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Key Points:

- Snow/ice enhanced Weather Research and Forecasting (WRF-ice) model reasonably captures surface conditions of ice sheet, ice shelf, and sea ice when verified against the in situ and satellite observations
- Blowing snow effects within WRF-ice include moistening and cooling of the air, cloud enhancement, and intensified precipitation
- Blowing snow can contribute to positive surface mass balance by increasing precipitation and reducing surface runoff and sublimation

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Case Study of Blowing Snow Impacts on the Antarctic Peninsula Lower Atmosphere and Surface Simulated With a Snow/Ice Enhanced WRF Model

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Abstract To better capture the air-snow-ice interaction, a snow/ice enhanced Weather Research and Forecasting (WRF-ice) model has been developed. This study examines the performance of WRF-ice and its blowing snow component during a strong cyclone event from October 23 to 27, 2017 over the Antarctic Peninsula, which is characterized by a synoptic cyclone crossing the northern part of the Peninsula and an embodied mesoscale cyclone over the Weddell Sea. Evolution of the cyclone is accurately reproduced in the 5-km resolution WRF-ice simulation, and the simulated near-surface conditions agree well with station and satellite observations. Numerical simulations show that the process of blowing snow sublimation can be prominent within the lower atmosphere when the air is dry, and produces moistening and cooling effects. Over relatively warm and humid areas, cloud enhancement by blowing snow can lead to either colder or warmer surfaces because of competing effects of longwave and shortwave cloud radiative forcings. In particular, additional moisture from blowing snow sublimation can slightly intensify precipitation over the mountains. Surface energy budget analysis indicates that absorbed shortwave (S_a), incoming longwave (L_d), and outgoing longwave (L_u) are dominant components of surface energy budget. When increases in L_d , L_u , and sensible heat flux are combined with decreases in S_a and latent heat flux due to blowing snow effects, a negative surface net heat flux ($\sim 0.5 \text{ W/m}^2$) occurs during daytime. A positive domain-total surface mass balance ($\sim 0.43 \text{ Gt}$) is generated during the simulated cyclone event due to increases in precipitation, decreases in runoff, and decreases in sublimation.

1. Introduction

The Antarctic Peninsula, separated from South America by the Drake Passage, is the most inhabited part of Antarctica (Meredith et al., 2017). During the late 20th century, the Peninsula was the fastest-warming region on Earth (Vaughan et al., 2003). Accompanied by the warming, there have been profound and rapid changes, including the disintegration of Larsen A and most of Larsen B ice shelves in 1995 and 2002, respectively (MacAyeal et al., 2003; Rott et al., 1996) and a partial collapse of Larsen C in 2017 (Hogg & Gudmundsson, 2017). However, since the late 1990s, there has been summer cooling in the Antarctic Peninsula, resulting from a higher frequency of cold east-southeast winds due to enhanced cyclonic circulation in the northern Weddell Sea (Oliva et al., 2017; Turner et al., 2016). Furthermore, several numerical modeling studies have been conducted to better understand the key mechanisms and processes that influence the Peninsula region (e.g., Bozkurt et al., 2020; Powers et al., 2012; Vaughan, 2006; C. Zhang & Zhang, 2018).

In addition, a near real-time weather forecast–Antarctic Mesoscale Prediction System (AMPS) over the entire Antarctic region has been carried out with the polar-optimized Weather Research and Forecasting (polar WRF) model for more than a decade (Powers et al., 2012). WRF (Skamarock et al., 2005, 2008, 2019) is the state-of-the-art high-resolution atmospheric model that has been applied, tested, and improved for the polar region (e.g., Bromwich et al., 2013; Cassano et al., 2011; K. M. Hines et al., 2015; Powers et al., 2012; Wille et al., 2016, 2017; X. Zhang et al., 2013). Modifications made to the WRF by K. Hines and Bromwich (2008), Bromwich et al. (2009), K. Hines et al. (2011), and Wilson et al. (2011, 2012) led to the development of polar WRF. The polar optimizations are primarily within the land surface module of WRF and improve the representation of heat transfer through snow and ice. One of the recent components added

to polar WRF includes allowing the specification of spatially varying sea ice thickness and snow depth over sea ice (K. M. Hines et al., 2015). In addition to the improvements made within polar WRF, X. Zhang et al. (2013) also coupled a thermodynamic sea ice model (X. Zhang & Zhang, 2001) to WRF in their efforts to generate the Chukchi-Beaufort High-Resolution Atmospheric Reanalysis (Liu et al., 2014; X. Zhang et al., 2013). Evaluations of WRF and polar WRF have been conducted by a number of studies in various contexts for the polar region (e.g., Bromwich et al., 2013; K. M. Hines et al., 2019; Tastula et al., 2012; Wille et al., 2017, 2016; Wilson et al., 2011, 2012; X. Zhang et al., 2013). For example, evaluation of polar WRF over all of Antarctica by Bromwich et al. (2013) concluded that polar WRF has cold summer and warm winter biases in 2 m air temperature. A positive surface wind bias has also been observed. Wille et al. (2017) evaluated the boundary simulation within AMPS at 3-km resolution against field campaign observations at the Ross Ice Shelf and the European Centre for Medium-Range Weather Forecasts (ECWMF) Re-Analysis (ERA). Their results suggest that AMPS has a 10%, year-around dry bias within the boundary layer. Recently, summer cloud microphysics from AMPS and a newer version of polar WRF 3.9 were compared against the Atmospheric Radiation Measurement West Antarctic Radiation Experiment (AWARE) observations by K. M. Hines et al. (2019). It is found that in the AMPS forecast simulation, liquid clouds are rarely observed. In contrast, they are frequently observed in AWARE. Although more accurate cloud liquid water content and cloud radiation forcing are captured by the advanced microphysics scheme in polar WRF 3.9, all simulations underpredict cloud fraction. The dry bias of atmospheric humidity over Antarctica also occurs in other numerical weather prediction models and has been ascribed to a lack of blowing snow impacts (Amory et al., 2015; Barral et al., 2014).

In situ and remote sensing observations suggest that blowing snow is a common phenomenon over the Antarctic region (Amory, 2020; Gossart et al., 2017; Leonard & Maksym, 2011; Palm et al., 2018, 2011), with a frequency as high as 70% (Palm et al., 2018, 2011). Blowing snow affects not only the surface mass balance and hydrological processes via redistribution of snow accumulation but also the shortwave and longwave radiation budget at the surface (Mahesh et al., 2003). Studies of persistent wind influence on Antarctic surface mass balance indicate that only atmospheric models which include blowing snow parameterizations are likely to provide accurate estimation of surface mass balance (Das et al., 2013; Grazioli et al., 2017). And indeed Palm et al. (2017) estimated an average snow transport of 3.7 Gt per year from Antarctic continent to ocean based on the 11-year remote sensing blow snow observations.

To parameterize the blowing snow processes and their corresponding impacts on atmospheric conditions, Déry et al. (1998) developed a blowing snow model named PIEKTUK. This model was able to predict the spectral number density of blowing snow and the effects of blowing snow particle sublimation on air temperature and humidity. For the sake of increasing computational efficiency, the one- and two-moment blowing snow models were further implemented and can predict the mixing ratio and total number concentration of blowing snow (Déry & Yau, 1999, 2001). The one-moment blowing snow model (Déry & Yau, 1999) was coupled with the Regional Atmospheric Climate Model and applied to evaluate the climatological effect of blowing snow on the Antarctic surface mass balance (Lenaerts, Broeke, et al., 2012; Lenaerts & van den Broeke, 2012; Lenaerts et al., 2012b).

To better capture the snow-ice-air interactions over the polar region, a snow/ice enhanced WRF model (WRF-ice) has been developed based on previous efforts. These efforts include the coupling of thermodynamic sea ice by X. Zhang et al. (2013) and the coupling of a thermodynamic snow-firn by Yao et al. (2016). Several other improvements such as variable sea ice/snow thickness adopted from polar WRF are also included. Within WRF-ice, the double-moment blowing snow physics model developed by Déry and Yau (2001) has been coupled. This allows full interactions between blowing snow processes and the atmospheric boundary layer to be represented in the model.

In this study, we examine blowing snow effects by using WRF-ice to model a cyclone system passing through the Antarctic Peninsula area in October 2017. The goals of this study are twofold. First, the performance of WRF-ice in modeling near-surface meteorological conditions is evaluated by comparing it to both in situ automatic weather stations observations and Moderate Resolution Imaging Spectroradiometer (MODIS) satellite retrievals over ice sheets, ice shelves, and sea ice. The variables used for evaluation of the WRF-ice model include surface pressure, 10-m wind speed, surface temperature, 2-m temperature, and dew point temperature. The in situ station observations used in this study are from the Integrated Surface Dataset

(ISD; Smith et al., 2011) obtained from the National Centers for Environmental Information (NCEI) and have been quality-controlled. The MODIS data used in this study include the products of land surface temperature (LST; Wan et al., 2015a, 2015b), and ice surface temperature (IST; Hall & Riggs, 2015). Within the study domain, only two radiosonde observations are available from the Integrated Global Radiosonde Archive (IGRA; Durre et al., 2006) at NCEI. Comparisons against the IGRA observations suggest that the WRF-ice model basically captures the observed profiles of temperature, humidity and wind, which however are not presented in details due to limited data availability. Second, blowing snow impacts on surface energy and surface mass balance are investigated. This is accomplished by comparing control (blowing snow on) and sensitivity (blowing snow off) simulations.

The remainder of this paper is organized as follows. Section 2 provides an overview of the snow and ice physics within WRF-ice. Sections 3 and 4 briefly document the model set up, simulation design, as well as metrics for surface energy budget and surface mass balance. The simulation results and verifications against both in situ station and satellite observations are presented in Section 5. Section 6 examines blowing snow and its interactions with the lower atmosphere and surface. This is done with the aid of a sensitivity simulation experiment in which blowing snow processes are deactivated. In this way, a conceptual model summarizing the blowing snow effects within WRF-ice is developed.

2. Snow and Ice Physics within WRF-Ice

Within WRF-ice, three extra snow/ice modules have been coupled with the Advanced Research WRF (ARW) model. They include a thermodynamic snow-firn module (Yao et al., 2016), a thermodynamic sea-ice module (X. Zhang & Zhang, 2001), and a double-moment blowing snow module (Dry & Yau, 2001).

As a substitute for the original unified Noah land surface scheme within ARW, an upgraded snow-firn module which can capture comprehensive snow-firn-ice physics has been included in WRF-ice (Yao et al., 2016). The snow-firn module used in this study has 20 layers of snow-firn for a total depth of 10 m. The evolution of snow grain size, snow density, firn density, temperature, and liquid water within each snow-firn layer are calculated. Liquid water from melt and rainfall percolates through the snow-firn layers, where it can be stored, refrozen, and/or removed as runoff. More details on the snow-firn module can be found in Yao et al. (2016), in which verifications against MODIS surface temperatures indicate improvements in surface temperature simulations over Antarctica.

Sea ice in the polar region plays important roles in the environment. This includes modifying air-sea exchange and impacting atmospheric and oceanic conditions. Furthermore, the sea ice extent and thickness in the polar region can vary dramatically throughout the year and even within a single day (Zwally et al., 2002). Therefore, it is necessary to accurately characterize the spatial distribution of sea ice. To this end, improvements made by polar WRF to include varying sea ice and snow thickness (K. M. Hines et al., 2015) has been adopted in WRF-ice. In addition, to improve the modeling of sea ice surface temperature and energy fluxes, a thermodynamic sea ice module developed by X. Zhang and Zhang (2001) has been coupled to WRF-ice by X. Zhang et al. (2013). Sea ice temperature, snow surface temperature, snow density, and melt are calculated based on the energy budget at the sea ice or snow surface. Verifications against the Surface Heat Budget of the Arctic (SHEBA) measurements from X. Zhang et al. (2013) indicate that the coupled sea ice module improves the simulation of sea ice and snow surface variables.

WRF-ice is coupled with the double-moment blowing snow module developed by Dry and Yau (2001), in which the full interaction between blowing snow and planetary boundary layer is included. Within the blowing snow module, the zeroth moment (total number concentration N [m^{-3}]) and third moment (mixing ratio Q [kg/kg]) of the blowing snow particle size distribution are predicted by the processes of eddy diffusion, blowing snow sedimentation, and sublimation. Blowing snow sublimation and its associated thermodynamic tendencies are added to atmospheric temperature and humidity within the planetary boundary layer of WRF-ice. In this study, a total of 24 vertical layers from the surface to 1-km above ground are utilized within the blowing snow module. Following Dry et al. (1998), blowing snow activates when the surface is snow-covered, and the total wind speed at 10 m height is above a certain threshold. The wind speed threshold U_i in this study is calculated as a function of near-surface air temperature (T_a in unit of $^\circ\text{C}$) following Li and Pomeroy (1997), by the following equation:

Figure 1. Model domain with 5-km grid spacing for the Antarctic Peninsula region. Terrain elevation (m a.s.l.) is shaded in color. Locations of nine ISD/NCEI surface stations used for WRF-ice evaluation are labeled in magenta circles. ISD, Integrated Surface Dataset; NCEI, National Centers for Environmental Information; WRF, Weather Research and Forecasting

$$U_t = 6.975 + 0.0033 \times (T_a + 27.27)^2 \quad (1)$$

There are also more comprehensive considerations of snow properties in this threshold calculation (e.g., Lehning & Fierz, 2008; Lenaerts, Broeke, et al., 2012; Lenaerts, van den Broeke, et al., 2012), which will be tested and adopted within WRF-ice in the future.

Snow albedo, a critical parameter determining the amount of shortwave radiation absorbed by the surface, is highly variable in space and time (Klok et al., 2003). The physically based snow albedo parameterization scheme by Gardner and Sharp (2010), was implemented and tested within the thermodynamic sea-ice module by X. Zhang et al. (2013) and applied to all types of snow/ice surfaces including ice sheets, ice shelves, and sea ice. The parameterization includes a base snow albedo, which varies with snow grain size and also accounts for contributions from cloud optical thickness, solar zenith angle and impurities in the atmosphere. The version of this parameterization used in this study accounts for all these factors except impurity impacts. Evaluations against the SHEBA measurements indicated that the Gardner and Sharp albedo scheme captured the observed variation in albedo relatively well (X. Zhang et al., 2013). This helped improve the simulation of surface net shortwave flux and temperature.

3. Simulation Design and Model Setup

The simulation case used in this study is a cyclone system that impacted the entire Antarctic Peninsula and embodied a mesoscale cyclone over the Weddell Sea during October 23–26, 2017. This particular event is unusual for its strong surface winds, from which a polynya was formed in front of Rydberg Peninsula and Stange Ice Shelf according to NASA's satellite images. The mesoscale cyclone developed from the synoptic cyclone impinging on the Antarctic Peninsula and generated snowfall over the northern part of the Peninsula and the Weddell Sea. In contrast, over the Bellingshausen Sea, clear skies were present during most of the study period. Thus, the blowing snow effects under different sky conditions can be examined in this case study.

Two simulation experiments are conducted with the WRF-ice model. In the control simulation, named as CNTL, the blowing snow processes and the corresponding interactions with the atmosphere are included. In the sensitivity simulation, named Blow-off, the blowing snow physics and its effects are deactivated. The model domain centered at (68.25°S, 65°W) covers the Antarctic Peninsula region with 300 by 300 grid points at 5-km horizontal grid spacing (Figure 1). Fifty-eight terrain following vertical levels from the surface to 10 hPa are used in which 14 layers are within the lowest 1 km and the bottom layer centered around 7-m above ground level (AGL). The base state temperature is set to 273.15 K and 10 grid points are set for the lateral boundary relaxation zone. The snow/ice modules described in Section 2 are used to capture the thermodynamic processes within snow-firn and sea ice. The 2.5-order Mellor-Yamada-Nakanishi-Niino boundary layer parameterization scheme (Nakanishi & Niino, 2006) is used for the parameterization of vertical diffusion of heat and moisture flux processes. The radiation processes are parameterized using the Rapid longwave and shortwave Radiative Transfer Model (Iacono et al., 2008). The Morrison two-moment scheme (Morrison & Gettelman, 2008) is utilized for the parameterization of microphysics processes. All the simulations are initiated at 0000 UTC October 23, 2017 and integrated for 72 h with a time step of 10 s. The first 6 h of the simulation are used for model spin up and discarded from analysis. The remained 66-h modeling results are presented in this study. The initial and boundary conditions at 6-h intervals are provided by the fifth ECMWF reanalysis (ERA5; Hersbach et al., 2020) at 0.25 × 0.25 resolution.

Since more than 3,200 km² of ice disintegrated from the Larsen A and B ice shelves in 1995 and 2002 (MacAyeal et al., 2003; Rott et al., 1996), Bedmap2, the new suite of gridded products (Fretwell et al., 2013) describing surface elevation, ice thickness and the seafloor of Antarctica is used to update the static geographical

surface conditions within WRF-ice. After modification, Larsen A and most of Larsen B ice shelves are replaced by water, and the corresponding surface conditions (e.g., surface temperature and albedo) have also been updated. Moreover, as there is no snow over sea ice within the ERA5 reanalysis data set, an extra daily snow depth over sea ice has been added to the WRF-ice initial conditions via the nearest-neighbor interpolation method. The daily snow depth data are from the Advanced Microwave Scanning Radiometer-Earth Observing System (AMSR-E) and the Advanced Microwave Scanning Radiometer 2 (AMSR2) Unified L3 daily data set at 12.5 km spatial resolution (Meier et al., 2018). In addition, the high-resolution coastlines used in all figures are generated with the Quantarctica data developed by the Norwegian Polar Institute (Matsuoka et al., 2018).

4. Metrics for Surface Energy Budget and Mass Balance

Accurate estimation of snow/ice mass changes under climate forcing is one of the major motivations for developing the WRF-ice model to use in the study area. Thus, the impacts of blowing snow on the surface energy fluxes and mass balance are examined by comparing the simulation results of the CNTL and Blow-off experiments. The surface net heat flux Q (W/m^2) is calculated by:

$$Q = L_d - L_u + S_a + Q_{gr} - Q_h - Q_l \quad (2)$$

Where L_d and L_u are the incoming and outgoing longwave radiation flux, and S_a is the solar radiation absorbed by the surface. L_d , L_u , and S_a are all positive. Q_{gr} is the ground heat flux, Q_h and Q_l are the surface sensible and latent heat fluxes, with positive (negative) as the upward (downward) fluxes.

The surface mass balance (SMB) for the entire snow-firn or snow-seaice layer, in mm water equivalent (w.e.), is calculated by:

$$SMB = P - R - E + D - S_s \quad (3)$$

where P is the accumulated precipitation, R is the total runoff from liquid precipitation and melt of entire snow-firn or snow-seaice layer, E and D are the blowing snow erosion and deposition, and S_s is the surface snow sublimation. Note that the blowing snow sublimation is removed from the blowing snow deposition D .

5. Simulation Results and Verifications

5.1. Cyclone Evolution during October 23–26, 2017

The cyclone system simulated in this study includes a synoptic cyclone that covers the entire Antarctic Peninsula and embodies a mesoscale cyclone over the Weddell Sea. Mesoscale cyclones are short-lived but intense low-pressure systems with a radius of less than 500 km and are generally embedded in larger structures (Pezza et al., 2016). Mesoscale cyclones play important roles in the atmospheric circulation of the Southern Hemisphere, not only for the poleward atmospheric moisture transportation (Zou et al., 2004), but also for their vigorous destructivity (Rasmussen & Turner, 2003). The synoptic patterns associated with the simulated cyclone system during October 23–26, 2017 are depicted in Figure 2, with the sea level pressure (SLP), surface temperature, and wind fields from the ERA5 reanalysis. This event was characterized by a synoptic cyclone center crossing the Drake Passage and the northern part of the Peninsula. As the synoptic cyclone center impinged on the western side of the Peninsula, the system underwent deformation due to orographic blocking, indicated by an increase of central SLP from 960 to 970 hPa (Figures 2a–2e). While the synoptic cyclone center crossed the Peninsula, strong warm advection occurred. In addition, a substantial reconstitution of the system on the leeward side occurred. In particular, a mesoscale cyclone embedded in the synoptic cyclone was present over the northeast coast, with a central SLP of 966 hPa at 0000 UTC October 24 (Figure 2c). During October 24–25, the mesoscale cyclone intensified and continued to linger while moving slightly southward along the east coast (Figures 2d–2h). After 0600 UTC October 25, the synoptic cyclone center moved out of the Peninsula and the mesoscale cyclone decayed (Figure 2i).

The CNTL simulation captures the detailed evolution of the cyclone system. In particular, it reproduces the deformation of the synoptic cyclone center when it impinged on the western side of the Peninsula. It

Figure 2. Synoptic features associated with a mesoscale cyclone along the east coast of the Antarctic Peninsula between 1200 UTC October 23 and 1200 UTC October 25 from ERA5 reanalysis. 2-m temperature (K) in shaded, 10-m wind vectors (m/s) in gray arrows, and sea level pressure (hPa) in black solid contours, at 2-hPa intervals

also reveals the generation and evolution of the mesoscale cyclone along the east coast of the Peninsula, with a central SLP of 966 hPa at 0000 UTC October 24 to 982 hPa at 1200 UTC October 25 (Figures 3a–3d). During 0000–1200 UTC October 24, the warm air advection toward the east coast of the Peninsula from the Weddell Sea and the cold air from the mountain ranges form a mesoscale coastal front next to the east coast, and the accumulated precipitation shows an orientation parallel to the front (Figures 3e and 3f). After 1200 UTC October 24, the mesoscale cyclone continues lingering southward along the east coast, with strong cold southerly barrier winds dominating the area. These winds arise from the blocking effect of the Peninsula's mountain range. After 0000 UTC October 25, the mesoscale cyclone dissipates and precipitation decreases (Figures 3c, 3d, 3g, and 3h). Meanwhile, the warm southeastward maritime advection across the Drake Passage is almost perpendicular to the mountain ranges over the northern tip of the Peninsula and

Figure 3. WRF-ice simulated sea level pressure (contours, at 2-hPa intervals), 2-m temperature (shaded, K), 10-m horizontal winds (arrows, m/s) (a–d), 12-h surface accumulated precipitation (shaded, mm) (e–h) between 1200 UTC October 23 and 1200 UTC October 25 at 12-h intervals. Sea ice edges are marked by gray solid contours in panels (e–h). WRF, Weather Research and Forecasting

Table 1
Summary of Nine Surface Stations and Datasets From ISD/NCEI

Name	Location	Interval	Elevation	Owner	Variables ^a
Base Bernardo	(63.31°S, 56.68°W)	6 h	10 m	Chile	$P_{slp}, T_{2m},$ Td_{2m}, V_{10m}
Bellignshausen	(62.18°S, 58.88°W)	6 h	14 m	Russia	
Great Wall	(62.22°S, 58.97°W)	6 h	10 m	China	
Base Carlini	(62.23°S, 58.63°W)	3 h	11 m	Argentina	
Palmer	(64.77°S, 64.08°W)	6 h	8 m	USA	$P_{slp}, T_{2m},$ Td_{2m}, V_{10m}
Esperanza	(63.4°S, 56.983°W)	3 h	13 m	Argentina	
Vernadsky	(65.25°S, 64.267°W)	3 h	11 m	Ukraine	
San Martin	(68.117°S, 67.133°W)	3 h	4 m	Argentina	
Rothera	(67.570°S, 68.124°W)	1 h	32 m	UK	$P_{slp}, T_{2m},$ Td_{2m}, V_{10m}

ISD, Integrated Surface Dataset; NCEI, National Centers for Environmental Information.

^a Variables P_{SLP} , T_{2m} , Td_{2m} , and V_{10m} are the abbreviations for sea level pressure, 2-m temperature and dew point, and 10-m wind speed.

causes another precipitation maximum over the northern part of its windward slope (Figures 3e–3h). The near-surface temperature over the east side of the Peninsula decreases substantially during 0000 and 1200 UTC October 25, especially during daytime. This is found to be related to the radiation cooling of cloud cover over this area.

Moreover, evolution of the cyclone system simulated by the CNTL experiment is compared with the AMPS forecast. As mentioned before, AMPS is the operational weather forecast for Antarctica conducted by the National Center for Atmospheric Research, in which a high-resolution of 2.67 km domain covers the Antarctic Peninsula. Comparison shows that the cyclone system captured by AMPS in its Antarctic Peninsula domain is quite similar to the CNTL simulation. Deformation of the synoptic cyclone center when impinging on the western side of the Peninsula and generation of mesoscale cyclone along the east coast of the Peninsula are all predicted by AMPS (not shown). However, the synoptic cyclone from AMPS is stronger by 4–6 hPa during 0000 UTC and 1200 UTC October 24, with a central SLP of 960 hPa, and the southeastward movement is slightly slower. Consequently, the precipitation produced by AMPS is stronger than the CNTL simulation, particularly during October 24, with greater area covered for the 12-h accumulated precipitation >30 mm. These differences may also be related to different grid resolution and microphysics schemes utilized in AMPS and CNTL. Such impacts from grid resolution and microphysics schemes are worthy of further investigation but beyond the scope of this study.

5.2. Evaluation of WRF-Ice Near-Surface Modeling Against Station Observations

To accurately simulate the blowing snow impacts, near-surface conditions need to be realistically captured. The performance of WRF-ice near surface conditions, including the SLP, temperature, dew point, and wind speed at the lowest model level (~7-m AGL), is evaluated against a total of nine in situ station observations acquired from the ISD/NCEI data set. To avoid diagnostic uncertainties, no adjustments are made to the simulated atmospheric conditions although there exists a slight height differences with the verification data. The locations of these stations are indicated as magenta dots in Figure 1. More information about the stations and the observed variables is summarized in Table 1. Station surface conditions (such as surface cover types and elevation) are critical information for near-surface modeling verification, especially for stations located at the complex coastal region of the Peninsula. Therefore, model results from the grid point closest to the station with the same surface cover type and smallest elevation difference are used to conduct the near-surface modeling verification. As difference between the CNTL and Blow-off simulations is small at the station locations, where the blowing snow sublimation is relatively weak, only the verifications of the CNTL simulation against the station observations are presented herein.

Table 2 summarizes the biases, root-mean-squared errors (RMSE) and correlation coefficients of the near-surface meteorological fields from the WRF-ice CNTL simulation verified against the nine ISD/NCEI

Table 2*Statistical Parameters for WRF-Ice Evaluation Against Station Observations Over the Antarctic Peninsula Region Between 0600 UTC October 23 to 0000 UTC October 26, 2017*

Name	Bias	RMSE	Correlation	Bias	RMSE	Correlation
P_{SLP} (hPa) (p -value < 0.0001)			$T_{2\text{m}}$ (°C) (p -value < 0.005)			
Base Bernardo	0.41	4.35	0.88	0.81	1.64	0.79
Bellignshausen	−0.96	3.05	0.94	−0.10	0.48	0.92
Great Wall	−0.84	2.59	0.96	−0.62	1.07	0.81
Base Carlini	−0.63	1.69	0.98	−0.68	0.97	0.95
Palmer	2.79	3.14	0.99	0.75	1.48	0.84
Esperanza	−2.38	2.75	0.98	2.89	4.38	0.48
Vernadsky	−1.00	2.12	0.99	−0.21	1.79	0.87
sSan Martin	−0.73	1.64	0.99	0.48	1.52	0.26
Rothera	−0.87	1.03	0.99	0.88	1.86	0.14
$V_{10\text{m}}$ (m/s) (p -value < 0.005)			$Td_{2\text{m}}$ (°C) (p -value < 0.005)			
Base Bernardo	−0.47	3.41	0.43	2.45	3.06	0.87
Bellignshausen	0.79	1.98	0.68	0.10	0.61	0.94
Great Wall	−1.73	4.32	0.89	−0.23	0.81	0.90
Base Carlini	1.43	3.25	0.56	−1.67	1.82	0.92
Palmer	1.36	2.35	0.77	−1.91	2.70	0.68
Esperanza	1.44	3.32	0.42	4.40	5.93	0.53
Vernadsky	6.81	7.80	0.25	0.16	0.97	0.91
San Martin	0.29	3.64	0.78	0.87	2.19	−0.12
Rothera	3.49	6.83	0.64	1.70	2.75	0.38

RMSE, root-mean-squared error; WRF, Weather Research and Forecasting.

surface stations. Figure 4 compares the time series of SLP, 2-m temperature and dew point, and 10-m wind speed between the CNTL simulation and station observations during 0600 UTC October 23 to 0000 UTC October 26, 2017. Biases of SLP are generally negative ranging from −0.63 to −2.38 hPa for most stations, except two stations including Base Bernardo and Palmer with positive values of 0.41 and 2.79 hPa. Strong correlation coefficients (0.88–0.99, with p -value < 0.0001) of SLP for all stations reflect a high skill of the CNTL simulation in capturing the cyclone's evolution and movement (Table 2). The pressure time series in Figures 4a1–4i1 shows that the CNTL simulation accurately captures the pressure variation during the cyclone event. In particular, SLPs in both the CNTL simulation and observations decrease to ~970 hPa between 0600 UTC October 23 and 0000 UTC October 24, and gradually increase to ~990 hPa around 0000 UTC October 26 as the cyclone center moves further southeastward.

The absolute biases of 2-m temperature are less than 1°C for all stations, except Esperanza. The absolute biases of 2-m dew point are relatively greater for some stations, but still less than 2°C except at the Base Bernardo and Esperanza stations. The corresponding correlation coefficients range from 0.79 to 0.95 (with p -value < 0.005) for 2-m temperature and 0.68 to 0.94 (with p -value < 0.005) for 2-m dew point, except at the Esperanza, San Martin, and Rothera stations (Table 2). This suggests that the overall observed patterns of near-surface air temperature and dew point can generally be reproduced by the CNTL simulation. The poor correlations (<0.4) at San Martin and Rothera, can be ascribed to highly complex features that cannot be completely resolved by the model. This is especially true for the Rothera station. The CNTL simulation has very strong diurnal variation in 2-m temperature at Rothera and does not capture the observed decrease in 2-m dew point at both the San Martin and Rothera stations during the last day of simulation (October 25, 2017; Figures 4h2–3 and 4i2–3). The observed 2-m temperature and dew point at the Esperanza station from 0600 UTC October 23 to 0600 UTC October 25 are very different from the CNTL simulation. By comparing with the observations at a very close station (Base Bernardo) where the CNTL simulation reasonably

Figure 4. Observed (dashed lines) and the CNTL simulated (solid lines) time series of sea level pressure (hPa, black lines in the first row, a1–i1), 2-m temperature (°C, magenta lines in the second row, a2–i2) and dew point (°C, green lines in the third row, a3–i3), and 10-m wind speed (m/s, blue lines in the fourth row, a4–i4) for nine stations marked by magenta dots in Figure 1 during 0600 UTC October 23 and 0000 UTC October 25.

captures the observed variations in both 2-m temperature and dew point (c.f., Figures 1, 4a2–3, and 4f2–3), it becomes clear that the measurements at Esperanza might have some issues. Besides, the time series comparisons between the CNTL simulation and station observations for 10-m wind speed (Figures 4a4–4i4) suggest that the observed near-surface wind speeds can generally be reproduced by the CNTL simulation, with absolute biases < 2 m/s at most stations (Figures 4a4–4f4 and 4h4). But at the stations of Vernadsky and Rothera, wind speed is overestimated (Figures 4g4 and 4i4). Again, these relatively large discrepancies might partially be ascribed to highly complex topographic features of the Peninsula coast which cannot be completely resolved by the model. The slight inconsistency in the height of the compared near-surface air temperature, dew point and winds may also contribute to the discrepancies.

5.3. Evaluation of WRF-Ice Surface Temperatures Against MODIS

The performance of WRF-ice surface temperature is evaluated using the MODIS products of LST over ice sheet surfaces and IST over sea ice surface. MODIS data are acquired from two satellites known as Terra and Aqua, by which the MODIS products are grouped as MOD and MYD, respectively. Preliminary comparisons of the WRF-ice simulations against the MODIS LST products indicate that cloud contamination exists, and this contamination is worse in the MOD LST. For this reason, additional quality control procedure is added, applying a threshold of 0.01 g/kg total cloud mixing ratio from the WRF-ice simulation to remove the cloud contamination. Then the in-house quality controlled MYD LST is used for the evaluation. Cloud contamination is not found in the IST fields.

Figure 5 shows the WRF-ice evaluations against the MYD surface temperatures at 0400 UTC and 2100 UTC October 25, 2017. Due to cloud contamination, the coverage of MOD surface temperatures during the study period is very limited and therefore not included in the evaluation. It should be noted that surface temperatures over the ice-sheet (LST) are available during both daytime and nighttime (Figures 5a and 5c), while surface temperatures over sea ice (IST) are only available during daytime (Figure 5b). The WRF-ice surface temperatures over the areas with available LST or IST at the corresponding times are shown in

Figure 5. Surface temperatures over the ice sheet (LST) and sea ice (IST; K) from MODIS observations (a–c), and the corresponding WRF-ice simulated surface skin temperatures (d–f) at 0400 UTC and 2100 UTC October 25, 2017. Differences between MODIS observations and the WRF-ice simulations are depicted (g–i). Note that IST is only available during daytime. Sea ice edges are marked by the gray solid contours in panels (a–c). IST, ice surface temperature LST, land surface temperature; MODIS, Moderate Resolution Imaging Spectroradiometer; WRF, Weather Research and Forecasting

Figures 5d–5f. Differences between MYD LST, IST, and the WRF-ice simulation are depicted in Figures 5g–5i. Comparisons indicate that the WRF-ice model can generally capture the surface temperature features over both the ice sheet and sea ice. During daytime, the RMSE and bias scores for LST (IST) are 3.55 and 0.30 K (3.11 and 0.23 K), respectively; and during nighttime, the WRF-ice model seems to underestimate the LST, with RMSE of 4.69 K and bias of -1.75 K (Figure 5). These discrepancies might be caused by limitations in WRF's ability to represent strong temperature inversions, especially during nighttime. Based on the evaluations of polar WRF over Antarctic sea ice, Tastula et al. (2012) noted that the strength of temperature inversions in the WRF model is overestimated in general and the surface temperature tends to be accordingly underestimated. Future research improving the surface boundary layer simulation within WRF would be beneficial for a more precise prediction of polar atmospheric conditions. Given that the aim of this study is

Figure 6. Differences between the CNTL and Blow-off simulations at 0500 UTC October 25 for (a) column-integrated water vapor mixing ratio (shaded, 10^{-4} kg/kg), (b) column-integrated cloud water mixing ratio (shaded, 10^{-4} kg/kg), (c) surface temperature (shaded, K) and horizontal winds (arrows, m/s), and (d) accumulated precipitation between 0600 UTC October 23 and 0500 UTC October 25 (shaded; mm). The 10-m wind vectors from the CNTL simulation is overlaid to panel (b). Blue dashed lines in panels (a) and (b) denote the cross-section locations in Figure 7. Sea ice edge at 0500 UTC October 25 is marked by a gray solid contour in panel (a), and box “A” and “B” in panel (d) are the mountain areas used in Figure 8.

to evaluate the blowing snow effects within WRF-ice, we focus on examining the discrepancies between the CNTL and Blow-off simulations in the next section.

6. Blowing Snow and Its Interactions with Lower Atmosphere and Surface

6.1. Blowing Snow Impacts on the Lower Atmosphere

Strong near-surface winds approaching 20 m/s (c.f., Figure 4c4) associated with the cyclone system simulated during October 23–26, 2017 make it an ideal case to investigate blowing snow and its associated thermodynamic effects within WRF-ice. To facilitate this investigation, simulation results from the CNTL and Blow-off experiments are compared. As described in Section 3, Blow-off has the same model configuration as CNTL except blowing snow and its effects are deactivated. Differences between the CNTL and Blow-off simulations are examined using variables which include water vapor, clouds, surface wind, surface temperature and total precipitation (Figure 6). The vertical distribution of blowing snow and its impacts on temperature and humidity of the lower atmosphere are analyzed under different thermal conditions (cold and dry vs. warm and humid [Figure 7]). Blowing snow impacts in mountainous environments (box “A” and “B” in Figure 6d) are also examined (Figure 8). Moreover, the surface energy budget and mass balance are compared between the CNTL and Blow-off simulations to gain physical insights regarding the potential impacts of blowing snow on changes to ice sheets and ice shelf (Figure 9). Finally, a conceptual model for blowing snow impacts on the lower level atmosphere under different conditions is proposed (Figure 10).

Figure 7. Vertical cross-sections of the blowing snow sublimation (shaded, $10^{-11} \text{ kg kg}^{-1} \text{ s}^{-1}$), blowing snow mass content (red contours, 10^{-5} kg/kg), and total number concentration (blue contours; 10^2 m^{-3}) (a and b) along the two blue dashed lines in Figures 6a and 6b at 0500 UTC October 25. Panels (c and d) and (e and f) are the corresponding vertical slices of the temperature (shaded, K) and total water mixing ratio (shaded, 10^{-4} kg/kg) differences between the CNTL and Blow-off simulations. Wind vectors in the CNTL simulation and total cloud water mixing ratio difference between the CNTL and Blow-off simulations are also overlaid (a and b, arrows) and (c and d, magenta contours, 10^{-4} kg/kg)

Comparisons between the CNTL and Blow-off simulations suggest that blowing snow tends to increase water vapor in the lower atmosphere via blowing snow sublimation. This helps to generate more clouds and precipitation when sufficient moistening and/or upward motion are present. As shown in Figures 6a and 6b, column-integrated water vapor mixing ratio and cloud water mixing ratio over the entire model height increase substantially, which indicates that water vapor over the Bellingshausen Sea and the west side of the Peninsula increases prominently. It also suggests that more clouds are generated over relatively warm and humid areas of the northern Bellingshausen Sea due to additional moisture condensation effect. This is consistent with the 3-year observational study by Barral et al. (2014), in which blowing snow was also found to raise the near-surface atmospheric moisture content close to saturation and impact the turbulent flux within the boundary layer. Surface temperature and horizontal wind speed increase over the areas where

Figure 8. Time series of the column-integrated hourly blowing snow sublimation (red line, mm) averaged over area “A” and hourly precipitation increase (blue line, CNTL-Blow-off) averaged over area “B” marked in Figure 6d(b). Boxes “A” and “B” denote the mountain areas where strong blowing snow sublimation occurs and precipitation is enhanced, respectively.

enhanced clouds are present (Figures 6b and 6c). The maximum increases in surface temperature and wind speed approach 8 K and 4 m/s, respectively. Over the central Peninsula and southern part of its east coast (e.g., over area “B”), the precipitation is increased by around 2% due to blowing snow. This topographic lifting and the blowing snow-induced moistening work together to increase the precipitation rate (Figure 6d).

To further explore the blowing snow effects under different conditions, we next examine the vertical distribution of blowing snow and its impacts on the lower atmosphere along two locations marked by the blue dashed lines in Figure 6. Figure 7 shows sublimation, total number concentration (N_{ib}) of blowing snow, mixing ratio (Q_b) of blowing snow (a/b), differences of temperature and cloud mixing ratio (c/d), precipitable total water mixing ratio (e/f) between the CNTL and Blow-off simulations under two conditions of cold and dry versus relatively warm and humid. Blowing snow effects are different between these two conditions. For example, at 0500 UTC October 25, 2017 over the cold and dry areas of the Bellingshausen Sea, strong blowing snow sublimation occurs at the near surface and within the layer of 200–1,000 m above ground (Figure 7a).

On the other hand, there is little sublimation under the warm and humid mid cloudy sky (Figure 7b). Simulated temperature and humidity profiles show that the driest air occurs around 450-m AGL (not shown), which results in peak blowing snow sublimation at around 450-m AGL (Figure 7a). This is consistent with previous studies which show that the blowing snow sublimation rate is greater under drier conditions (Dai & Huang, 2014; Groot Zwaftink et al., 2013). Furthermore, blowing snow sublimation causes around 0.5 K cooling, particularly in the upper air (Figure 7c). There is less cooling at the near-surface, where sublimation cooling is compensated by the enhanced turbulence (not shown).

Over the cloudy areas of the Bellingshausen Sea, there are positive cloud anomalies between 200 and 600 m AGL, which are locally generated and/or advected from upstream. Temperature profiles generally become warmer under the enhanced clouds, and cooler within/above the clouds (Figure 7d). Obviously, the enhanced cloud radiative forcing warms up the near surface temperature and the warming effect is generally

Figure 9. Time series differences between the CNTL and Blow-off simulations during 0600 UTC October 23 and 0000 UTC October 25 for the surface energy budget terms (a) of downward and upward longwave radiation (L_d blue, L_u orange), absorbed shortwave radiation (S_a red), ground heat flux (Q_{gf} brown), sensible (Q_h green) and latent heat fluxes (Q_l magenta), and the net flux (Q black dashed); and for the mass balance terms (b) of precipitation (P cyan), surface runoff (R brown), blowing snow erosion (E blue) and deposition (D orange), surface snow sublimation (S_s purple) and the net mass balance (SMB black dashed). Note that the surface energy budget and mass balance terms are averaged over the entire model domain except the ocean area.

Figure 10. Conceptual model of blowing snow impacts on the low-level atmosphere under (a) cold and dry conditions, (b) warm and humid conditions, and (c) in mountainous environment. Blue and red arrows in (b) represent the longwave and shortwave radiation, respectively, and the thickness of arrows indicates their relative strength.

stronger under lower and thicker clouds. This is consistent with the surface temperature responses to the enhancement of clouds in Figure 6c. The enhanced cloud layer can also reduce upward energy transfer, which accordingly results in cooler temperature within and above the clouds. During daytime, an enhanced cloud layer has both cooling and warming effects via modifying the shortwave and longwave radiative forcings. When the cloud is relatively low and thick, the warming effect from the longwave radiative forcing can exceed the cooling effect from reduced shortwave radiation. For higher and thinner cloud layers these warming and cooling effects can cancel each other. On the other hand, the cooling effect can also be dominant (not shown since the cloud enhanced area during daytime is relatively small in this case study). Similar conclusions have also been noted in previous studies which show that cloud properties, such as cloud water-bearing content, cloud thickness, and so on, can influence cloud longwave and shortwave radiative forcings (Matus & L'Ecuier, 2017; Miller et al., 2015; Zhao & Garrett, 2015). For both clear and cloudy sky conditions, blowing snow generally helps to increase the precipitable total water (Figures 7e and 7f).

Moreover, Figure 8 shows the time series of the column-integrated hourly blowing snow sublimation and hourly precipitation difference between the CNTL and Blow-off simulations averaged over the mountain areas “A” and “B” (as marked in Figure 6d). “A” and “B” are the mountainous areas where relatively strong sublimation and precipitation increases occur. The column-integrated hourly blowing snow sublimation at area “A” starts to increase at around 0600 UTC October 23, reaches a maximum of around 0.043 mm at 1800 UTC October 23, and later on decreases rapidly (Figure 8). The fast decrease of blowing snow sublimation is caused by the immediate cloud development in the area (not shown). Accordingly, when the enhanced precipitable total water contributed by the blowing snow sublimation over “A” is transported downstream to the area “B”, where an upward slope exists, the hourly precipitation over “B” is enhanced, with the peak increase around 0.023 mm at 1000 UTC October 24. Therefore, blowing snow sublimation helps to generate more clouds and precipitation when sufficient moistening and lifting are present.

6.2. Blowing Snow Impacts on the Surface Energy and Mass Balances

To gain insights on how blowing snow quantitatively impacts surface energy fluxes and mass balances during this event, Figure 9 shows the time series differences of area-averaged surface energy fluxes and mass balances between the CNTL and Blow-off simulations from 0600 UTC October 23 to 0000 UTC October 26, 2017. The averaged area includes the entire model domain except the ocean area. It should be noted that the major components of net surface heat flux are downward shortwave, downward longwave, and outgoing longwave radiation. The sensible and latent heat fluxes as well as ground heat flux are much smaller by comparison (not shown). The time series differences of each energy budget term between the CNTL and Blow-off simulations as shown in Figure 9a demonstrate that blowing snow tends to cause negative anomalies (up to -2 W/m^2) in the latent heat flux (Q_l) and positive anomalies (up to 2 W/m^2) in the sensible heat flux (Q_h) during the simulation period. Note that since the cyclone event during October 23–26, 2017 happened in a transition season from austral winter to summer, surface temperature and

water vapor mixing ratio in the study domain are generally warmer and greater than the air above, which results in overall upward sensible and latent heat flux (not shown). Moreover, ascribed to increased water vapor and cooler temperature within the lower atmosphere where blowing snow sublimation dominates, the vertical temperature gradient increases while moisture gradient decreases. As a result, the sensible heat flux increases and latent heat flux decreases (c.f., green and magenta lines of Figure 9a). These results are consistent with previous studies which highlight the changing characteristics of Antarctic surface energy balance caused by changing seasons (Déry & Yau, 2002; Van As et al., 2005; van den Broeke et al., 2005).

Along with the enhancement of clouds after 1200 UTC October 24, there are increases of downward long-wave radiation L_d and decreases of absorbed shortwave radiation S_a . The outgoing longwave radiation L_u increases slightly during nighttime but does not change much during daytime, apparently limited by the maximum temperature of melting point for the ice/snow surface. The difference in ground flux Q_{gr} between the CNTL and Blow-off simulations is minor. As a result, the net surface heat flux (Q) decreases by around 0.5 W/m^2 during daytime and increases slightly (by around 0.1 W/m^2) during nighttime with blowing snow.

The surface energy budget plays a critical role in determining surface mass balance (Lenaerts, Broeke, et al., 2012; Lenaerts & van den Broeke, 2012; Lenaerts, van den Broeke, et al., 2012; Miller et al., 2015; van Wessem et al., 2016, 2018). Thus, blowing snow's contributions to changes in surface mass balances are also examined (Figure 9b). The differences in each surface mass balance term between the CNTL and Blow-off simulations show that blowing snow helps to increase precipitation P via moistening the air, especially over areas with topographic lifting, by up to $\sim 2.4 \times 10^{-2} \text{ mm w.e. per day}$ averaged over the entire model domain except the ocean area, which is $\sim 0.15 \text{ Gt}$ integrated over the study area and entire 66-h study period. On the other hand, blowing snow reduces surface sublimation (S_s) by $\sim 1.9\text{--}4.8 \times 10^{-2} \text{ mm w.e. per day}$ for the same reason of decreased latent heat flux, amounting to an areal total of $\sim 0.22 \text{ Gt}$ for the 66-h period. In addition, enhanced cloud radiative cooling effect during daytime reduces melt and accordingly runoff R by up to $\sim 1.4 \times 10^{-2} \text{ mm w.e. per day}$. A slight increase in surface runoff R occurs during nighttime. The areal total of reduced runoff amounts to $\sim 0.06 \text{ Gt}$ for the 66-h period. The budget terms of blowing snow erosion E and deposition D which denote the divergence (negative) and convergence (positive) of blowing snow are only present in the CNTL simulation, shown as blue and orange solid lines in Figure 9b. Difference of blowing snow erosion E and deposition D represents the snow mass loss to the atmosphere via blowing snow sublimation, which is very small though. As a result, the net surface mass balance increases by $\sim 2.9\text{--}8.9 \times 10^{-2} \text{ mm w.e. per day}$, amounting to an areal total of $\sim 0.43 \text{ Gt}$ for the entire 66-h study period when blowing snow effects are included.

6.3. Conceptual Model of Blowing Snow Effects within WRF-Ice

Based on the examinations of the CNTL and Blow-off sensitivity simulations, a conceptual model depicting the blowing snow effects under different atmospheric and environmental conditions, including cold and dry conditions, warm and humid conditions, and mountainous environment, is proposed in Figure 10.

Over the cold and dry areas during clear sky conditions, blowing snow occurs from near-surface to 800-m AGL and blowing snow sublimation is mainly confined between 300- and 600-m AGL. Accordingly, the atmospheric temperature decreases and water vapor increases (Figure 10a). On the other hand, as the cold and moist air resulting from blowing snow sublimation moves to relatively warm and humid coastal areas, the stability of the lower atmosphere diminishes and a positive cloud anomaly occurs between 200- and 600-m AGL (Figure 10b). The longwave and shortwave cloud radiative forcings vary with the height and thickness of enhanced clouds. This further impacts the air and surface temperatures. Over the area with lower and thicker clouds, the warming effect from enhanced longwave radiation exceeds the cooling effect from reduced shortwave radiation. In this case, the air and surface temperatures under the clouds increase. However, for the higher and thinner clouds, the warming effect becomes weaker (Figure 10b). Furthermore, in mountainous areas with enhanced lifting, blowing snow can enhance precipitation by increasing water vapor via snow sublimation (Figure 10c). Above all, the impact of blowing snow and its associated physical processes vary with the atmospheric and geographic conditions.

It should be noted that this conceptual model derives from the simulation case study, where the major physical processes associated with blowing snow effects are synthesized and illustrated for conditions specific to the simulated event. The atmospheric state in reality is much more complex and varies with season and location. Continuing studies with more cases are needed to confirm and improve the findings presented here.

7. Summary and Conclusions

This study evaluates the performance of the snow/ice enhanced WRF-ice model and the blowing snow effects within WRF-ice over the Antarctic Peninsula by modeling a strong cyclone system that impacted the entire Antarctic Peninsula during October 23–26, 2017. The lifespan of the cyclone system captured by the WRF-ice model in the CNTL simulation indicates that when a synoptic cyclone center crossed the Drake Passage and impinged on the west side of the Peninsula, a substantial reconstitution occurred to the leeward side. This created a mesoscale cyclone embodied within the synoptic cyclone. The key simulated near-surface meteorological variables, including SLP, 2-m temperature, dew point, and 10-m wind speed within the CNTL simulation are verified against observations at nine ISD/NCEI stations over the Antarctic Peninsula. The verification shows that the WRF-ice reasonably captures the observed near-surface conditions, although there still exist some discrepancies in 2-m temperature, dew point, and 10-m wind speed. This is especially true for the stations along the coastal area with highly complex features. Moreover, the simulated surface temperature is compared against the MODIS MYD land ice and sea ice temperature products. Comparisons indicate that the surface temperature features over both land and sea ice are well reproduced by WRF-ice model. However, some cold bias of land ice surface temperature exists at night.

To further examine blowing snow effects within the WRF-ice model, a sensitivity simulation (Blow-off) is conducted, in which blowing snow processes are deactivated. Using the results of both the CNTL and Blow-off simulations, the blowing snow impacts on the lower atmosphere and surface mass balance are investigated. Differences between the CNTL and Blow-off simulations suggest that blowing snow enhances water vapor in the lower atmosphere under cold and dry atmospheric conditions. Besides, it can generate enhanced clouds under warm-humid conditions. It can also enhance precipitation when sufficient moistening and upward lifting are all present in the mountainous environment during the cyclone event. Surface winds and temperature also increase with the enhanced clouds. Based on our analyses, a conceptual model illustrating the blowing snow effects under different conditions mentioned above is proposed. Again, it should be noted that blowing snow effects synthesized in the conceptual model are derived from conditions specific to the simulated event. Continuing studies with cases from different locations and weather conditions are needed to further confirm and improve the model.

Surface energy budget analyses suggest that downward shortwave, downward longwave, and outgoing longwave radiation are the dominant components responsible for surface heat fluxes. Blowing snow can cause negative anomalies (up to -2 W/m^2) in latent heat flux and positive anomalies (up to 2 W/m^2) in sensible heat flux due to the cooling and moistening effects resulting from blowing snow sublimation, when the surface is warmer and wetter than the air above during the study period. The downward longwave radiation increases and shortwave radiation decreases along with enhanced clouds associated with blowing snow sublimation. Consequently, the net surface heat flux decreases by around 0.5 W/m^2 during daytime and increases by around 0.1 W/m^2 during nighttime all due to the presence of blowing snow. Surface mass balance analyses indicate that blowing snow contributes to intensified precipitation ($\sim 0.15 \text{ Gt}$ integrated over the entire model domain except the ocean area during the 66-h study period from 0600 UTC October 23 to 0000 UTC October 25, 2017) through moistening of the atmosphere, especially over the mountain areas. Surface sublimation decreases by an areal total of $\sim 0.22 \text{ Gt}$ with blowing snow for the same reason that causes the latent heat flux decrease. Surface runoff decreases mainly during daytime due to reduced downward solar radiation at the surface with an areal total of $\sim 0.6 \text{ Gt}$ for the entire study period. Blowing snow erosion and deposition play important roles in snow redistribution, however their net impact on the domain averaged mass balance is negligible. Thus, blowing snow contributes to a positive net surface mass balance by around 0.43 Gt integrated over the study area and the study period via increased precipitation, reduced runoff, and reduced surface sublimation.

Findings in this study indicate that the neglect of blowing snow and its interactions with the atmosphere can cause systematic biases in the forecasts of near-surface atmospheric fields, surface snow characteristics, and surface mass balances. These interactions include blowing snow transport and sublimation in numerical weather prediction models. While the WRF-ice simulation results are encouraging, these findings and the derived conceptual model of blowing snow effects are based on a single simulated case. Considering that atmospheric states in reality are much more complicated, further studies on different cases over various locations for a longer study period are needed. Because of the lack of in situ observations of blowing snow processes, the evaluations of blowing snow impacts within this study carry a degree of uncertainty. More reliable observations of blowing snow and regional climatology studies of blowing snow impacts are beneficial for a better understanding of its interactions with the lower atmosphere and surface.

Data Availability Statement

Modeling data produced by this study as well as the forcing and verification data are available online (<https://doi.org/10.7910/DVN/ZXZA7M>).

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