °C and 2%, respectively). Figs. 3a and 3b show MERRA-2 data compared to the surface station at Princess Elisabeth for data taken over 2009-2015. Here also MERRA-2 is colder and moister (about 4 °C and 6-8%, respectively). Also shown in Figs. 3c, d and e are the annual mean relative humidity at 2 meters above the surface in 2015 estimated by MERRA-2, ERA-Interim, and AMPS-Polar WRF showing that MERRA-2 is considerably moister than ERA-Interim or AMPS. From these comparisons it is likely that MERRA-2 does not exhibit a dry or warm bias and is rather slightly cold and moist.

2.2 Sublimation

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

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

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

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

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

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

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

Where $\beta(z)$ is the CALIPSO measured attenuated calibrated backscatter at height *z* (30 m resolution), $\beta_m(z)$ is the molecular backscatter at height *z* and *S* is the extinction to backscatter ratio (25). Here $\beta(z)$ represents the atmospheric backscatter profile through the blowing snow layer. Both $\beta_m(z)$ and $\beta(z)$ have units of m^{-1} sr^{-1}. We found that the values of N(z) obtained from Eq. (2) for the typical blowing snow layer range from about 5.0×10^4 to 1.0×10^6 particles per cubic meter. This is consistent with the blowing snow model results of Dery and Yau (2002) and the field observations of Mann et al., (2000). A plot of the average particle density for the blowing snow layer in Fig. 1 is shown in Fig. 5. Note that the decrease in particle number density below about 75 m is most likely due to attenuation of the lidar signal as it propagates through the layer. We did not attempt to correct for this and the overall effect is an under estimation of the particle density in this region.

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:

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

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

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

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:

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

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

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

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 ν 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}):

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

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

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:

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

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:

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

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

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

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 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).

2.3 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:

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

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

Fig. 5 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 blowing snow frequencies are located. Sublimation is highly dependent on the air temperature and relative humidity. For a given value of the blowing snow mixing ratio (q_b), the warmer and drier the air, the greater the sublimation. In Antarctica, it is considerably warmer along the coast but one would not necessarily conclude that it is drier there. However, other authors have noted that the katabatic winds, flowing essentially downslope, will warm and dry the air as they descend (Gallee, 1998, and others). We have examined the MERRA-2 relative humidity (with respect to ice) and indeed, according to the model, it is usually drier along the coast. The model data often shows 90 to 100 % (or even higher) relative humidity for interior portions of Antarctica, while along the coast it is often 70 % or less. It should be noted, however, that this model prediction has never been validated through observations. The combination of warmer and drier air makes a big difference in the sublimation as shown in Fig. 6. For a given relative humidity the sublimation can increase by almost a factor of 100 as temperature increases from -50 to -10 °C. For temperatures greater than -20 °C, sublimation is very dependent on relative humidity, but this dependence lessens somewhat at colder temperatures. Continental interior areas with very high blowing snow frequency that approach 75 % (like the Mega Dune region in East Antarctica) exhibit fairly low values of sublimation because it is very cold and the model relative humidity is high.

Fig. 7 shows the annual total sublimation for years 2007–2015. It is evident that the sublimation pattern or magnitude does not change much from year to year. The overall spatial pattern of sublimation is similar to the model prediction of Dery and Yau, (2002) with our results showing noticeably greater amounts in the Antarctic interior and generally larger values near the coast. As previously noted, most sublimation occurs near the coast due mainly to the warmer temperatures. The areas of sublimation maximums near the coast are consistently in the same location year to year, indicating that these areas may experience more blowing snow episodes and possibly more precipitation (availability of snow to become airborne). It is interesting to compare the sublimation pattern with current estimates of Antarctic precipitation. Precipitation is notoriously difficult to quantify over Antarctica due to the scarcity of observations and strong winds producing drifting and blowing snow which can be misidentified as precipitation. Precipitation is often measured by looking at ice cores or is estimated by models. But perhaps the most complete (non-model) measure of Antarctic precipitation come from the CloudSat mission. Palerme et al., (2014) used CloudSat data to construct a map of Antarctic precipitation over the entire continent (north of 82 S). They showed that along the East Antarctic coast and slightly inland, precipitation ranges from 500 to 700 mm swe yr^{-1} and decreases rapidly inland to less than 50 mm yr^{-1} in most areas south of 75 S. Their precipitation pattern is in general agreement with the spatial pattern of our sublimation results and the magnitude of our sublimation estimates is in general less than the precipitation amount, with a few exceptions. These occur mostly inland in regions of high blowing snow frequency such as the Megadune region and in the general area of the Lambert glacier. In these regions, our sublimation estimates exceed the CloudSat yearly precipitation estimates. When this occurs, it is likely that either the precipitation estimate is low or the sublimation estimate is too high. Otherwise it would indicate a net negative mass balance for the area unless transport of snow into the region accounted for the difference.

Table 1 shows the average sublimation over all grid cells in snow water equivalent and the integrated sublimation amount over the Antarctic continent (north of 82S) for the CALIPSO period in Gt yr^{-1}. Note that the 2006 data include only months June–December (CALIOP began operating in June, 2006) and the 2016 data are only up through October, and do not include the month of February (CALIOP was not operating). To obtain the integrated amount, we take the year average swe (column 1) multiplied by the surface area of Antarctica north of 82S and the density of ice. The average integrated value for the 9 year period 2007–2015 of 393 Gt yr^{-1} is significantly greater than (about twice) values in the literature obtained from model parameterizations (Lenaerts 2012b). Note also that this amount does not include the area poleward of 82S, the southern limit of CALIPSO observations. If included, and the average sublimation rate over this area were just 4 mm swe per year, this would increase the sublimation total by 10 Gt yr^{-1}. 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 (and grounded ice shelves) 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. 8 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 (16,662 km^2) 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 Fig. 8,, 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. 8 are two CALIPSO tracks (yellow lines) and their associated retrieved blowing snow backscatter (upper and lower images of CALIOP backscatter). Note that the yellow track lines are drawn only where blowing snow was detected by CALIOP and that not all the CALIOP blowing snow detections are shown. The green dots denote the coastline. Plainly seen along the coast near longitude 145–150E is blowing snow being carried off the continent. In this case, topography might have played a role to funnel the wind in those specific areas. Fig. 9 shows a zoomed in image of this area with the red lines indicating the approximate position of the coastline. Also note that, as evidenced by the times of the MODIS images, this transport began on or before October 13 at 23:00 UTC and continued for at least 7 hours. This region is very close to the area of maximum sublimation seen in Fig. 5

and shown to be quite stable from year to year in Fig. 7. Undoubtedly, this continent to ocean transport also occurs in other coastal areas of Antarctica and most often during the dark winter (when MODIS could not see it).

In an attempt to better understand the magnitude of this phenomena, we have computed the amount of snow mass being blown off the continent by computing the transport at 342 points evenly spaced (about 60 km apart) along the Antarctic coast using only the v component of the wind. If the v component is positive, then the wind is from south to north. The transport (Eq. (12) using only the v wind component) is computed at each coastal location and then summed over time at that location. The resulting transport is then summed over each coastal location to arrive at a continent-wide value of transport from continent to ocean. Of course this assumes that the coastline is oriented east-west everywhere. This is true of a large portion of Antarctica but there are regional exceptions. Thus we view the results shown in Table II to be an upper limit of the actual continent to ocean transport. Evident from Table 2 is that most of the transport for East Antarctica occurs in a relatively narrow corridor, with on average over half (51 %) of the transport occurring between 135E and 160E. This is obviously due to the very strong and persistent southerly winds (see Fig. S13) and high blowing snow frequency in this region and is consistent with the conclusions of Scarchilli et al., (2010). In West Antarctica, an even greater fraction (60 %) of the transport off the coast occurs between 80W and 120W.

In Fig. 10 we show the magnitude of blowing snow transport for the 2007–2015 timeframe in Mt km^{-1} yr^{-1} as computed from Eq. (12). 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. 10) exceed about 3 x 10^6 tons of snow per km per year. In Figs. S1 and S2 we display the MERRA-2 average 10 m wind speed and direction for the years 2007–2015. By inspection of Figs. S1 and S2 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 and within 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 the attenuation of the lidar signal above the blowing snow layer (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. The effect of attenuation within the layer is unaccounted for here and will also reduce the amount of calculated blowing snow sublimation.

With regard to 5) above, the method presented here cannot reliably detect blowing snow layers less than about 30 m thick. Therefore, sublimation associated with these layers is not accounted for. Other studies have shown that drifting snow sublimation within the salutation layer can be very significant (Huang et al., 2016). 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.

4.1 Sensitivity Analysis

A major limitation of this work is the uncertainty inherent in the meteorological data used for obtaining the temperature and moisture within the blowing snow layer. Re-analyses like MERRA-2 do not have the vertical or horizontal resolution to enable an accurate description of the temperature and moisture profile through the blowing snow layer. Also, as mentioned in section 2.1, MERRA-2, or more accurately the GEOS-5 model on which it is based, does not incorporate the effects of blowing snow sublimation on the moisture within the layer. Even so, we have already shown that MERRA-2 is moist compared to surface observations and to other models. Thus we do not feel that using the MERRA-2 moisture will cause a large overestimation of blowing snow sublimation. However, we feel compelled to examine the effects of mositure on the calcualted sublimation. To demosnstrate this we have taken one CALIPSO track with blowing snow and plotted the MERRA-2 humidity (wrt ice) and the calculated blowing snow sublimation along the track. We then increased the moisture amount by 5 and 10% to see the effect on the calculated sublimation. The temperature was not changed. In Figs. 12b – 12d the MERRA-2 relative humidity is the dark solid line, MERRA-2 temperature is the dotted line and the calculated blowing snow sublimation is the thin black line. The temperature and moisture shown are the MERRA-2 averages through the blowing snow layer. Figure 12b shows the unperturbed MERRA-2 moisture and the resulting blowing snow sublimation (integrated through the layer). In Fig 12c and 12d we have increased the MERRA-2 relative humidity by 5 and 10%, respectivly. The effect on the average blowing snow sublimation is marked. A 10% increase in relative humidity produces nearly a 30% reduction in the calculated blowing snow. This exercise demonstrates the non-linear effect of the moisture level on the calculated sublimation.

If we assume then that the error in moisture is 10%, we must accept that the resulting blowing snow sublimation could be 30% too high. But is that realistic, given the fact that the MERRA-2 data were shown to be moist compared to observation and other models? We do not think so. Rather we take the error in MERRA-2 moisture to be 5%. This produces an 18% over estimation of sublimation (Fig. 11b compared to Fig. 11c). This error must be combined with other errors such as extinction, particle radius and temperature. Here we assume the extinction error to be 20 %, the particle radius error 10 % and the temperature error 5%. In Eq. (6) these terms are multiplicative. The total error in sublimation is then:

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

This indicates that the sublimation values derived in this work should be considered to have an error bar of ±50 %. 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 assume wind speed has an error of 20 %, extinction 20 % and particle size 10 %, the total error in transport is:

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

5. Summary and Discussion

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

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

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

The overall spatial pattern of blowing snow sublimation is consistent with previous modelling studies (Dery and Yau, 2002 and Lenearts et al., 2012a). However, we find the Antarctic continent-wide integrated blowing snow sublimation to be larger than previous studies such as Lenaerts et al., (2012a) (393 ±196 vs roughly 190 Gt yr^{-1}), even though the observations include only the area north of 82° S. The maximum in sublimation is about 250 (±125) mm swe per year near the coast between longitudes 140E and 150E and seems to occur regularly throughout the 11 year data record. There are a number of reasons for the higher sublimation values in this study compared to prior estimates. 1) The depth of the layer: the average blowing snow layer depth as determined from the CALIOP measurements is 120 m. Layers as high as 200 - 300 m are not uncommon. It is likely that models such as those cited above do not always capture the full depth of blowing snow layers, thus producing a smaller column-integrated sublimation amount. 2) We only compute sublimation from blowing snow layers that are known to exist (meaning they have been detected from actual backscatter measurements). Models, on the other hand, must infer the presence of blowing snow from pertinent variables within the model. The existence of blowing snow is not easy to predict. It is a complicated function of the properties of the snowpack, surface temperature, relative humidity and wind speed. Snowpack properties include the dendricity, sphericity, grain size and cohesion, all of which can change with the age of the snow. In short, it is very difficult for models to predict exactly when and where blowing snow will occur, much less the depth that blowing snow layers will attain. 3) The lack of blowing snow physics within the MERRA-2 reanalysis. This produces perhaps the largest uncertainly in the derived results. It was shown that MERRA-2 is slightly colder and moister than some surface measurements and moister compared to other re-analyses. However, given the limited number of comparisons, a definitive conclusion on the accuracy of MERRA-2 data cannot be drawn. Since the model on which MERRA-2 re-analysis is based (GEOS-5) does not include blowing snow (and thus blowing snow feed backs on moisture and temperature), it is likely that our estimates of blowing snow sublimation are probably too high. However, the fact that we do not include blowing snow layers less than 30 m in depth and are not able to detect blowing snow beneath thick clouds layers means that we are missing potentially important contributions to sublimation. An addition, the retrieved blowing snow number density below about 80 m is probably too low for layers greater than 120 m in depth because of lidar signal attenuation. This will act to erroneously reduce the calculated sublimation. While we estimate an upper limit on the blowing snow sublimation error as 50%, we believe that the error is considerably less than that.

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

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

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

Year	Average Sublimation (mm we)	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***

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

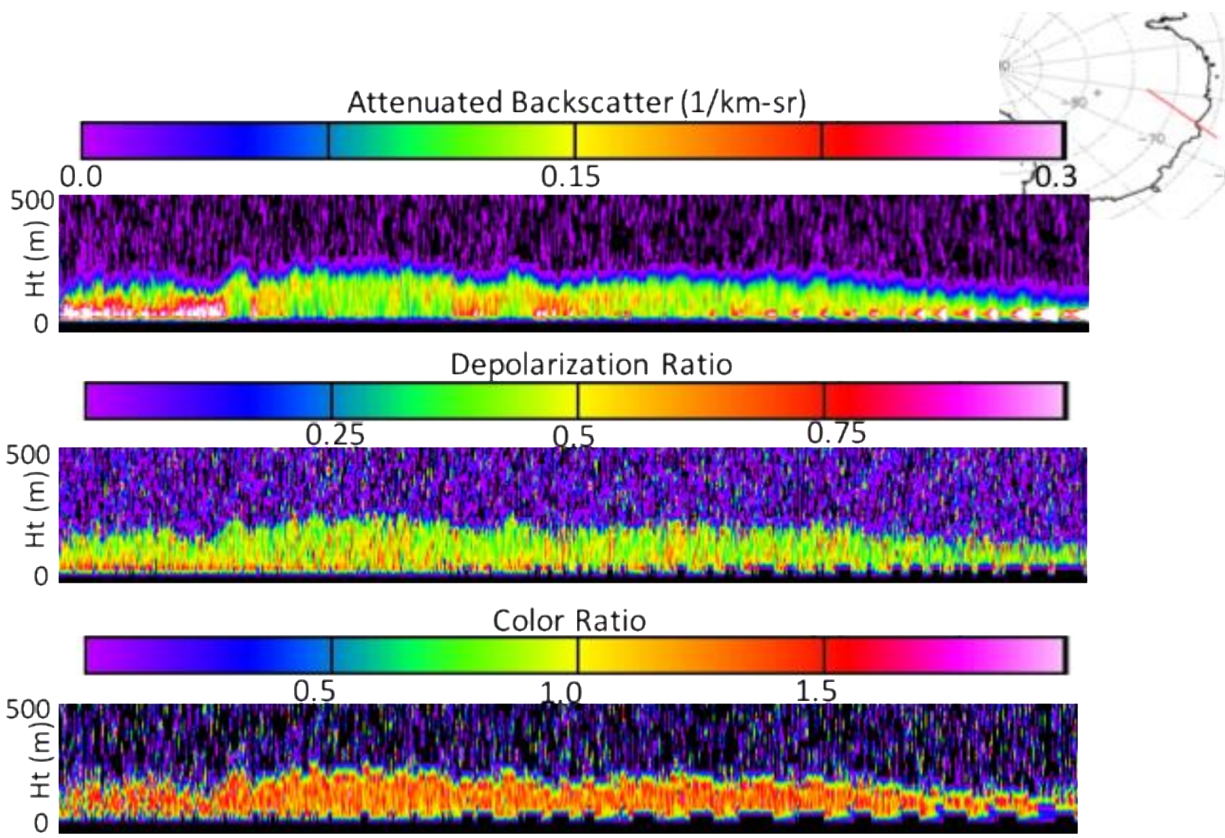

Figure 1. A typical Antarctic blowing snow layer as measured by CALIPSO on May 28, 2015 at 17:08:41 – 17:11:33 UTC. Displayed (from top to bottom) are the 532 nm calibrated, attenuated backscatter, the depolarization ratio at 532 nm, and the color ratio (1064 nm / 532 nm).

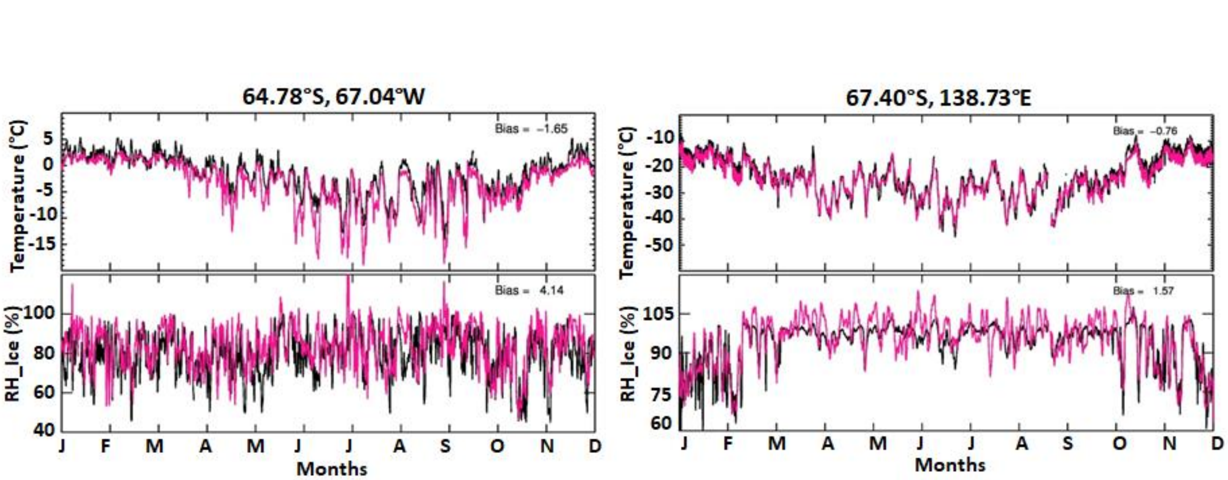

Figure 2. A comparison of 2 m MERRA-2 temperature and moisture (pink) with measurements from 2 AWS stations for 2016.

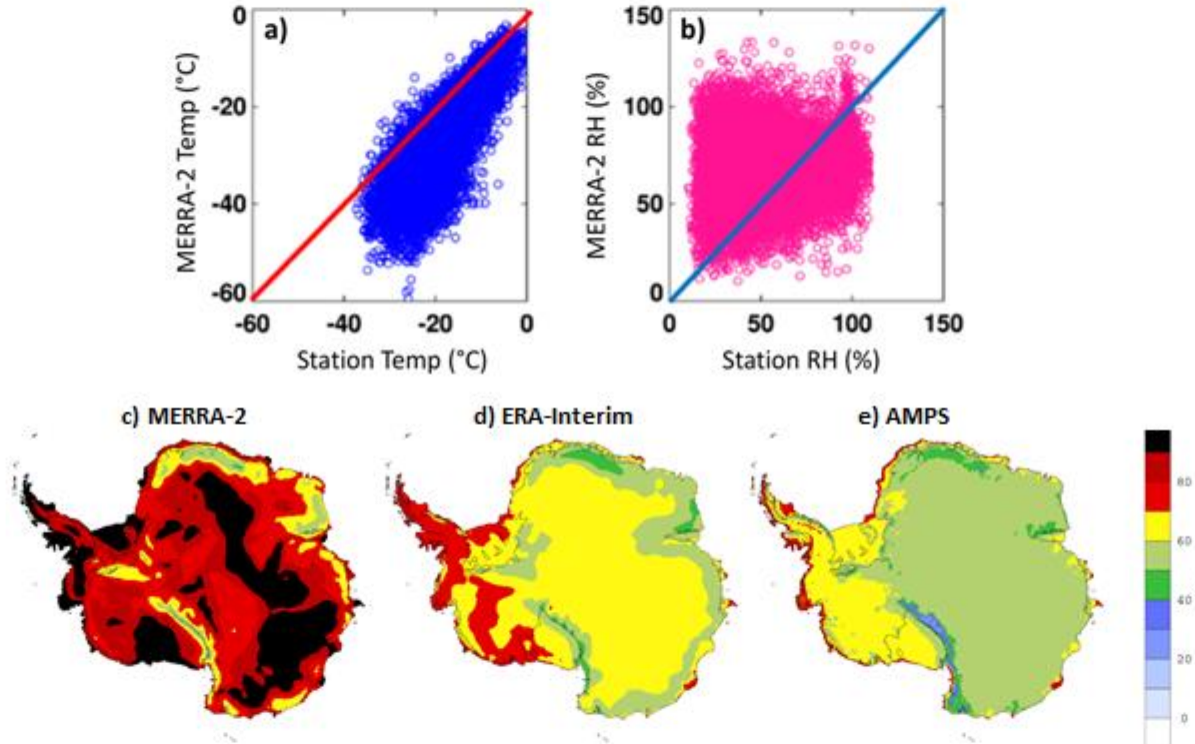

Figure 3. (a) A comparison of MERRA-2 2 m temperature and (b) relative humidity with respect to ice for the period 2009-2015 at Princess Elisabeth Station, Antarctica. (c-e) Annual mean relative humidity at 2 meters above the surface in 2015 estimated by (c) MERRA-2, (d) ERA-Interim, and (e) AMPS-Polar WRF.

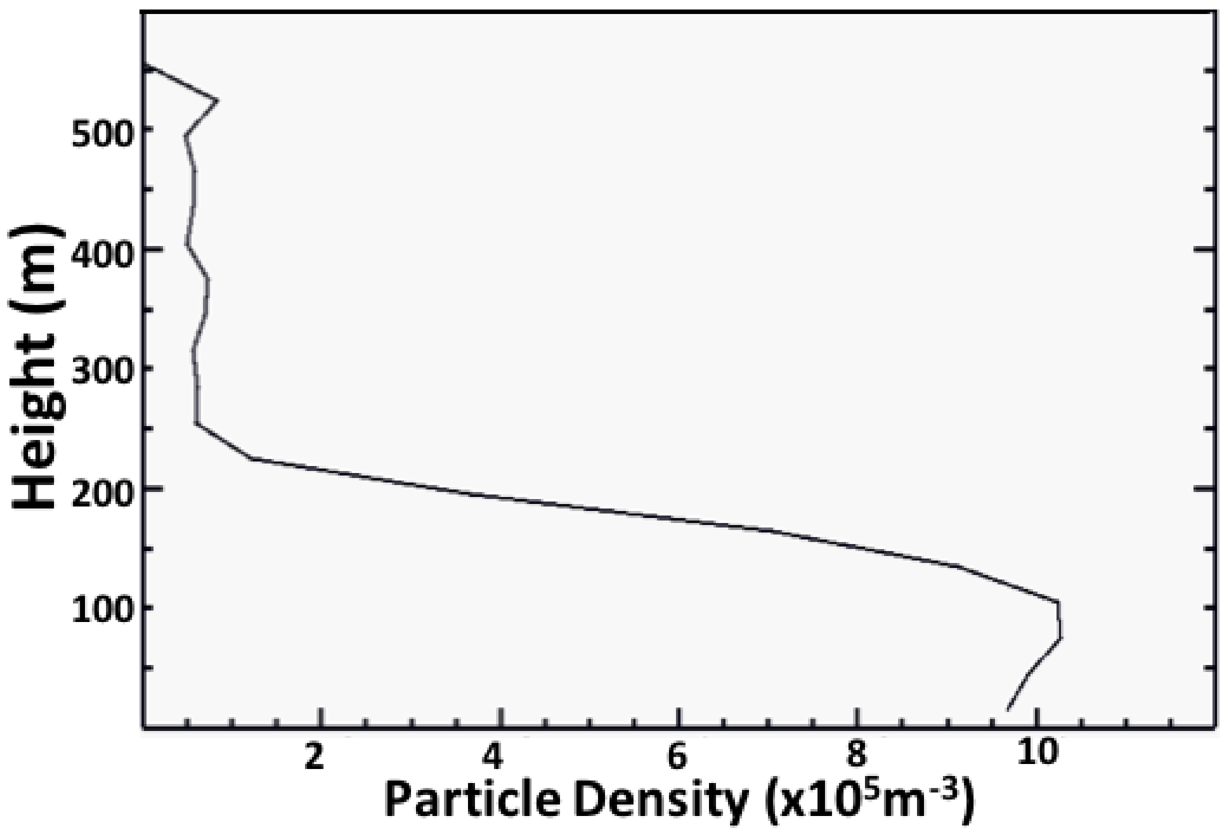

Figure 4. Average particle density profile (Eq. 2) through the blowing snow layer shown in Fig. 1.

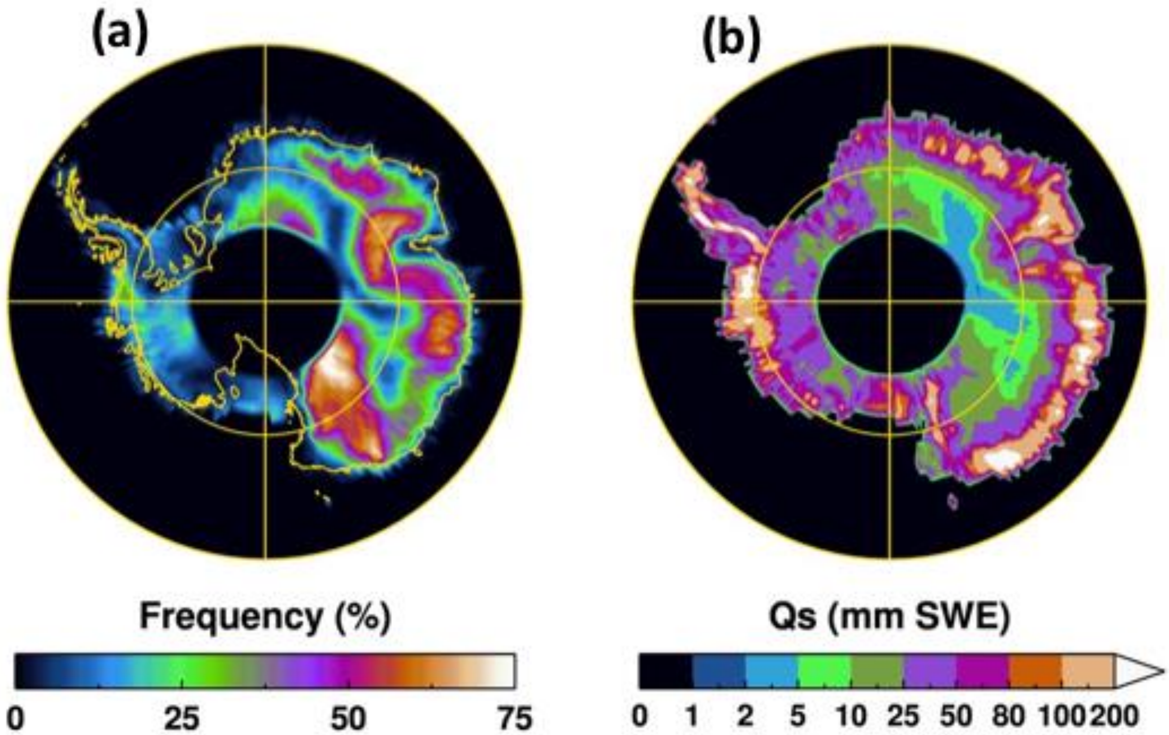

Figure 5. (a) The average April through October blowing snow frequency for the period 2007–2015. (b) The average annual blowing snow sublimation for the same period as in (a).

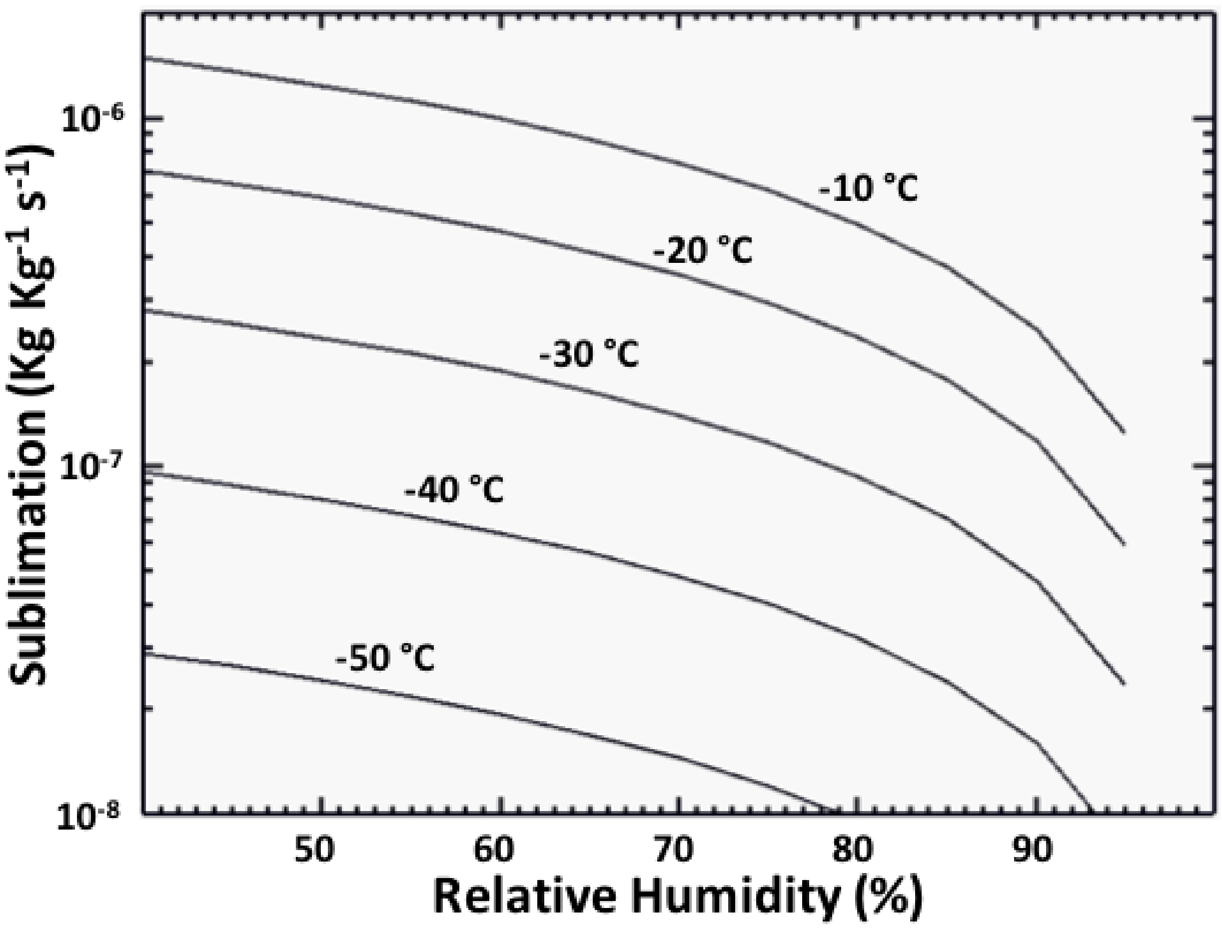

Figure 6. Computed blowing snow sublimation rate using Eqs. (3) and (4) as function of relative humidity for varying air temperatures. The particle density value used in Eq. (3) was 10^6 m^{-3} which corresponds to a blowing snow mixing ratio (qb) of 4.7x10-5 kg kg^{-1}

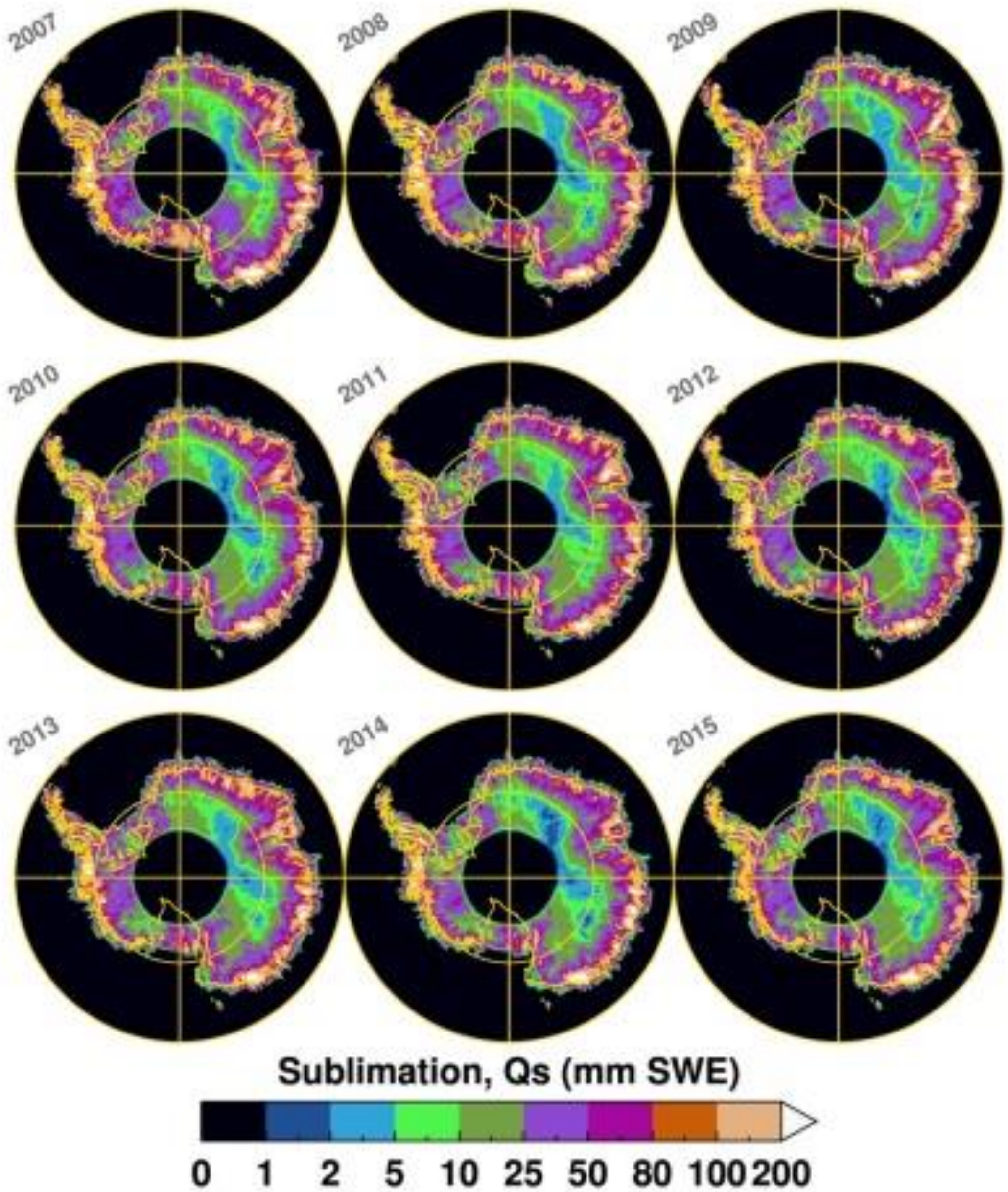

Figure 7. Blowing snow total sublimation over Antarctica by year for 2007–2015.

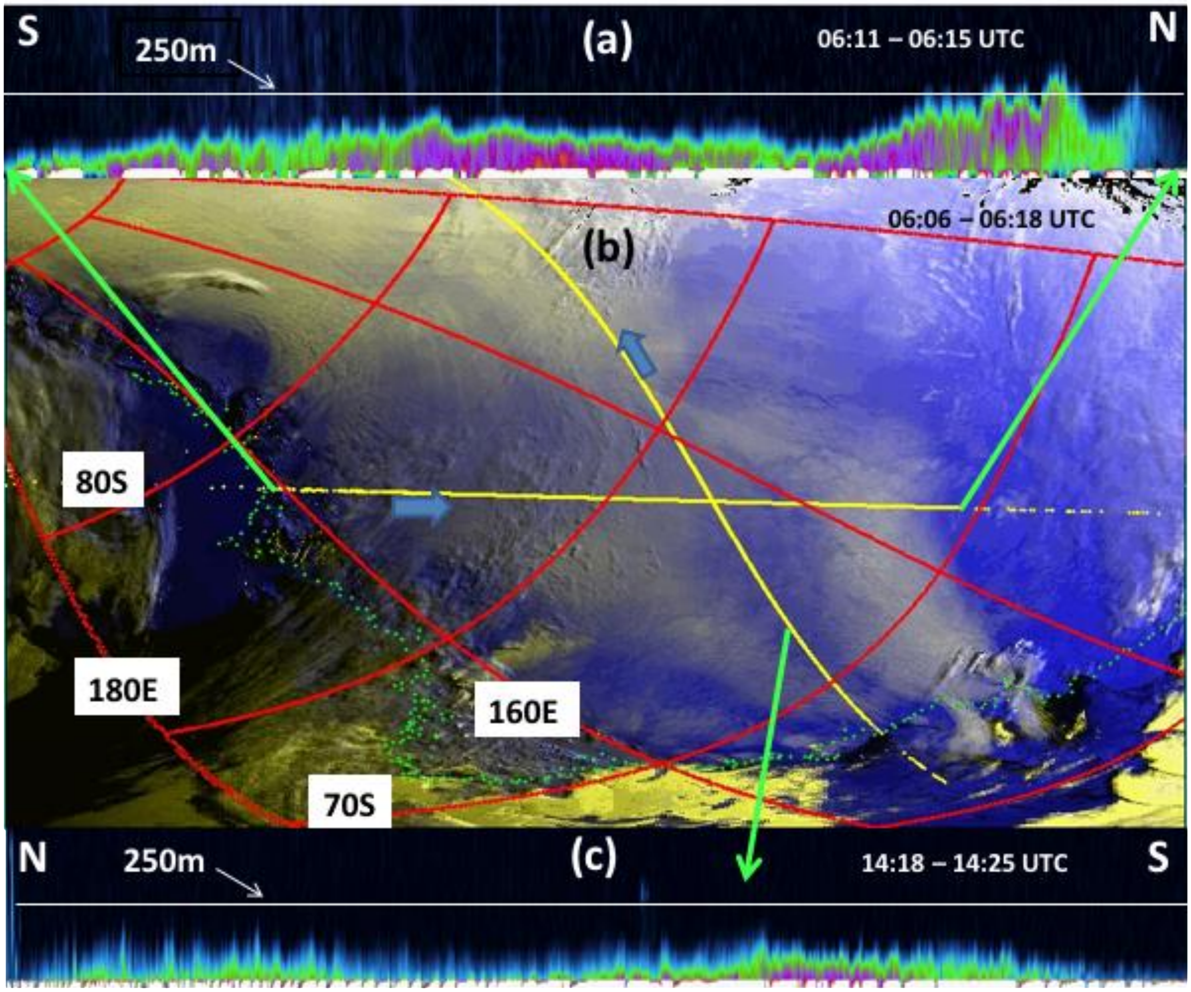

Figure 8. A large blowing snow storm over Antarctica with blowing snow transport from continent to ocean on October 14, 2009. (a) CALIOP 532 nm attenuated backscatter along the yellow (south to north) line bounded by the green arrows as shown in (b) at 06:11 – 06:15 UTC. (b) MODIS false color image at 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.

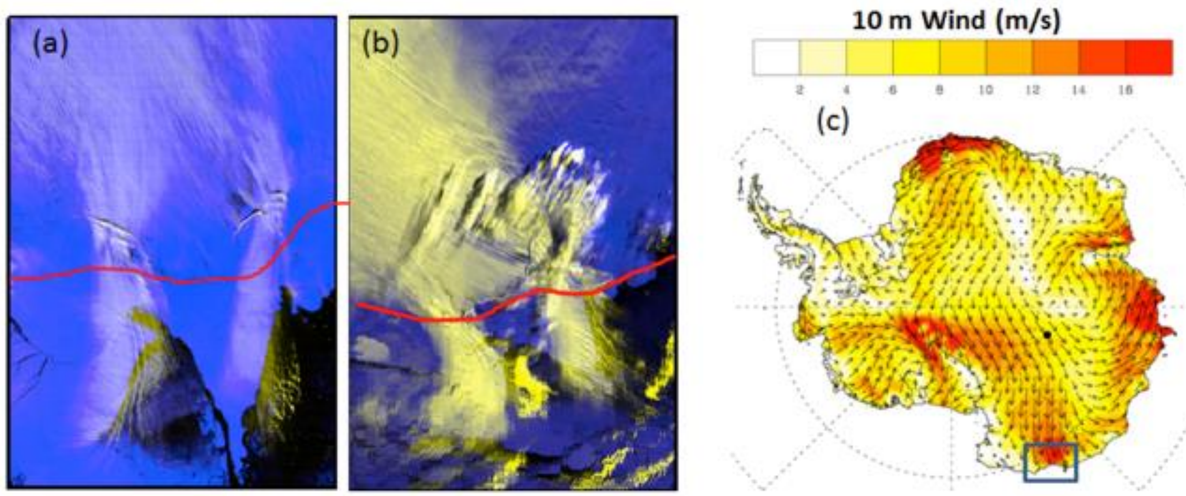

Figure 9. (a) MODIS false color image on October 13, 2009, 23:00 UTC and (b) October 14, 2009, 06:16 UTC. The red line is the approximate position of the coastline. (c) The 10 m wind speed from the AMPS model (Antarctic Mesoscale Prediction System) for October 14, 2009. The area covered by the MODIS images is roughly that indicated by the blue box in (c).

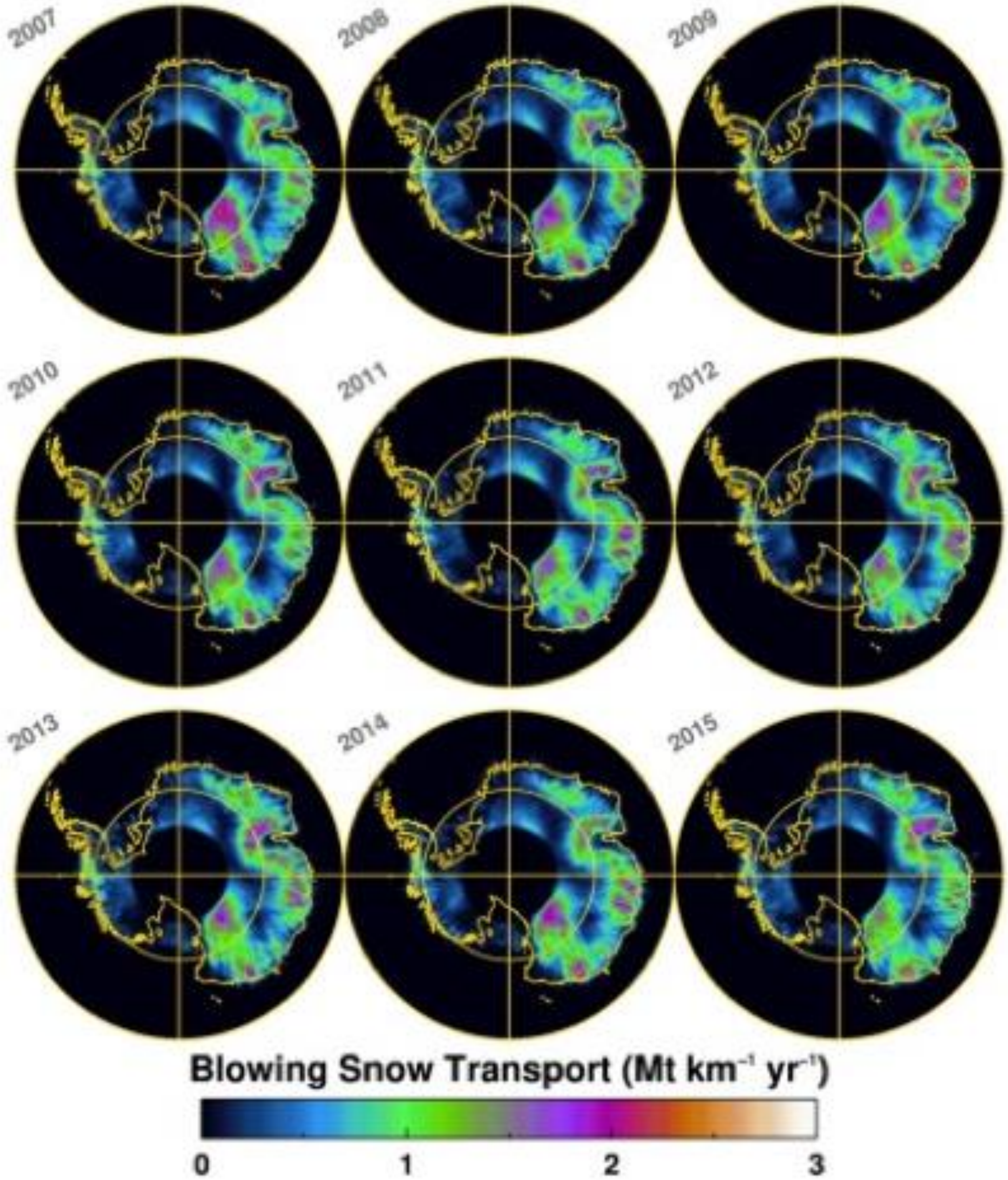

Figure 10. The magnitude of blowing snow transport over Antarctica integrated over the year for years 2007–2015.

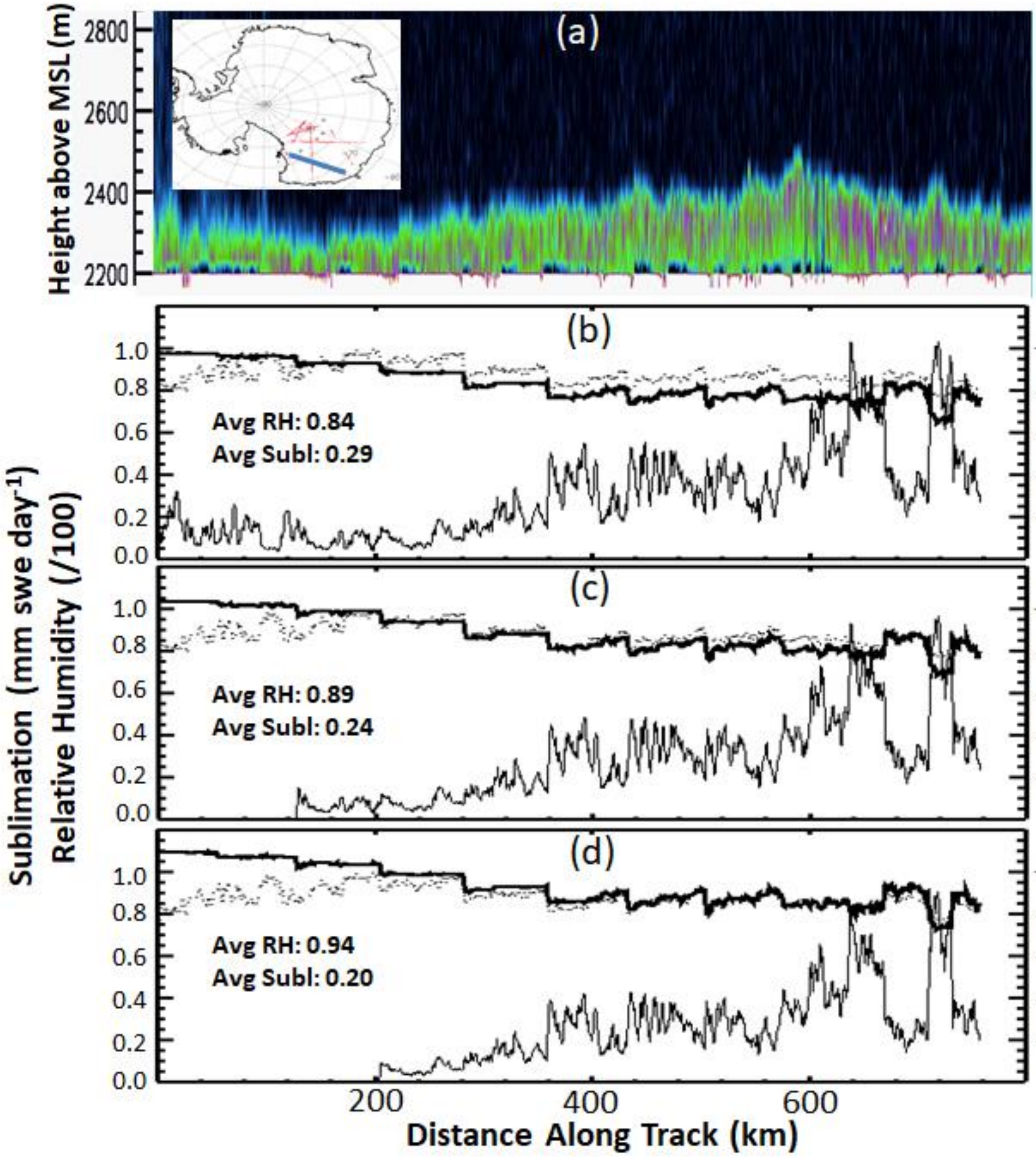

Figure 11. (a) CALIPSO backscatter showing blowing snow layer along the blue line in the map inset on 10/12/2010 at 05:51 UTC. (b) Average MERRA-2 moisture (dark black line), temperature (dotted line) and calculated sublimation through the blowing snow layer along the CALIPSO track. (c and d) Same as in (b) but increasing MERRA-2 humidity by 5 and 10%, respectively.