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New Perspectives on Blowing Snow in Antarctica and Implications for Ice Sheet Mass Balance

Stephen P. Palm, Yuekui Yang and Vinay Kayetha

Abstract

Blowing snow processes commonly occur over the earth's ice sheets and snow covered regions when near surface wind speed exceeds a threshold value. These processes play a key role in the sublimation and redistribution 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. Until recently, most of our knowledge of blowing snow was obtained from field measurements or modeling. Recent advances in satellite remote sensing have enabled a more complete understanding of the nature of blowing snow. Using 12 years of satellite lidar data, climatology of blowing snow frequency has been compiled, showing the spatial and temporal distribution of blowing snow frequency over Antarctica. Other characteristics of blowing snow such as backscatter structure and profiles of temperature, relative humidity, and winds through the layer are explored. A new technique that uses direct measurements of blowing snow backscatter combined with model meteorological reanalysis fields to compute the magnitude of blowing snow sublimation and transport is also discussed.

Keywords: blowing snow, climatology, sublimation, transport, thermodynamic structure, CALIPSO

1. Introduction

Antarctica is the coldest and windiest place on earth. The ice sheet reaches a height of 4 km near the center of East Antarctica and slopes toward the coasts. The sloping terrain together with radiative cooling of the surface and lower atmosphere produces drainage flows known as katabatic winds. In some places, wind speeds can attain hurricane force and blow for days on end. It is not surprising, then, that the snow present in Antarctica is almost always airborne. Snow lifted from the surface and carried aloft by wind is known as drifting and blowing snow. Drifting snow is generally defined as airborne snow confined to a maximum height of 2 m. Once snow particles attain a height greater than 2 m, it is considered blowing snow. Blowing snow is important for a number of reasons including ice sheet mass balance [1, 2], the water budget of high latitude regions [3], and the reconstruction of

paleoclimate records from the physical and chemical records obtained from ice cores [4]. Interaction with blowing snow is a major factor for changes in surface ice characteristics, such as rifts, crevasses, ridges, sastrugi, etc. and in deposition of snow on sea ice [5–7].

Most of our understanding of blowing snow comes from field measurements at the surface or from numerical modeling. Numerous observations conducted in Antarctica have measured the properties of blowing snow such as particle size, number density, mass flux, and atmospheric conditions associated with blowing snow episodes [8–10]. Most such studies are made in drifting or blowing snow conditions where the snow particles are confined to shallow layers. Particle size ranges from 50 to 450 μm with the largest particles occurring near the surface. These studies also provided data on the wind speed required to initiate blowing snow. This wind speed, known as the threshold velocity, depends on the properties of the snow on the surface such as age, temperature, density, sphericity, and cohesiveness [1]. Generally, the threshold velocity ranges from about 5 to 8 m s^{-1} . All field measurements of blowing snow in the literature are made below a height of 10 m. However, blowing snow can frequently reach heights of 100 m or more and little or nothing is known about the properties of these deep blowing snow layers [11, 12]. This is mainly because when they occur, the conditions are too harsh to make measurements in the field.

Because of the lack of observations over Antarctica, the temporal and spatial frequency of blowing snow was not known until recently. With the advent of active satellite remote sensing (lidar), it was shown that blowing snow occurs more frequently than 70% of the time over large regions of Antarctica and can reach heights of 500 m [13].

2. Blowing snow detection from satellite lidar

Because of the scarcity of manned weather observation stations in Antarctica, the only practical way of obtaining information on blowing snow over the entire continent is from satellite remote sensing. While there have been passive sensors in polar orbit since the 1960s, it is very difficult to detect blowing snow from passive visible or infrared (IR) channels. In the visible, the blowing snow is indistinguishable from the underlying snow surface, and in the IR, there is generally not enough temperature contrast between the surface and blowing snow to make the latter visible. In 2003, ICESat carried the Geoscience Laser Altimeter System (GLAS) into polar orbit [14]. GLAS had both altimetry and atmospheric channels and was the first satellite lidar to study the earth's surface and atmosphere. Atmospheric data from GLAS were used to develop a technique to detect blowing snow layers from satellite lidar data. In 2006, the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite was launched into polar orbit [15]. CALIPSO was specifically designed to study the earth's atmosphere using a multi-wavelength lidar. The data from CALIPSO have been used to study the climatology of blowing snow over Antarctica for the period 2006–2017 [16].

A lidar instrument transmits short pulses of (usually) visible or near IR laser light into the atmosphere, and a very small portion of the light is scattered directly backward from molecules, aerosols, and clouds. A telescope and associated detector electronics record the backscattered light into time-resolved “bins” usually 30–75 m in length. For satellite lidars like CALIPSO, the resulting backscatter profile extends from about 40 km in altitude to a kilometer or two below the ground. For blowing snow detection, only the backscatter data in the lowest 500 m above the surface are of interest. Since blowing snow is rooted at the ground, the algorithm first detects

the large ground return signal and then interrogates the bins immediately above for elevated levels of backscatter. If the ground return is not found, then it is impossible to detect blowing snow from the backscatter profile. The lack of a ground return indicates that overlying cloud layers have attenuated the lidar beam and reduced the backscatter signal from the ground to near zero. This occurs when the overlying cloud layers have an optical depth of about 3 or greater. In general, over East Antarctica, this only happens near the coast. In the interior, cloudiness is normally quite low, and what clouds are present are usually optically thin.

When the ground return is found, the backscatter level of the lidar profile bin directly above the ground is compared to a threshold value (about 10 times the local value of the molecular backscatter). If it is greater than the threshold and the 10 m wind speed (which is on the CALIPSO data product and is obtained from the GEOS-5 global analysis product) is greater than 4 m s^{-1} , then a scattering layer has been detected. The search then continues upward for the location of the top of the scattering layer. This is defined as the bin immediately below two consecutive bins that have signal levels less than a value of about twice the local value of molecular backscatter. A few tests are made on the scattering layer in an effort to remove diamond dust from the detections. Diamond dust is fairly common in Antarctica, especially over the high Antarctic Plateau in winter and can be many km in thickness and extend to the ground. To reduce misclassification of diamond dust as blowing snow, the top of the layer cannot exceed 500 m in height (above the ground) and the backscatter level within the layer must decrease with height. If these tests are passed, then the layer is assumed to be blowing snow. Finally, the optical depth of the blowing snow layer is estimated using an extinction to backscatter a value (lidar ratio) of 25 sr that is the typical value of ice crystals found in cirrus clouds [17].

3. Characteristics of blowing snow

Blowing snow is a dynamic boundary layer phenomenon, which is only slowly revealing its many secrets. The satellite lidar measurements discussed in Section 2 have increased our knowledge significantly. **Figure 1** shows examples of blowing snow as seen by CALIPSO. Each panel of **Figure 1** is a separate CALIPSO track over East Antarctica. They show the 532-nm attenuated backscatter as measured by the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) lidar and represent typical wintertime blowing snow events in Antarctica. Generally, the average height of the layers is about 120 m, but they can range from just a few meters to over 400 m in depth. The lidar attenuated backscatter values range from about $1.0 \times 10^{-5} \text{ m}^{-1} \text{ sr}^{-1}$ near the top of the layer to $5.0 \times 10^{-4} \text{ m}^{-1} \text{ sr}^{-1}$ in the lower third of the layer. This is consistent with blowing snow models that indicate both the particle size and number density decrease with height. On average, the optical depth of blowing snow layers (not shown) is about 0.2, but can range from 0.05 to 1.0 [13]. A defining feature of blowing snow as visualized by lidar is the cellular-like structure of the backscatter. Close inspection of **Figure 1** will reveal relatively small (2–3 km), regularly spaced cell-like structures. This is a very common feature of blowing snow layers and seems to suggest that convection is occurring within the layer.

A recent discovery aided by active (lidar) and passive satellite measurements is the size of blowing snow storms in East Antarctica. Evidence shows that these storms can cover an enormous area and last for a number of days [13]. An example of such a storm is shown in **Figure 2**, which is a MODerate resolution Imaging Spectroradiometer (MODIS) false color (RGB = 2.1, 2.1, 0.85 μm) image of a large

area of blowing snow covering an area about the size of Texas (695,622 km²) in East Antarctica on October 14, 2009. This false color technique is the best way to visualize blowing snow from passive sensors. The one drawback is that sunlight is

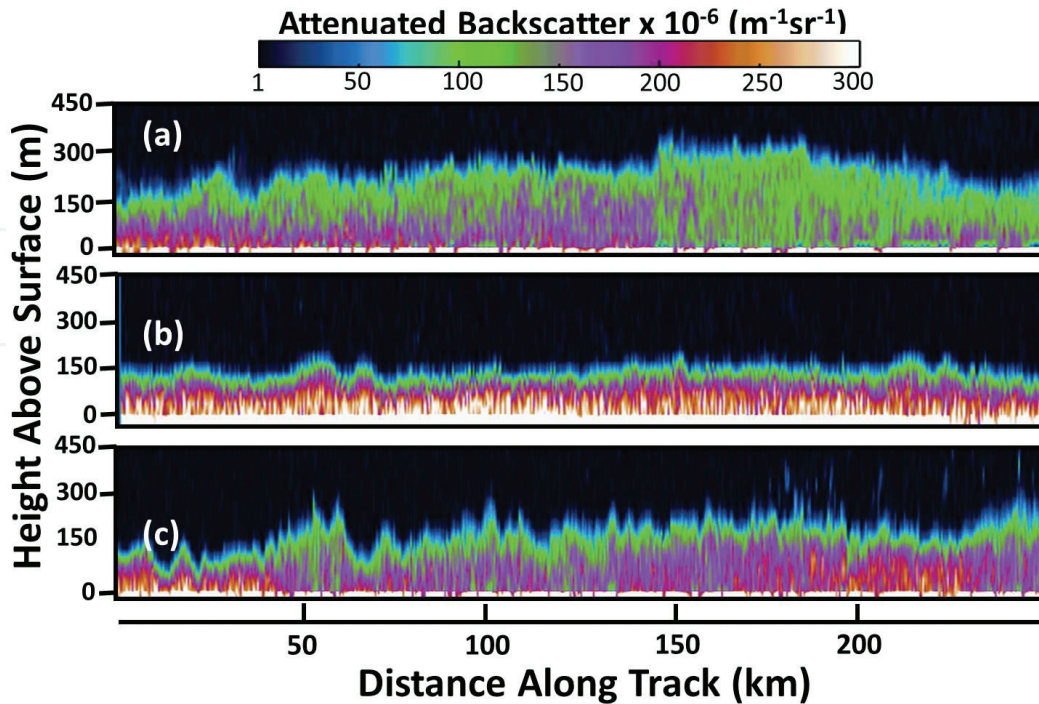


Figure 1. CALIOP 532-nm attenuated backscatter of typical wintertime blowing snow layers over East Antarctica for (a) September 18, 2007, 09:10 UTC, (b) September 7, 2010, 14:14 UTC, and (c) October 10, 2010, 12:34 UTC.

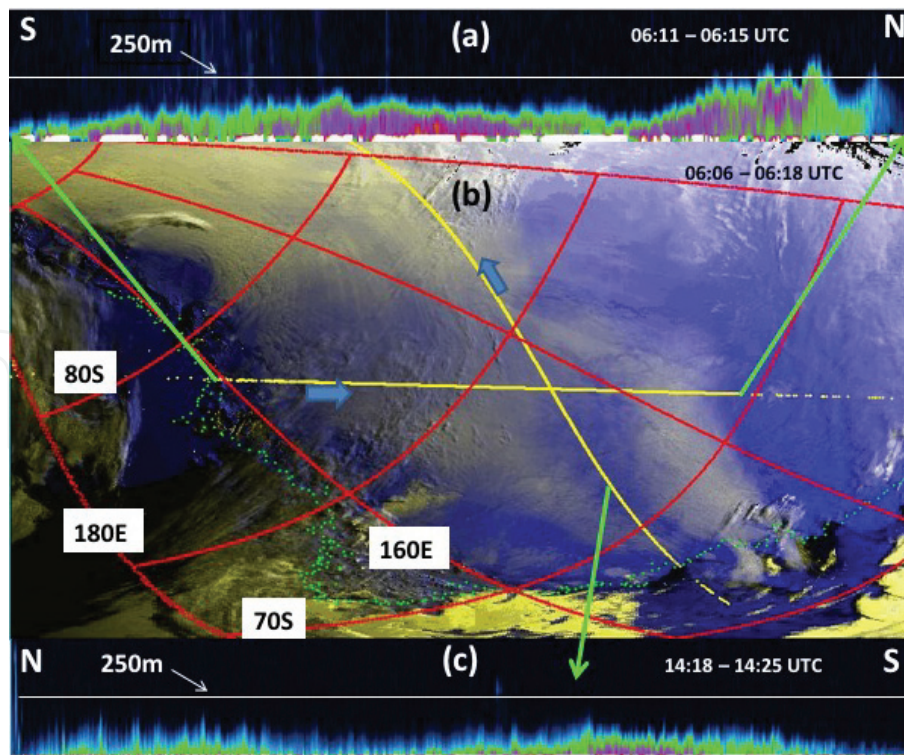


Figure 2. A large blowing snow storm over Antarctica with blowing snow transport from continent to ocean on October 14, 2009. (a) CALIOP 532-nm attenuated backscatter along the yellow (south to north) line indicated by the green arrow as shown in (b) at 06:11–06:15 UTC. (b) MODIS false color image at 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.

required. In **Figure 2**, blowing snow is a dirty gray-white, the ice/snow surface (in clear areas) is blue, and clouds are generally a brighter white. Also shown in **Figure 2** 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 for that day are shown. The green dots denote the coastline. Plainly seen along the coast near longitude 145–150 E is blowing snow being carried off the continent. This is an important but poorly understood component of the ice sheet mass balance equation and will be discussed further in Section 5.

The temperature and moisture structure of blowing snow layers are very important to the calculation of the sublimation rate of blowing snow particles (sublimation is discussed further in Section 5). This information is typically acquired by radiosondes, but during intense blowing snow episodes, they cannot be launched due to high winds. Thus, the temperature and moisture structure of these layers remain somewhat unknown. Surface measurements during blowing snow have shown that sublimation of the snow particles will cool and moisten the air. Eventually, the air becomes saturated and sublimation ends. However, this has been shown to be true only near the surface. Measurements higher up in the blowing snow layer have not been made. Likewise, models of blowing snow also indicate that the blowing snow layer quickly saturates and sublimation reduces to near zero in a matter of hours after initiation [3, 18]. However, the models may be missing physical processes that keep the layer from reaching saturation.

Recent work has utilized dropsonde measurements to better understand the thermodynamic structure of blowing snow layers [19]. The Concordiasi Project, which occurred in the austral fall of 2010 (September 28–December 8), utilized multiple high altitude, long duration balloons to launch 648 dropsondes over Antarctica and surrounding sea ice [20–22]. The dropsondes measured temperature, moisture, and wind from the lower stratosphere to the surface. Some of the sondes fell through blowing snow layers as measured by CALIPSO and MODIS. One example is shown in **Figure 3**, which shows the CALIOP measured 532-nm attenuated backscatter in the color image with the dropsonde measured temperature (red profile) and moisture (white profile) drawn on the image. The distance between the dropsonde and the CALIPSO track was less than 5 km, but the dropsonde data were acquired about 7 hours after the CALIPSO data. However, blowing snow was very

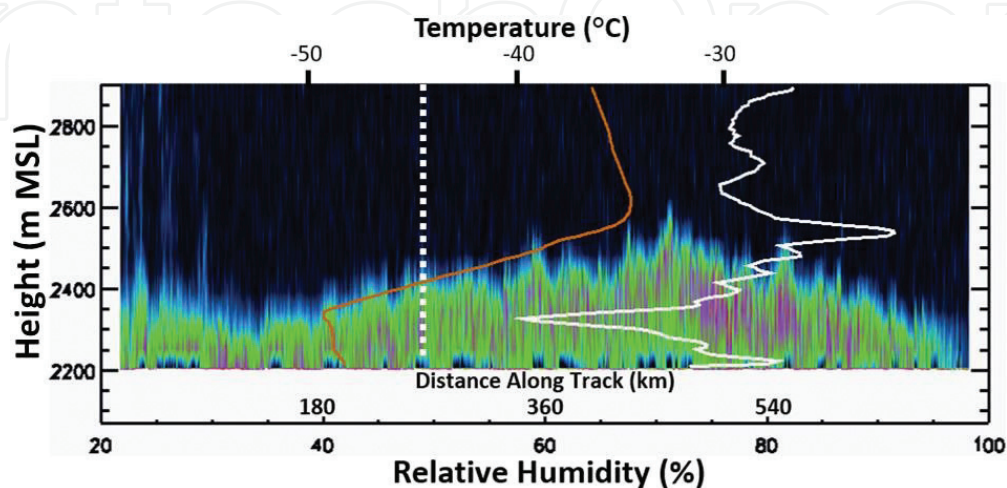


Figure 3. CALIPSO calibrated attenuated backscatter at 05:52:50–05:54:40 UTC on October 12, 2010 showing the blowing snow layer, the approximate position of the dropsonde (vertical dashed white line), temperature (red line, scale at top), and relative humidity with respect to ice (white line, scale at bottom) as measured by the dropsonde. Note, height scale is in m above the mean sea level (MSL), and the surface is at 2200 m MSL.

likely still occurring at the time of the dropsonde (12 UTC) since CALIOP and MODIS data showed blowing snow still occurring in the area 2 hours after the dropsonde. The temperature profile in **Figures 3** and **4a** shows the normal low-level inversion beginning at about 350 m above the surface or 2550 m above mean sea level (MSL). However, it does not continue to the surface, but rather, at the height of the top of the blowing snow layer ($\sim 140\text{--}150$ m above the surface), the temperature profile increases slightly as it descends. The average lapse rate in the lowest 150 m is almost adiabatic (-0.0088 versus $-0.0098^\circ\text{C m}^{-1}$), which is between moist and dry adiabatic. There are even regions of the temperature profile that have a lapse rate less than dry adiabatic (**Figure 4a** between 20 and 50 m above the surface).

The relative humidity (with respect to ice) profile in **Figures 3** and **4a** shows an ample structure both above and within the blowing snow layer. Well above the blowing snow layer, the relative humidity averages about 75%. As the dropsonde descends into the blowing snow layer, and at almost the exact height of the inflection point of the temperature profile (~ 150 m above the surface), the relative humidity begins to increase from a value of about 60% near the top of the layer to about 82% within roughly 10 to 20 m of the surface. From there, it decreases to a value of about 75% at the surface. Most importantly, note that the relative humidity is not saturated within the layer.

The wind speed (blue line in **Figure 4b**) reaches a maximum of almost 24 m s^{-1} (~ 53 miles per hour) at an altitude of about 2350 m MSL, which is 150 m above the surface and very near the top of the blowing snow layer. From that altitude, the wind speed decreases linearly to roughly 15 m s^{-1} (33.5 miles per hour) close to the surface. The wind direction (red line in **Figure 4b**) varies from about 155° at 2400 m altitude to 184° at 50 m above the surface. The magnitude of the wind speed and directional shear in the lower 150 m (corresponding to the blowing snow layer) will undoubtedly produce turbulence in the layer and promote mixing. It is apparent that the mixing has destroyed the temperature inversion at the surface (assuming it existed prior to the onset of high wind speeds and blowing snow) by the process of entrainment of warmer air from above and/or adiabatic warming of the descending katabatic flow.

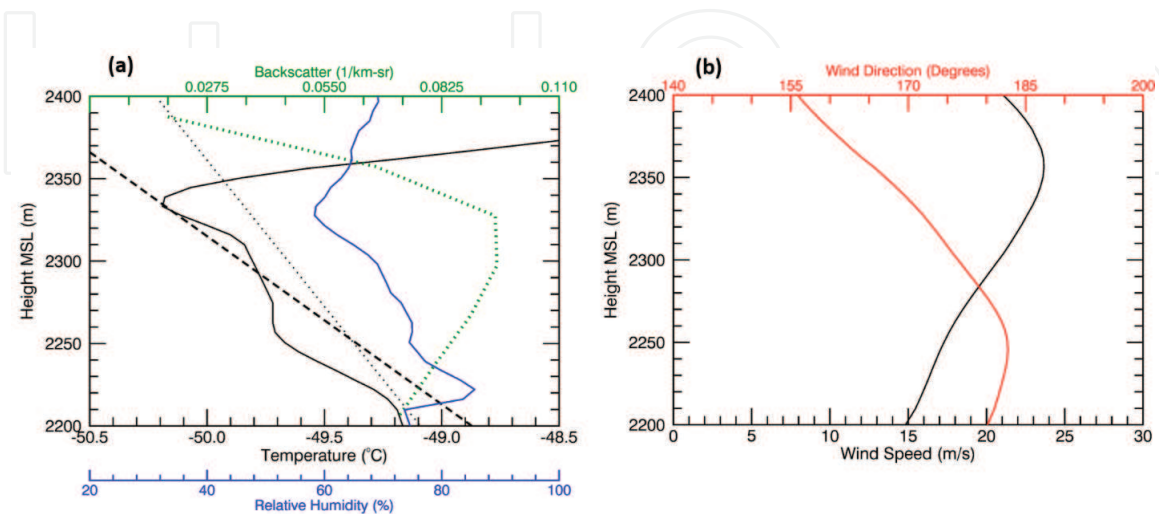


Figure 4. (a) A magnified view of the dropsonde temperature (black solid line) and humidity (blue solid line) profiles on October 12, 2010 at 12:49 UTC and CALIPSO average backscatter (green dotted) profile at 05:50 UTC for just the lowest 200 m above the surface for the data in **Figure 3**. Also shown is the dry (black dashed line) and moist (dotted black line) adiabatic lapse rates. (b) Dropsonde wind speed (black) and direction (red) through the blowing snow layer shown in **Figure 3**. Location of dropsonde: 71.61 S , 143.44 E .

4. Climatology of blowing snow

Understanding the spatial coverage and temporal changes of blowing snow is crucial if we are to fully understand how it impacts Antarctic climate, mass balance, and hydrology. Because of the harsh climate of Antarctica and the scarcity of observations, there are few direct observations of blowing snow that cover long time periods. Most observations are near the coasts and are limited in time, covering months or a few seasons [9, 10, 23]. The CALIPSO mission has enabled the construction of a 12-year climatology of blowing snow over Antarctica. This constitutes the basis of a longer term record that can be used to examine variability and trends. The algorithm described in Section 2 has been applied to the CALIOP backscatter data acquired between 2006 and 2017. The result, shown in **Figure 5**, is the average wintertime (April–October) blowing snow frequency for that period. Note that, CALIOP began operating in June of 2006, and thus April and May are missing from that year's average. **Figure 5** indicates that large areas of Antarctica experience blowing snow more than 50% of the time. Notable patches of even higher blowing snow frequencies are found over the Megadune region east of the Transantarctic mountains, south of 75 S from 120 to 160 E, and near the Lambert Glacier along 60 to 80 E longitude. Note also that these frequency values do not contain shallow (<30 m) blowing snow layers or drifting snow, since that is the vertical resolution of the CALIOP data. In addition, blowing snow that occurs during synoptic storms is also not included as most of the time these storms contain clouds thick enough to obscure the ground (attenuate the lidar beam so that the ground cannot be detected). The latter point has a large effect only near the East Antarctic coasts and most of West Antarctica where precipitation events via synoptic storms are more frequent.

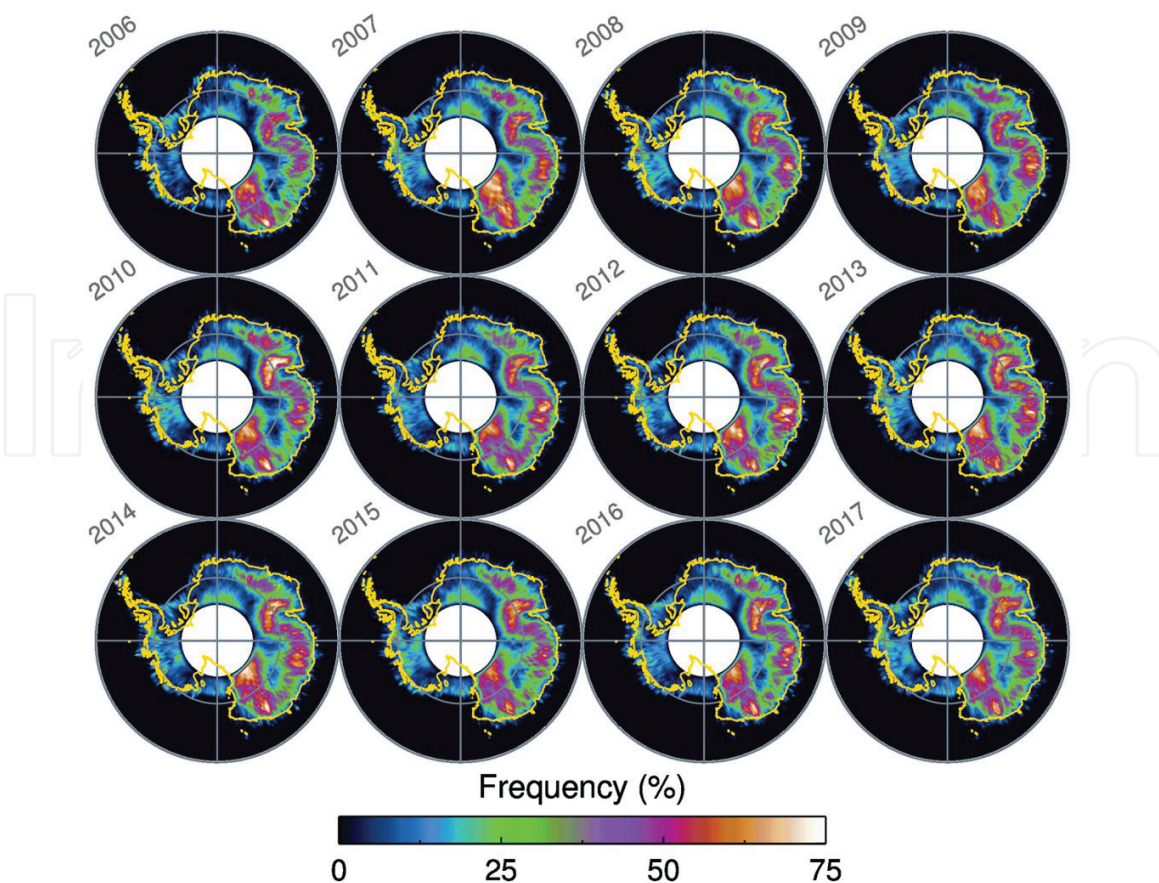


Figure 5.
The average blowing snow frequency over Antarctica for the winter months (April–October), of each year 2006–2017.

Shown in **Figure 6** is the intra-annual frequency of blowing snow for the period of 2006–2017. This is made by creating the average blowing snow frequency for each month during that period and then averaging each of the 12 months. Note that 2006 does not have data prior to June, and thus, January to May represent the average of 11 months. Blowing snow is prevalent in all months except November through February, though in these summer months, it still occurs at a reduced frequency. It is striking how fast the blowing snow frequency increases from February to March and how fast it decreases from October to November. These time periods generally coincide with the setting and rising of the sun, respectively. We hypothesize that the abrupt increase/decrease in blowing snow frequency is due to an increase/decrease in the katabatic wind flow, which is related to the increase/decrease in radiative cooling when the sun sets/rises. From April through September, large regions of Antarctica experience blowing snow more than 70% of the time with the overall spatial pattern and magnitude of blowing snow frequency remaining fairly constant through the 12 year period.

As the climate warms, it is expected that Antarctic precipitation will increase [24–27]. Climate models predict on average about a 7.4% increase in precipitation per degree of atmospheric warming [28]. Given the scarcity of observations and the problems of measuring precipitation in Antarctica, it is very difficult to verify any changes in Antarctic precipitation if indeed it is occurring. However, since blowing snow depends, at least partially, on the availability of snow [1, 29], it is reasonable to suggest that the frequency of blowing snow over Antarctica could increase as precipitation increases in a warming climate.

Using the winter average (April–October) blowing snow frequency for each year, we computed the trend of the blowing snow frequency for each $1 \times 1^\circ$ grid box over Antarctica for the 12 years of data. The student t-test for significance was

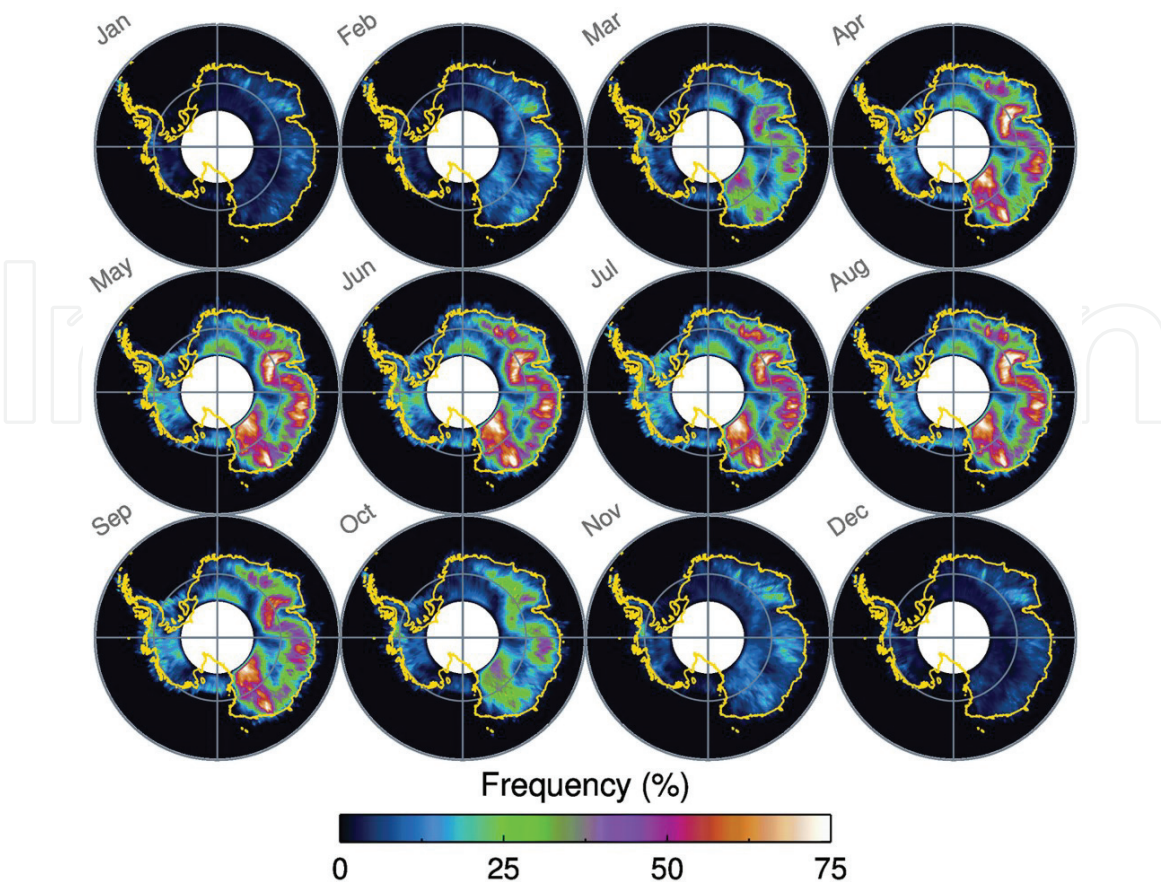


Figure 6.
The average monthly blowing snow frequency over Antarctica for the period of 2006–2017.

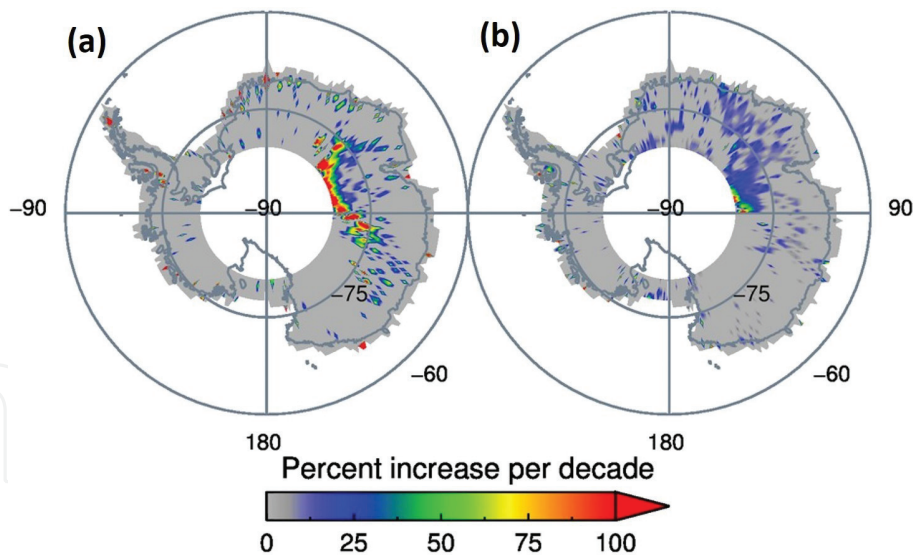


Figure 7. Percent increase in blowing snow frequency (a) and 10 m wind speed (b) per decade for areas with trends significant at the 95% or greater level. The grid box size for the plot is $1 \times 1^\circ$.

applied to each of the time series. The grid boxes that had positive trends significant at the 95% level or greater are shown in **Figure 7a**. Some very small areas of significant negative trends were also found (not shown). The color scale refers to the percentage increase in blowing snow frequency over a 10-year period. Here, we see an increase in blowing snow frequency up to 100% per decade over East Antarctica bounded between 45 and 95 E longitude. The observed geographic preference for increasing blowing snow frequency lies along the sloping edge of the ice divide. This feature in the time series trend map is intriguing. The red areas on the map in **Figure 7a** indicate grid boxes that experienced near a 100% increase in blowing snow frequency.

Since blowing snow is dependent on both the availability of snow and the near surface winds (along with the surface roughness and the snow properties), we computed the trends in 10-m wind speeds to see if the wind speed was increasing in the areas of increasing blowing snow frequency. The 10-m wind speed used was taken directly from the CALIPSO data product (V4-00), which uses the NASA Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2) reanalysis. In **Figure 7b**, we see spatially similar increasing trends in 10-m wind speed in some of the areas that show increasing blowing snow frequency. However, there are a few other small areas where an increasing trend in wind speed is observed with no significant trends in blowing snow frequency.

5. Blowing snow sublimation and transport

Blowing snow sublimation and transport are two important terms in the ice sheet mass balance equation (Eq. (1)). 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.

$$S = \int (P - E - M - Q_t - Q_s) dt \quad (1)$$

Of the terms in this equation, precipitation is by far the greatest in magnitude followed by Q_s and Q_t . Until recently, 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 reanalysis 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 [30, 31] develop and test a parameterization of blowing snow sublimation. Dery and Yau [2] utilize the model with the ECMWF reanalysis covering 1979–1993 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) year⁻¹. Lenaerts et al. utilized a high resolution regional climate model (RACMO2) to simulate the surface mass balance of the Antarctic ice sheet [32]. They found drifting and blowing snow sublimation to be the most significant ablation term reaching values as high as 200 mm year⁻¹ swe along the coast. There has been some recent work done on blowing snow sublimation and transport from field measurements (see for instance

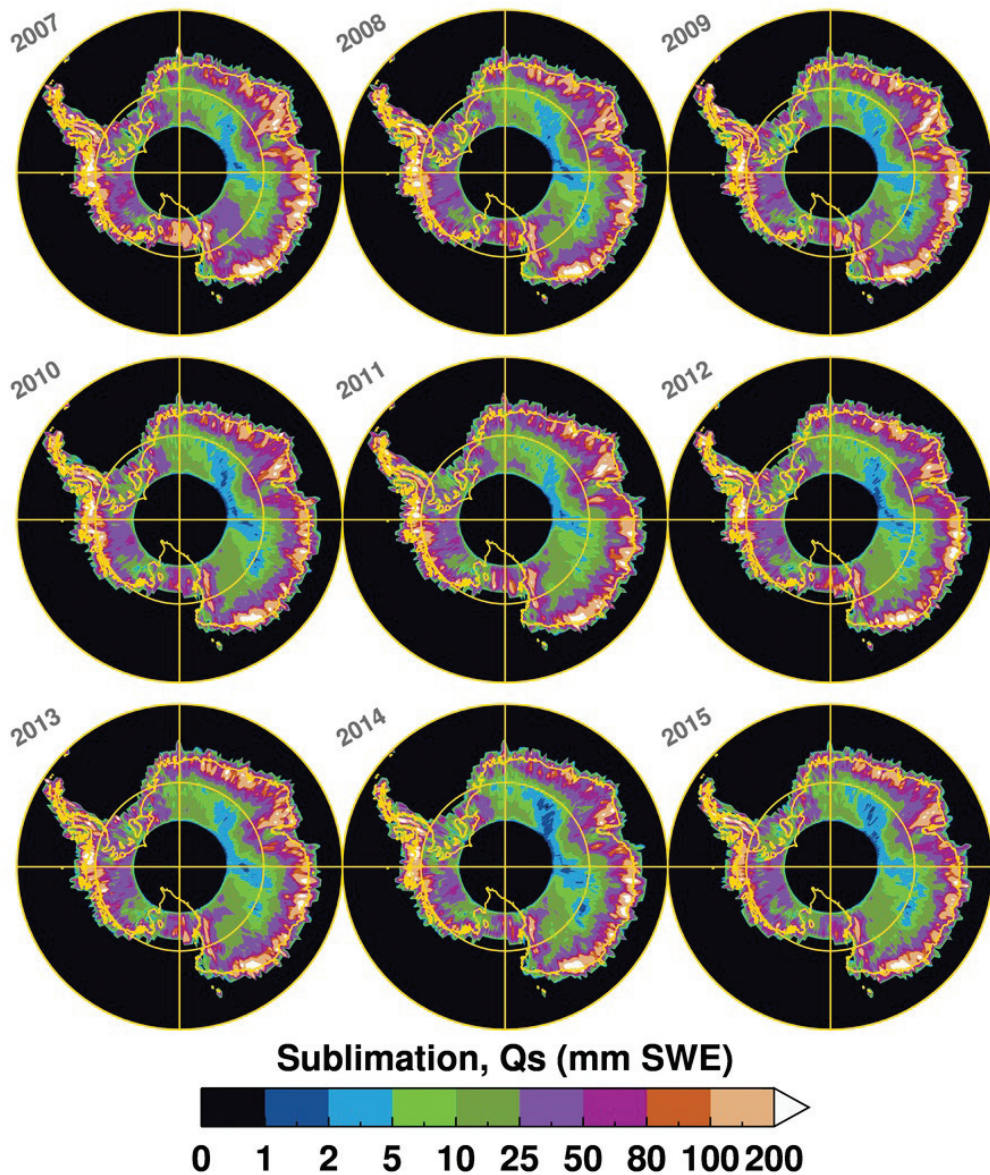


Figure 8.
Blowing snow total yearly sublimation over Antarctica for the period of 2007–2015.

[33, 34]), but the data are sparse, and the measurements are only available within the surface layer (<10 m).

While transport of blowing snow is considered 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. [35] 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 [12]. 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.

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 model parameterizations of blowing snow sublimation and transport. However, these quantities can also be computed using direct retrieval of blowing snow layers from CALIOP attenuated backscatter and model reanalysis fields of temperature, moisture, and winds [36]. Even with this method, the

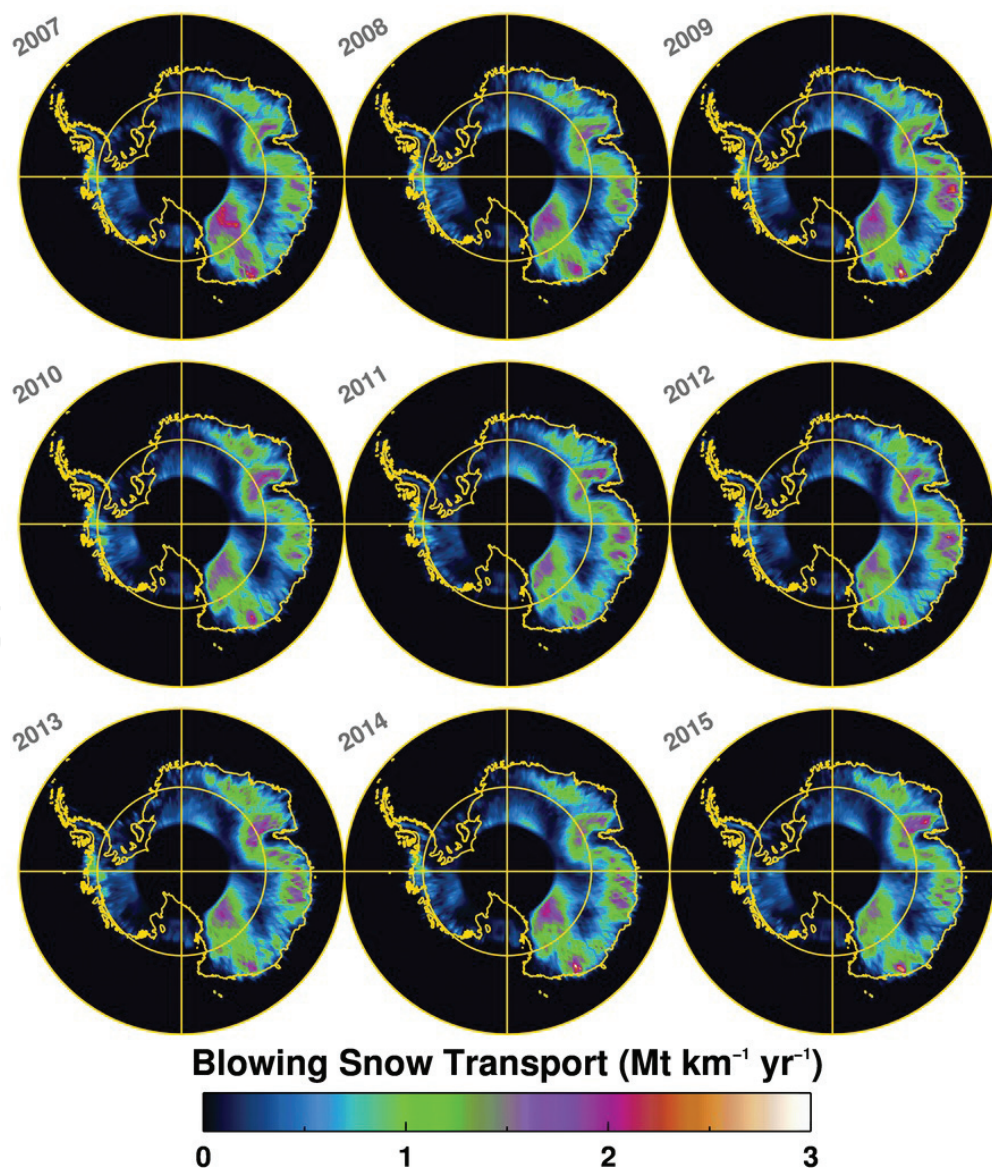


Figure 9.
The magnitude of blowing snow transport over Antarctica integrated over the year for years 2007–2015.

accuracy of the resulting sublimation is highly dependent on the accuracy of the model temperature and moisture fields. Using the CALIOP blowing snow backscatter and the MERRA-2 reanalysis, blowing snow sublimation was computed for the period of 2007–2015 (**Figure 8**). 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 (see **Figure 5**). Sublimation is highly dependent on the air temperature and relative humidity. For a given value of the blowing snow particle density, 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 [37, 38]. Continental interior areas with very high blowing snow frequency that approaches 75% (like the Megadune region in East Antarctica) exhibit fairly low values of sublimation because it is very cold and the model relative humidity is high. **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 82 S) for the CALIPSO period in Gt year^{-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, the year average swe (column 1) is multiplied by the surface area of Antarctica north of 82 S and the density of ice. The average integrated value for the 9-year period 2007–2015 of 393 Gt year^{-1} is significantly greater than (about twice) the values in the literature obtained from model parameterization [39].

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 (**Figure 9**). 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 **Figure 2**, at the bottom right of the MODIS

Year	Average sublimation (mm swe)	Integrated sublimation (Gr year^{-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
Average	43.5	393.4

*2006 and 2016 consist of only 7 and 9 months of observations, respectively.

Table 1.

The year average sublimation and the integrated sublimation over the Antarctic continent (north of 82 S) for 2006–2016.

image where snow is being blown off the coast. It turns out that this is quite a common phenomenon. Palm et al. [36] computed the mass of snow being carried off the continent by the process of blowing snow. They determined that in total about 3.68 Gt of snow is blown off the continent each year.

6. Summary and conclusion

Active remote sensing in the form of satellite lidar has given us a new perspective on, and an increased understanding of, blowing snow over Antarctica. We now know that large blowing snow storms are frequent, reach heights of 500 m, and often cover an area roughly the size of the state of Texas. From April to October, blowing snow occurs over 50% of the time over large areas of East Antarctica with some areas experiencing blowing snow 75% of the time in winter. The greatest blowing snow frequency is seen in the Megadune region of East Antarctica (south of 75 S and 120 to 160 E) and near the Lambert Glacier (60 to 80 E). Most areas of high blowing snow frequency coincide with areas of high average wind speed and/or high surface roughness. Blowing snow is prevalent in 8 months of the year, with only November through February devoid of areas of blowing snow frequency greater than 50%. Blowing snow frequency increases markedly from February to March and decreases significantly from October to November. This behavior is likely the result of katabatic wind speed increasing/decreasing as the sun sets/rises in the fall/spring.

Dropsonde and CALIOP backscatter data were utilized to investigate the temperature, moisture, and wind structure through the depth of blowing snow for the first time. The temperature structure through the layer is near isothermal, with the average lapse rate close to moist adiabatic. Above the blowing snow layer, the temperature profile is strongly stable (an inversion). The relative humidity was the greatest near the surface or slightly above (80%) and decreased through the depth of the layer with a minimum of about 60% near the layer top. Saturation was not reached within the layer indicating that sublimation of blowing snow particles was ongoing. Wind speed was 15 m s^{-1} near the surface and rapidly increased to 24 m s^{-1} near the layer top. The wind direction was constant in the lowest 50 m but backed by 25° in the upper 100 m of the layer. The near-isothermal temperature structure within the layer is likely due to the turbulent mixing of warm air from the inversion above the layer and caused by wind speed and directional shear. It is also possible that the relative humidity structure is influenced by the same process (entrainment of warmer and dryer air from above the layer), which keeps the layer from reaching saturation despite the sublimation of blowing snow particles. These results have potentially important implications for the amount of water vapor that is sublimated into the atmosphere during blowing snow episodes and also for ice sheet mass balance.

Blowing snow events identified by CALIPSO and meteorological fields from MERRA-2 were used to compute the blowing snow sublimation and transport rates. The 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 reanalysis 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, the results show that the 2006–2016 Antarctica average integrated blowing snow sublimation is about $393 \pm 196 \text{ Gt year}^{-1}$, which is considerably larger than previous model-derived estimates [2, 39]. The maximum blowing snow transport amount of

5 Megatons $\text{km}^{-1} \text{year}^{-1}$ occurs over parts of East Antarctica and aligns well with the blowing snow frequency pattern. The amount of snow transported from continent to ocean was estimated to be about 3.7 Gt year^{-1} . These continent wide estimates of blowing snow sublimation and transport based on the direct measurements of blowing snow layers are the first of their kind and can be used to help model and constrain the surface-mass budget over Antarctica.

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