Abstract—In this paper, we present model and measurement results for time-series angular dependencies of C-band HH and VV normalized radar cross-sections (NRCS) over first-year snow-covered sea ice during a winter–spring transition period. Experimental scatterometer and physical data were collected near Cambridge Bay, Nunavut, Canada, between May 20 and May 28, 2014, covering a severe storm event on May 25. We use the small perturbation scattering theory to model small-scale surface scattering, the Mie scattering theory to estimate the level of volume scattering in snow, and the Kirchhoff physical optics model to compute the large-scale surface scattering component. We observed good agreement between the model and experimental HH and VV NRCS. Before the storm, $R^2$ between model and experimental NRCS was 0.88 and 0.82 for VV and HH, respectively. After the storm, $R^2$ was 0.81 and 0.78 for VV and HH, respectively. Our model results suggest an overall increase in surface roughness after the storm event, supported by LiDAR measurements of the snow surface topography. Before the storm, the large-scale and small-scale surface scattering from the air-snow interface as well as volume scattering components dominated. After the storm, the large- and small-scale scattering contributions increased, while the volume scattering component considerably dropped. We attribute these effects to the increase in surface roughness and snow moisture content during the poststorm period. Our results could aid in interpretation of time-series synthetic aperture radar images with respect to physical properties of snow and ice during the winter–spring transition period.

Index Terms—Electromagnetic wave scattering, Kirchhoff model, layered media, rough interfaces, small perturbation theory, snow-covered sea ice, surface and volume scattering.

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I. INTRODUCTION

The Arctic sea ice cover has been undergoing significant changes over the past three decades [1]. An overall reduction in sea ice volume has been accompanied by a general replacement of multiyear (MY) ice with seasonal first-year (FY) ice, which has impacted a range of physical, chemical, and biological processes in the Arctic marine ecosystem [2]–[4]. Furthermore, the increasing number of hazardous ice features in the Arctic Ocean, such as ice islands and thick MY ice, represents a significant threat to ongoing industrial infrastructure development in the Arctic [5]. It is, therefore, important that short-term (weekly) and long-term (interannual) changes in the Arctic sea ice cover can be accurately monitored at both local and regional scales.

Active microwave remote sensing has proved to be an efficient tool for studying both the dynamic [6]–[9] and thermodynamic [10]–[14] characteristics of snow-covered sea ice. Observations from the ongoing synthetic aperture radar (SAR) satellite missions, such as Canadian RADARSAT-2 (C-band), European Sentinel-1A and 1B (C-band), TanDEM-X (X-band), and Japanese PALSAR-2 (L-band), currently provide the most effective and frequent data over Arctic sea ice for both scientific and operational tasks.

However, the Earth observation paradigm at a number of space agencies is currently being shifted from a single complex multifunctional satellite platform to several satellite platforms (often called satellite constellations). Three-satellite Canadian RADARSAT Constellation mission [15] in addition to the presently operational two-satellite European Sentinel-1 [16] missions will further increase the quantity and repeat sampling frequency of C-band SAR imagery over the Arctic Ocean. This anticipated increase in the temporal resolution of satellite data will provide a user with the potential for monitoring, at high spatial and temporal resolution, the evolution of the normalized radar cross-section (NRCS) over a specific region in the Arctic Ocean. However, in spite of this increase in data provision, interpretation of such time-series data is limited by our understanding of the complex processes relating electromagnetic scattering within the snow-covered sea ice to the geophysical and thermodynamic characteristics of the system, particularly in a changing Arctic environment.
Electromagnetic scattering models play a key role in linking physical characteristics of snow-covered sea ice and the NRCS response. A number of studies have been devoted to examining and comparing experimental and model NRCS from sea ice with respect to the physical properties under various environmental conditions [17]–[19]. Previous studies on time-series evolution of NRCS were conducted for artificially grown young sea ice under controlled environment [20], [21]. A recent time-series study of NRCS evolution with subhour temporal resolution from artificially grown frost flowers was reported in [22]. Several studies were devoted to the examination of time-series NRCS response from the Arctic sea ice acquired by spaceborne SAR [11], [14], [23] and spaceborne scatterometer [24] observations. In these studies, the temporal resolution of spaceborne NRCS observations varied from several hours to several days.

However, modeling and measurement of time-series C-band NRCS response with a high temporal resolution (i.e., at a subhour scale) over the natural FY snow-covered sea ice during a seasonal shift have not been conducted. At the same time, such studies are particularly important to understand changes in NRCS and scattering mechanisms in relation to changes in physical properties of snow and ice under various environmental conditions.

In [12], a model for electromagnetic wave scattering from rough boundaries interfacing inhomogeneous media based on the first-order approximation of the small perturbation theory was proposed. This model demonstrated good agreement with ship-based scatterometer data collected over winter snow-covered sea ice in the southern Beaufort Sea [13] where the snow grains were not large enough to generate a significant level of volume scattering. In spring, with enlarging snow grains due to the snow metamorphism processes, the volume scattering in snow cover becomes important [25], [26]. Furthermore, the small-scale surface roughness could be superimposed on large scale changes in snow topography produced by winds, which characteristically become more pronounced in the spring [27]. For example, sastrugi are common large-scale snow features caused by wind erosion and, therefore, align parallel with the wind direction [28]. The large-scale roughness creates an additional scattering mechanism that may play an important role in the total scattering. Studying contributions of surface scattering (from both the small- and large-scale surface roughness) and volume scattering during the spring transition period are particularly important to better understand the impact of individual physical components of the system under changing environmental conditions. We note that, in this paper, we deal with smooth undeformed FY sea ice in the Canadian Arctic Archipelago; therefore, we assume that only the air–snow interface has both the small- and large-scale roughness components while the ice–snow interface has small-scale roughness only. The small-scale roughness has root mean square (rms) height at a scale of millimeters (∼1–3 mm) and correlation length at a scale of centimeters (∼1–5 cm). The large-scale roughness has rms height at a scale of centimeters (∼2–6 cm) and correlation length at a scale of meters (3–8 m). Surface roughness could be described by stationary or nonstationary (fractal) geometry [29]. Surface roughness measurements, in this paper, indicate that the power spectral density of the roughness over undeformed sea ice is two-scale; therefore, surface roughness in our case could be described by stationary random functions. We also note that in the recent study [30], it was shown that the fractal properties of sea ice roughness observed in some of the past studies [31] may have been artificially introduced by the field data sampling technique (i.e., limited sampling interval or extent).

A commonly used integral equation model (IEM) [32] has a larger range of validity for roughness parameters compared with the small perturbation theory. However, the IEM model is found to be limited by the fact that it is applied to a rough surface on top of a homogeneous half-space. A solution for radar backscatter from a submerged rough surface (i.e., snow–ice rough interface) could not be derived within the IEM framework only; therefore, often, a combination of the IEM and the radiative transfer theory is implemented. Furthermore, the IEM model does not account for a layered vertical structure of the medium (i.e., when the dielectric constant is a function of depth).

A number of studies [33]–[35] use a combination of the radiative transfer theory and surface scattering theories (such as small-perturbation theory and IEM). The surface scattering solutions are used in order to adequately describe electromagnetic fields at boundaries. The use of radiative transfer theory allows for account for multiple scattering effects. On the other hand, description of wave interference, which could be quite strong (e.g., within a snow layer), might not be accurately performed as the phase information is not considered by such theories.

A recently published study [36] uses the same experimental data set as in this paper in order to investigate feasibility of retrieval of physical properties of snow and ice. In their study, only the period of time with stable environmental conditions was considered, while the second part of the experiment with changing environmental conditions was not analyzed. A single temporarily averaged snow and sea ice profile representing the period with stable conditions was a subject for inversion. For the inversion process, roughness information and several measured physical parameters of snow and ice served as a priori information. In [36], volume and small-scale surface scattering components were considered, while the large-scale surface scattering was not considered. Evaluating temporal evolution of scattering mechanisms was not a research objective in [36].

In this paper, we pursue the following main objectives: 1) develop (by combining and extending available models) and validate a scattering model accounting for scattering from small- and large-scale surface roughness as well as volume scattering from snow grains within an inhomogeneous snow cover; 2) to examine whether variations in time-series C-band scatterometer data collected over snow-covered FY sea ice during a winter–spring transition period can be accurately simulated by this model; and 3) to investigate how the relative contributions from various scattering mechanism evolve under changing environmental conditions.
II. MICROWAVE SCATTERRING THEORY

The geometry of electromagnetic wave scattering from snow-covered sea ice is shown in Fig. 1. The air–snow interface represents a superposition of small- and large-scale roughness components, i.e., at the scale of snow grains (0.001–0.01 m) and snow drifts (1–10 m), respectively. Both of them can be described by uncorrelated stationary random functions. The snow–sea ice interface is characterized by a small-scale roughness, which can be defined through a stationary random function. Complex dielectric constant (CDC) of snow and sea ice is described by \( \varepsilon_s(z) \) and \( \varepsilon_i(z) \), which are continuous functions of the vertical coordinate \( z \). Also, snow grain sizes in snow vary over the vertical coordinate. For a given sublayer \( n \) (from \( -d_{n-1} \) to \( -d_n \), \( n = 1 \ldots N_u \) radius of snow grains is assigned to be \( r_n \). The rough ice–water interface at the base of the sea ice cover could be potentially included in the scattering geometry. However, in this paper, we deal with FY ice with thickness \( \sim 1.8 \) m, which has high total attenuation (in C-band) due to the presence of brine inclusions throughout the ice layer.

An electromagnetic plane wave is scattered by the air–snow and snow–ice rough interfaces and by the snow grains at different depths. Therefore, the total radar backscatter coefficient can be expressed as a superposition of the following four components:

\[
\sigma_{pp} = \sigma_{pp}^{(\text{air-snow})} + \sigma_{pp}^{(\text{snow-ice})} + \sigma_{pp}^{(\text{vol})} + \gamma \sigma_{pp}^{(\text{air-snow})},
\]

Subscript \( p \) indicates either vertical (\( p = V \)) or horizontal polarization (\( p = H \)). In (1), \( \sigma_{pp}^{(\text{air-snow})} \) represents the surface scattering contribution from the small-scale component of the air–snow interface; \( \sigma_{pp}^{(\text{vol})} \) is the surface scattering contribution from the large-scale roughness of the air–snow interface; \( \sigma_{pp}^{(\text{snow-ice})} \) represents surface scattering contribution from the rough snow–ice interface; \( \gamma \) is the factor accounting for the volume scattering loss within the snow layer; \( \sigma_{pp}^{(\text{air-snow})} \) is the volume scattering component originating from the snow grains within the snow layer.

A. Surface Scattering From Small-Scale Roughness

To model the small-scale surface scattering components in (1), we use the small perturbation scattering theory for wave scattering from rough interfaces separating layered media. This model was introduced in [12] and applied to winter snow-covered sea ice in [13]. According to [12] and [13], NRCS for vertical and horizontal polarizations in the monostatic case from the air–snow and snow–ice interfaces is expressed as follows.

Small-Scale Air–Snow Roughness:

\[
\sigma_{HH}^{(\text{air-snow})} = \frac{k_0^4|\Delta \varepsilon_s|^2}{4\pi} [(1 + \Re \chi_H(q_0))^4 \tilde{K}_s(-2q_0)]
\]

Small-Scale Snow–Ice Roughness:

\[
\sigma_{HH}^{(\text{snow-ice})} = \frac{k_0^4|\Delta \varepsilon_i|^2}{4\pi} [(1 + \Re \chi_H(q_0))^4 \tilde{K}_s(-2q_0)]
\]

\[
\sigma_{VV}^{(\text{snow-ice})} = \frac{k_0^4|\Delta \varepsilon_i|^2}{4\pi} \left[ \frac{\sin^2 \Theta_0}{\varepsilon_i(0)} [1 + \Re \chi_V(q_0)]^2 + [1 - \Re \chi_V(q_0)]^2 \cos^2 \Theta_0 \right]^{1/2} \times \tilde{K}_s(-2q_0).
\]

Note that the cross-polarization components in the monostatic case within the first-order approximation of the small perturbation theory are zeros.

In (2)–(5), \( k_0 \) is the wavenumber in vacuum; \( \Theta_0 \) is the incidence angle; \( \varepsilon_i(-d) \) is the CDC of ice at the snow–ice interface; \( \varepsilon_s(-d) \) is the CDC of snow at the snow–ice interface \( z = -d \); \( \Delta \varepsilon_i = \varepsilon_i(0) - 1 \) is the dielectric contrast between snow and ice at the air–snow interface; \( \Delta \varepsilon_s = \varepsilon_s(-d) - \varepsilon_s(0) \) is the dielectric contrast between ice and snow at the snow–ice interface; \( q_0 \) is the projection of the wave vector in the air onto the horizontal plane.

In (2) and (3), \( \Re \chi_{H,V} \) values are complex reflection coefficients at horizontal and vertical polarizations from the entire snow-covered sea ice structure. In (4) and (5), the auxiliary coefficients \( L_{H,V} \) and \( M_{H,V} \) are defined as follows:

\[
L_{H,V}(q_0) = \frac{\omega_0(q_0)}{\omega_0(q_0) - r_{H,V}(q_0) R_{H,V}(q_0)} [1 + r_{H,V}(q_0)]
\]

\[
M_{H,V}(q_0) = \frac{\omega_0(q_0)}{k_0} \frac{T_{H,V}(q_0)}{1 - r_{H,V}(q_0) R_{H,V}(q_0)} [1 - r_{H,V}(q_0)]
\]

where \( \omega_0(q_0) = (k_0^2 - q_0^2)^{1/2}, u_s(q_0) = (k_0^2 \varepsilon_i(-d) - q_0^2)^{1/2}, \) and \( r_{H,V} \) and \( T_{H,V} \) are complex reflection and transmission coefficients for the inhomogeneous snow layer when the wave is incident from the half-space with CDC \( \varepsilon_s(-d) \) at horizontal and vertical polarizations; \( r_{H,V} \) values are reflection coefficients from the inhomogeneous sea ice when the wave is
incident from the half-space with CDC $\varepsilon_r(-d)$ at horizontal and vertical polarizations.

The reflection and transmission coefficients associated with inhomogeneous media can be computed using the invariant embedding approach [37], [38]. A detailed numerical recursive scheme for this method can also be found in [13]. This scheme uses the dielectric profiles of snow and sea ice as inputs. To calculate the reflection coefficients for sea ice, we used a dielectric profile of the upper 50-cm layer of ice, as there is no need to calculate reflection coefficients for the entire ice thickness (more than 1 m) as the wave attenuates significantly within sea ice (in C-band).

In (2)–(5), $K_{s1}$ and $\tilde{K}_i$ are the spatial power spectral densities (i.e., Fourier transforms of the autocorrelation functions) of the small-scale air–snow–ice rough interfaces, respectively. LiDAR measurements conducted in this paper and in [30] suggest that the exponential function provides best fit to the autocorrelation functions of snow and ice surfaces. Therefore, the autocorrelation functions of the small-scale air–snow rough interface and snow–ice rough interface can be described by an exponential dependence

$$K(\rho) = h^2 \exp\left(-\frac{\rho}{L}\right)$$

where $\rho$ is the magnitude of the position vector in the horizontal plane, and $h$ and $L$ are rms height and correlation length of a rough interface, respectively. Cases $K(\rho) = K_{s1}(\rho)$, $h = h_{s1}$, and $L = L_{s1}$ correspond to the small-scale component of the air–snow interface. Cases $K(\rho) = K_i(\rho)$, $h = h_i$, and $L = L_i$ correspond to the snow–ice rough interface. Spectrum of (8) can be found as follows [32]:

$$\tilde{K}(q) = \frac{2\pi L^2 h^2}{(1 + q^2 L^2)^{1.5}}.$$  

Generally speaking, the large-scale component of the air–snow rough interface can be considered in the small-perturbation theory through integration of the NRCS multiplied by the probability distribution function (pdf) of slopes over all possible slopes in two dimensions as shown in [39]. However, pdfs of large-scale slopes in our case are very narrow as discussed in Section V-B. Therefore, our results obtained by integrating over all possible slopes practically do not change the NRCS calculated for a zero-sloped surface. Hence, in our case, only zero slopes contribute to the total NRCS response from the small-scale roughness.

B. Surface Scattering From Large-Scale Roughness

We model scattering from large-scale roughness based on the Kirchhoff (physical optics) model [39], [40]. According to this model, the NRCS from a large-scale rough surface can be written as follows:

$$\sigma_{pp}^{(\text{air-snow})} = k_0^2 \cos^2 \Theta_0 |(i_{pp}(\Theta_0))|^2 \exp\left[-(2k_0 h_{s1} \cos \Theta_0)^2\right] \times \frac{1}{\pi} \sum_{m=1}^{\infty} \frac{4k_0^2 \cos^2 \Theta_0 m!}{m!} \tilde{K}_{s1}^{(m)}(2k_0 \sin \Theta_0)$$

where $\tilde{K}_{s1}^{(m)}$ is the Fourier transform of the $m$th power of the surface autocorrelation function. Our LiDAR measurements revealed that the large-scale topography in this paper can be described by the exponential autocorrelation function. Therefore, $\tilde{K}_{s1}^{(m)}$ can be written as follows:

$$\tilde{K}_{s1}^{(m)}(q) = \frac{2\pi m h_{s1}^m q^2 L_{s1}^2}{(m^2 + q^2 L_{s1}^2)^{1.5}}$$

where $h_{s1}$ and $L_{s1}$ are the rms height and correlation length of the large-scale roughness component of the air–snow interface, respectively. In our formulations, we assume that the rough surface is isotropic. According to [39], the validity conditions for the Kirchhoff approximation model can be formulated as follows:

$$k_0 L_{s1} > 6, \quad L_{s1}^2 > 0.8 h_{s1}^2 \lambda$$

where $\lambda$ is the wavelength of the incident wave. The first condition in (12) indicates that the correlation length should be equal or larger than the wavelength. The second condition in (12) (specifically written for the exponential autocorrelation function) indicates that relatively high rms heights can be tolerated as long as the average radius of curvature is larger than the wavelength. Therefore, the rms height of large-scale roughness in this model can be comparable to the wavelength. Our surface roughness measurements (presented below) confirm that validity conditions (12) are well met. We note that it is not possible to determine rms slope included in the validity conditions in [39] as we deal with the exponential correlation function.

C. Volume Scattering

Snow grains at the bottom of the snow layer (i.e., basal layer) normally enlarge by the end of winter due to metamorphism processes [26] and are brine wetted due to icing processes throughout the season [25]. Therefore, volume scattering in C-band within a spring snow cover may be considerable. Size of snow grains in the basal layer is relatively large with respect to the wavelength and usually decreases toward the snow surface. To quantify the level of volume scattering within the snow cover, we consider $N_v$ layers over the snow depth. Since the snow grains in a given layer $n$ (such that $n = 1 \ldots N_v$) are randomly oriented, they can be approximated by a collection of spheres with an effective radius $r_{pp}$. An equation for volume scattering contribution from a single volume scattering layer is presented in [32] and [41]. We extended this formulation to several volume scattering layers with different physical characteristics as follows. The total volume scattering from all $N_v$ layers can be presented as a sum of backscatter returns from an individual layer

$$\sigma_{pp}^{(\text{vol})} = \sum_{n=1}^{N_v} \sigma_{pp}^{(n)}$$

where $\sigma_{pp}^{(n)}$ is the volume backscattering coefficient from layer $n$, which can be expressed as follows:

$$\sigma_{pp}^{(n)} = \frac{1}{2} \left[ \frac{k_{s1}^{(n)}}{k_{s1}^{(n)} + k_{sh}^{(n)} + k_{as}^{(n)}} \right] \left[ T_{pp}^{(n)}(\Theta_0, \Theta_0) \right]^2 \left[ T_{pp}^{(n)}(\Theta_0, \Theta_0) \right]^2 \times \cos \Theta_0 F_{pp}^{(n)} g_{pp}^{(n)}(n-1).$$
In the last equation

\[
\delta^{(n)} = \left[ 1 - \exp \left( -\frac{2\Delta_n}{\cos \Theta_{\text{in}}} \left( \kappa_s^{(n)} + \kappa_{\text{sh}}^{(n)} \right) \right) \right] \quad (15)
\]

\[
\gamma^{(n)} = \prod_{i=1}^{n} \exp \left[ -\frac{2\Delta_i}{\cos \Theta_{\text{in}}} \left( \kappa_s^{(i)} + \kappa_{\text{sh}}^{(i)} \right) \right] \quad (16)
\]

To describe scattering from a single grain within the snow layer, we rely on the Mie scattering theory that represents an exact solution of electromagnetic wave scattering from a sphere with an arbitrary radius [42]. In this case, the scattering and absorption coefficients \( \kappa_s^{(n)} \) and \( \kappa_{\text{sh}}^{(n)} \) within layer \( n \) can be written as follows:

\[
\kappa_s^{(n)} = N_n \pi r_n^2 Q_s \quad (17)
\]

\[
\kappa_{\text{sh}}^{(n)} = N_n \pi r_n^2 Q_\ell \quad (18)
\]

where \( N_n \) is the number density of grains, which is \( N_n = 3\rho_n/4\pi r_n^3 \); \( \rho_n \) is the volumetric fraction of ice grains within layer \( n \). \( Q_s \) and \( Q_\ell \) are the scattering and absorption efficiencies, respectively. Equations for the efficiencies within Mie’s theory can be found in [43]. In our simulations, we accepted the dielectric constant of ice grains to be equal to the dielectric constant of pure ice, i.e., 3.15.

In (14), \( P_{pp}^{(n)} \) represents the phase function for the backscattering case. A formulation for four elements of the phase matrix (HH, VV, HV, and VH) is given in [41] and [44]. Substituting the scattered electric field in far zone [43] into this formulation, it is possible to obtain the phase functions as follows:

\[
P_{HH}^{(n)} = P_{VV}^{(n)} = \frac{4\pi N_n}{\kappa_s^{(n)}} \left| E_s \right|^2, \quad (19)
\]

where \( E_s \) is the scattered electric field in the backward direction in far zone. A formulation for the electric scattered field in far zone can be found in [43].

The additional absorption coefficient \( \kappa_{\text{sh}}^{(n)} \) in layer \( n \) associated with the loss within the host medium can be written as follows:

\[
\kappa_{\text{sh}}^{(n)} = 2k_0 \text{Im} \sqrt{\varepsilon_{\text{sh}}^{(n)} (1 - \rho_n)}, \quad (20)
\]

where \( \varepsilon_{\text{sh}}^{(n)} \) is the CDC of the host medium within the layer \( n \). We compute this value as an average of \( \varepsilon_s(z) \) function within the interval from \( -d_{n-1} \) to \( -d_n \).

In (14), \( T_{pp}^{(n)}(\Theta_0, \Theta_j^{(n)}) \) is the complex transmission coefficient with respect to the electric field (for horizontal polarization, \( p = H \)) or the magnetic field (for vertical polarization, \( p = V \)) when the wave is incident from the air and propagates through the snow layers located above a given layer \( n \). \( T_{pp}^{(n)}(\Theta_0, \Theta_j^{(n)}) \) is the transmission coefficient similar to the previous one, but when the wave propagates from the layer \( n \) toward the air half-space. In (14), \( |T_{pp}^{(n)}(\Theta_0, \Theta_j^{(n)})|^2 |T_{pp}^{(n)}(\Theta_0^{(n)}, \Theta_0)|^2 \) represents a product of power transmission coefficients for electromagnetic wave propagating toward and outward layer \( n \).

In (15) and (16), \( \Delta_n = d_n - d_{n-1} \) is the thickness of layer \( n \) and \( \cos \Theta_{\text{in}}^{(n)} \) is the cosine of the angle of refraction within layer \( n \), which can be found according to Snell’s law.

The exponent in (15) denotes the two-way attenuation within layer \( n \). Product of exponents \( \gamma^{(n-1)} \) given by (16) represents volume scattering loss within the layers located above layer \( n \). The loss due to volume scattering in the entire snowpack is \( \gamma = \gamma^{(N_n)} \).

We note that brine and water droplets in snow are very small compared with the wavelength in C-band, and their number density is very small as well. Therefore, volume scattering from brine and water droplets is not considered here.

III. EXPERIMENTAL DATA

Time-series C-band polarimetric scatterometer and physical data were collected on smooth landfast FY sea ice in Dease Strait near Cambridge Bay, Nunavut, Canada, between May 20 and May 28, 2014, as a part of the interdisciplinary ICE-CAMPS field program. A storm event occurred on May 25 and divided the experiment into pre-storm and post-storm phases. The experimental site was located at 69.026°N, 105.337°W. Sea ice thickness was around 1.8 m and did not change appreciably throughout the experiment. Snow thickness at our physical sampling stations varied from 13 to 24 cm with an average value of 18 cm. The ice thickness was measured by drilling a 2-inch-diameter hole and deploying an ice thickness measurement tape. The snow thickness was measured by orienting a snow ruler perpendicularly to the ice surface across the face of the snow pit wall.

A. Scatterometer Data

A fully polarimetric C-band (5.5 GHz) scatterometer system was mounted at 3.1-m height on a stationary platform frozen in the ice surface as shown in Fig. 2. The platform design allowed snow to move almost freely through the tower structure, in order to minimize accumulation in the vicinity of the platform. Scatterometer measurements of snow-covered sea ice were collected at different incidence angles (with respect to nadir) in the elevation from 20° to 60° with 5° step. At each incidence angle, the effective NRCS for a −30° to 30° swath in the azimuth direction was computed, where 0° azimuth corresponds to the direction perpendicular to the
tower front side facing the North. The half power beamwidth of the scatterometer antenna is 5.4°. The estimated scatterometer footprint area varied with the incidence angle from 0.08 m² at 20° to 0.54 m² at 60°. The estimated area on the ground corresponding to the effective NRCS (from 60° azimuth swath) varied from 0.39 m² at 20° to 6.67 m² at 60° incidence angle.

The scatterometer was run continuously except for time periods where instrument operation was not possible due to storm weather conditions, precipitation events, or owing to logistical issues. An absolute calibration of the scatterometer was performed on May 28 using a trihedral corner reflector after the experiment was finished. Calibration errors that may have a biasing effect (due to range inaccuracies, and inaccurate estimates of incidence angle) are assumed to be ±0.5 dB according to [45]. More information on the scatterometer system and its operation in the field can be found elsewhere [18], [22], [45].

B. Physical Data

Throughout the nine-day experiment, we conducted detailed physical sampling of snow and sea ice in support of the time-series scatterometer observations. The frequency of physical sampling varied from one to three times a day. Twelve physical sampling sessions were conducted before the storm, and eight sampling sessions were conducted during the poststorm phase of the experiment. The size of snow grains was measured at 2-cm vertical intervals across the snow profile, by photographing disaggregated gains on a grid plate. Snow density profiles were captured with 2-cm resolution using a snow density cutter and a gravimetric approach. Snow salinity profiles were recorded by melting each snow sample and measuring salinity of the melt water at room temperature with an HACH SENSION5 portable conductivity meter (accuracy of ±0.1 ppt). The temperature profile of snow was measured with 2-cm resolution using a temperature probe (Traceable Digital Thermometer, Control Company, accuracy of ±0.01 °C). Volumetric contents of brine and pure ice in snow were estimated through the measured snow density, salinity, and temperatures as shown in Section IV-A.

Temperature and salinity of sea ice as the functions of depth were captured by examining an ice core. The temperature of the extracted ice core was measured at 5-cm resolution by drilling a hole and inserting a temperature probe. Then, the ice core was cut into 5-cm lengths placing them in sealed plastic bags and melting them and equilibrating to room temperature. To measure the bulk salinity profile, salinity of each melted core was cut into 5-cm lengths placing them in sealed plastic bags and melting them and equilibrating to room temperature. The extracted ice core was measured at 5-cm resolution by a Leica C10 laser scanner (precision <0.1 mm). Volumetric contents of brine and pure ice in snow were estimated through the measured snow density, salinity, and temperatures as shown in Section IV-A.

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Temperature and salinity of sea ice as the functions of depth were captured by examining an ice core. The temperature of the extracted ice core was measured at 5-cm resolution by drilling a hole and inserting a temperature probe. Then, the ice core was cut into 5-cm lengths placing them in sealed plastic bags and melting them and equilibrating to room temperature. To measure the bulk salinity profile, salinity of each melted core was cut into 5-cm lengths placing them in sealed plastic bags and melting them and equilibrating to room temperature. The extracted ice core was measured at 5-cm resolution by a Leica C10 laser scanner (precision <0.1 mm). Volumetric contents of brine and pure ice in snow were estimated through the measured snow density, salinity, and temperatures as shown in Section IV-A.

C. Meteorological Data

Various meteorological parameters were continuously logged over the duration of our experiment at an on-ice tower located near the scatterometer experimental site. Collected atmospheric parameters included air temperature, air pressure, wind speed, and relative humidity. Shortwave and long-wave radiation fluxes were also measured by the meteorological station. Furthermore, temperatures in the snowpack at 0, 3, 6, and 9 cm from the snow–ice interface, temperature at the air–snow interface, as well as a temperature profile of the ice cover with 5-cm resolution were measured throughout the experiment cycle using thermocouple arrays (Type T, Omega Inc.).

D. Surface Roughness

To capture changes in roughness of the air–snow rough interface, both small- and large-scale components of surface roughness were measured on May 20, 24, 27, and 30 at a rectangular noninvasive site located approximately 0.4 km to the North West of the scatterometer site. Since our experiment was conducted on smooth and undeformed FY sea ice, the snow surface roughness conditions did not vary considerably around our scatterometer site. Therefore, we assumed that any significant changes in surface roughness captured at the LiDAR measurement site could adequately indicate changes in snow roughness at the scatterometer site.

Terrestrial LiDAR measurements were sequentially conducted from four platforms placed at the site corners, using a Leica C10 terrestrial laser scanner (precision <1 mm). By this technique, measurements of the small-scale roughness could be acquired (at a sampling interval of around 0.5 cm) within small 5 m × 5 m areas close to the sampling platforms, in addition to measurements of the large-scale roughness (at a sampling interval of around 5 cm) across the entire 80 m × 160 m site. The Gaussian beam divergence of the Leica C10 laser scanner is 0.14 mrad and the maximum range to the snow surface was 6 and 90 m for the small- and large-scale measurements, respectively. Based on these scanning parameters, the maximum size of the laser footprint on the target was 0.54 mm² for the small-scale measurements and 125 mm² for the large-scale measurements. Raw LiDAR data were resampled to regular grids and detrended with respect to their principal plane using orthogonal least-squares regression [30]. Large-scale topography was also removed from the small-scale roughness grids using a second fast Fourier transform-based detrending technique. Processed LiDAR grids were then used to calculate estimates for surface roughness parameters using the method of [30], which were then calibrated [46] to produce accurate roughness observations. The estimated surface roughness parameters from LiDAR measurements included rms height and correlation length of both the small- and large-scale snow roughness components. Note that surface roughness of the snow–ice interface could not be measured by the LiDAR.

E. Satellite Data

Two dual-polarization (HH–HV) RADARSAT-2 ScanSAR images with 50-m resolution over our experimental site were available within the duration of the experiment. The first image was taken on May 24, 2014, 7:50 a.m. (LST) and the second image was acquired on May 25, 2014, 7:21 a.m. (LST). A region of interest with 11×11 pixels encompassing the experimental site and surrounding area was analyzed in both images. Incidence angle, average HH and HV NRCS, and corresponding standard deviations over the 11×11 pixels region of interest were calculated.
IV. DIELECTRIC MODELING

In this section, we present dielectric mixture models for modeling CDC of snow and sea ice. For both media, a refractive dielectric mixture model (i.e., power model with a power of 0.5) is used similar to [13].

A. Snow

A dielectric mixture model for estimating CDC of wet snow is proposed in [47] and [48]. At the same time, direct measurements of CDC of brine-wetted snow in the microwave band and a corresponding dielectric mixture model are not available in the literature. The mixture model for CDC of wet snow [47] could be possibly extended to modeling CDC of the brine-wetted wet snow. However, implementation of such an extension is not straightforward, and it would require a special investigation. The refractive dielectric mixture model for dry snow was used in [49]. In general, the refractive mixture model has been proven to be the most applicable for wet soils [50]. We note that the dielectric properties of pure ice and soil rock particles are close to each other, and the ranges of porosity in snow and soil are comparable as well. Therefore, we assume that the refractive dielectric mixture model could be applied to the brine-wetted snow.

To model CDC of snow, we implement the power dielectric mixture model as follows:

\[
\varepsilon_s = \varepsilon_{ds}^a + (\varepsilon_{bs}^a - 1) W_b + (\varepsilon_{ws}^a - 1) W_w, \tag{21}
\]

where \( W_b \) is the volumetric fraction of brine, \( W_w \) is the volumetric fraction of fresh water, \( \varepsilon_{bs} \) is the dielectric constant of brine, which can be estimated as a function of temperature and frequency according to Stogryn and Desargant [51], and \( \varepsilon_{ws} \) is the CDC of fresh water in snow, which can be estimated from [52].

In (21), CDC of \( \varepsilon_{ds} \) dry snow (i.e., when \( W_b = W_w = 0 \)) is written as follows:

\[
\varepsilon_{ds}^a = 1 + (\varepsilon_{pi}^a - 1) W_{pi} \tag{22}
\]

where \( \varepsilon_{pi} = 3.15 \) is the dielectric constant of pure ice [40].

Following the recent study [13], we implement the refractive dielectric mixture model with \( \alpha = 0.5 \). Volumetric content of brine \( W_b \) is estimated according to Drinkwater and Crocker [53]. Volumetric fraction of fresh water \( W_w \) could not be measured or estimated adequately; therefore, this parameter was varied in our simulations. Volumetric content of pure ice \( W_{pi} \) is estimated based on measured snow density and volumetric fractions of brine and water.

B. Sea Ice

There are a number of dielectric mixture models for estimating CDC of sea ice. At the same time, there is no a universal model for modeling CDC of sea ice based on a significant statistics of experimental data. For example, the Polder–van-Santen/de Loor (PVD) mixture model described in [54] is often utilized. However, experimental and modeled CDC values according to PVD model do not agree very well [54]. A simple dielectric mixture model for describing CDC of sea ice at 4 GHz is presented in [48]. However, CDC of sea ice depends on frequency. For example, in the recent experimental study [55], it is seen that both the real and imaginary parts of measured sea ice CDC are decreasing with increasing frequency. There is no such a simple linear relationship between sea ice CDC and brine volume at 5.5 GHz. The refractive dielectric mixture model showed a very good agreement with experimental data at 10 GHz as shown in [49]. Therefore, in this paper, we use the refractive dielectric mixture model for estimating CDC of sea ice.

The CDC of sea ice can be estimated using the power mixture model for an isotropic two-phase medium consisting of pure ice and brine inclusions as follows:

\[
\varepsilon_i^a = \varepsilon_{pi}^a (1 - V_b) + \varepsilon_{bi}^a V_b \tag{23}
\]

where \( \varepsilon_{bi} \) is the CDC of brine inclusions in ice, which follows the Debye relaxation model with the temperature-dependent parameters presented in [51], and \( V_b \) is the volume fraction of brine inclusions, which can be estimated through the Frankenstein and Garner empirical equations [54]. Following [13] and [49], we implement the refractive dielectric mixture model with \( \alpha = 0.5 \).

V. RESULTS AND DISCUSSION

A. Environmental Conditions

We observed relatively calm and stable weather conditions between May 20 and May 24. In the afternoon on May 24, a snowfall event occurred, followed by a severe storm overnight. Storm wind speeds exceeding 17 m/s were registered by the meteorological station nearby our experimental site (as shown in Fig. 3). A drizzle precipitation event happened on May 26 between 1 and 4 P.M. (LST), leaving very wet snow behind. The scatterometer system was not operational during these events to prevent damage.

Fig. 4 demonstrates changes in snow temperature at four depths and at the snow surface, as well as changes in sea ice temperature profile (with 5-cm resolution) throughout the experiment captured by our meteorological station. Fig. 4(a) shows that the diurnal temperature variations in snow were quite moderate during the prestorm phase of the experiment. After the storm event, we observed higher air and snow temperatures and higher diurnal variability in air and snow temperatures. Observed changes in snow temperature (and therefore, indirectly, in snow moisture) can be attributed to changes in the energy budget of the snowpack. For instance, the increase in diurnal variability of snow temperature was driven by enhanced variations in short- and long-wave radiation after the storm (as shown in Fig. 5). Given the relatively high wind speed and air temperature on May 26 and 27 (Fig. 3), it is also likely that a high sensible heat flux delivered considerable energy to the snowpack in the period following the storm. Fig. 4(b) shows that the entire sea ice layer is warming over the poststorm phase of the experiment, likely due to the increase in heat flux into the ice from both the snowpack and underlying ocean.

Fig. 6 demonstrates the physical measurements of density, salinity, and temperature in the snowpack. Overall, snow
density did not change appreciably throughout the experiment, although the increased values are observed after the drizzle precipitation event on May 26 [Fig. 6(a)]. Similar to the meteorological station data [shown in Fig. 4(a)], the increased temperature and its variability in snow are observed during the poststorm period [Fig. 6(c)]. Fig. 6(b) shows that the bulk salinity in snow has decreased after the storm event, which is likely caused by fresh snowmelt and fresh water deposited into the system by the drizzle precipitation event (occurred on May 26). Furthermore, our field notes indicate that snow was much moister during the poststorm phase of the experiment compared with the prestorm phase. A snowball could be made; moisture from snow could be pressed out by moderately squeezing the snow in hands. These above-mentioned factors indirectly indicate higher levels of snow moisture content and its variability after the storm. Variability of snow moisture throughout the experiment is discussed as follows (see Fig. 12 and associated discussion in Section V-D).

B. Surface Roughness

Table I demonstrates the rms height and correlation length of both the small- and large-scale components of the air–snow interface extracted from LiDAR scans conducted before the storm (on May 20 and May 24) and after the storm (on May 27 and May 30). One may observe that the rms height of both the small- and large-scale roughness did not change...
Fig. 6. Manually measured physical properties of snow with 2-cm resolution. 0 cm indicates snow–ice interface. Time zero = 00 A.M. of May 20, 2014, LST. 0 cm indicates snow–ice interface. (a) Density. (b) Salinity. (c) Temperature.

TABLE I

<table>
<thead>
<tr>
<th>Date</th>
<th>Small-scale roughness</th>
<th>Large-scale roughness</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>RMS height [cm]</td>
<td>Correlation length [cm]</td>
</tr>
<tr>
<td>May 20</td>
<td>0.13</td>
<td>4.3</td>
</tr>
<tr>
<td>May 24</td>
<td>0.15</td>
<td>4.6</td>
</tr>
<tr>
<td>May 27</td>
<td>0.14</td>
<td>2.6</td>
</tr>
<tr>
<td>May 30</td>
<td>0.13</td>
<td>3.3</td>
</tr>
</tbody>
</table>

significantly over the storm event; however, correlation length of both the small- and large-scale components had considerably decreased.

Fig. 7 demonstrates the large-scale topography of the LiDAR experimental site along with the extracted normalized pdf of 5-cm slopes in a logarithmic scale before (May 24, 2014) and after (May 27, 2014) the storm event. One may observe that the poststorm pdf is wider compared with the prestorm one. We estimated that at the level of $10^{-4}$, the poststorm pdf is on average (i.e., over all azimuth angles) 23% wider than the prestorm pdf. This indicates that the number of occasions with high large-scale slopes of the air–snow interface had become larger after the storm, coinciding with a decrease in the measured correlation length, likely due to new sastrugi formation during the storm. It is seen that after the storm, the rough surface becomes anisotropic (i.e., the correlation length is a function of the azimuth angle). The correlation lengths presented in Table I after the storm are averaged over the azimuth angles. We discuss the potential impact of this anisotropy on the modeling results given in the following. Furthermore, the peak of the pdf after the storm is lower compared with the peak of the prestorm pdf. These results suggest that the air–snow interface had become “rougher” after the storm, which is also consistent with our qualitative visual observations.

Fig. 7(b) and (d) also shows that both pdfs of large-scale slopes before and after the storm are quite narrow. Therefore, accounting for the pdf of large-scale slopes in the small-perturbation theory (following [39]) does not practically affect the NRCS calculated for a horizontally oriented (zero-sloped) small-scale roughness. This observation led us to use the Kirchhoff approximation for modeling scattering from the large-scale snow-surface topography.

C. Modeling NRCS

For each of the 20 physical sampling data sets collected throughout the field experiment, we computed angular dependencies of VV and HH NRCS according to the scattering model presented in Section II. Prior to computing dielectric profiles of snow and ice as described in Section IV, the measured physical characteristics (temperature and salinity of ice and snow, and snow density) were interpolated at 2-mm resolution using the cubic interpolation. Two examples of estimated CDC profiles for snow and ice before and after the storm event are shown in Fig. 8. An increase in both the snow and sea ice CDC is observed after the storm event due to warmer temperatures and increased moisture in snow. An increase of CDC of snow with increasing snow moisture is also shown in [57]. Estimated sizes of snow grains at different depths from our photographs are shown in Table II.
Fig. 8. Examples of estimated CDC profiles of snow and ice using dielectric mixture models before (May 20, 9 P.M.) and after (May 25, 6 P.M.) the storm event. Note the increase in both the snow and ice CDC after the storm.

Table II

<table>
<thead>
<tr>
<th>Layer</th>
<th>0-2 cm</th>
<th>2-4 cm</th>
<th>4-6 cm</th>
<th>6-8 cm</th>
<th>8-10 cm</th>
<th>10-12 cm</th>
<th>&gt;12 cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Average radius [cm]</td>
<td>4</td>
<td>3</td>
<td>2</td>
<td>1</td>
<td>1</td>
<td>0.5</td>
<td>0.05</td>
</tr>
</tbody>
</table>

In our simulations, we varied surface roughness parameters of snow and ice and snow water content as shown in Table III. Roughness statistics of the hidden snow–ice interface could not be measured in the field. However, the FYI in our case was very smooth. Therefore, the range of the roughness parameters for the snow–ice interface could be assumed to be similar to the range of parameters selected for the small-scale roughness of the air–snow interface. For each set of parameters, a cost function between the modeled VV and HH NRCS and corresponding measured ones (averaged over ten scatterometer scans) is calculated. The minimum cost function identifies the set of parameters, which are selected for comparison with measured NRCS data in Section V-D. It is worthwhile to note that quite a wide range of parameters could be tested within our scattering model due to: 1) the model is not computationally expensive and 2) the model represents a superposition of independent scattering components; each scattering component is a function of its own input surface roughness parameters. We also note that our optimization process was chosen to be a noniterative, as such an optimization process ensures that a global minimum of the cost function is detected.

D. Time-Series of Experimental and Model NRCS

Fig. 9 demonstrates the time-series experimental and model HH NRCS at 20° and 45° incidence angles. (b) 30° and 55° incidence angles. RADARSAT-2 measurements at 45.9° on May 25, 2014, and at 30.4° on May 24 are marked by crosses. Vertical bars in RADARSAT-2 data indicate the standard deviation of NRCS within a 11 × 11 pixel area (at 50-m resolution) surrounding the experimental site. Time zero = 00 A.M. of May 20, 2014, LST.
time-series NRCS is observed. Our statistical analysis showed that, before the storm, $R^2$ between measured and modeled HH NRCS for all incidence angles was 0.82; after the storm, $R^2$ slightly decreased to 0.78. HH RADARSAT-2 measurement taken on May 24 (before the storm) agrees with corresponding scatterometer values considering the observed variability of RADARSAT-2 signal in the vicinity of our experimental site. HH RADARSAT-2 measurement taken on May 25 (during the storm) reasonably lies between NRCS values obtained before and after the storm event for both the experimental and model time-series data. We also observed good agreement between experimental and model VV NRCS during the prestorm and poststorm phases of the experiment (as shown in Fig. 10). $R^2$ between measured and modeled VV NRCS was 0.88 before the storm and 0.81 after the storm. Slightly lower agreement between the model and experimental NRCS after the storm can be associated with the anisotropic nature of the poststorm large-scale roughness, while in our measurement technique and modeling, we assume an isotropic case.

Experimental time-series HV NRCS along with available HV RADARSAT-2 measurements is shown in Fig. 11. Note that the model HV signal is not available as it is not accounted by our scattering theory. Both RADARSAT-2 HV measurements are substantially higher than the scatterometer values. We attribute it to the fact that the scatterometer HV values are below the noise-equivalent sigma zero (NESZ) of ScanSAR RADARSAT-2 mode (as shown in Fig. 11). Also, both available HV RADARSAT-2 values (at 30.4° on May 24 and at 45.5° on May 25) are almost the same. This indicates that, in our case, RADARSAT-2 HV NRCS is not sensitive to the incidence angle and the changes in snow-covered sea ice, due to the high NESZ in ScanSAR mode. For comparison, the scatterometer NESZ provided by the manufacturer (ProSensing) based on their calibration experiment is −40 dB.

During the prestorm phase of our experiment, the NRCS response was relatively stable. On May 24, HH, VV, and HV NRCS decreased, likely due to the increased amount of moisture in the snowpack caused by high air temperatures (as shown in Fig. 12). After the storm event, substantially higher values and variations of HH, VV, and HV NRCS are observed. We attribute this effect to the following factors: 1) higher values and variability of moisture in the snowpack driven by higher and more variable snowpack temperatures, in addition to the short drizzle event on May 26 and 2) an overall increase in surface roughness at the snow–air interface after the storm. Fig. 12 shows snow moisture over the period of the experiment estimated from the optimization process. During the prestorm phase of the experiment, one may observe relatively low values, following the diurnal cycle of the air and air–snow interface. After the storm and the drizzle event, high air and air–snow temperatures and their high variability led to the increase in the snow moisture
content and its variability. From time-series plots of NRCS (Figs. 9–11), it is seen that after the drizzle event on May 26, the NRCS at all polarizations increased earlier at low incidence angles and later at higher incidence angles. This time lag could be associated with the fact that during the water transport within the snowpack, the contributions of different scattering mechanisms evolved differently at different incidence angles. Our analysis of contributions from different scattering mechanisms for May 26 revealed that at low incidence angle, a larger increase in both the small- and large-scale surface scattering is observed as the snow moisture content increased compared with high incidence angles, whereas the volume scattering component was low at all incidence angles.

Fig. 13 shows the sensitivity of HH and VV NRCS to the snow moisture as the functions of the incidence angle and the snow moisture calculated for physical data collected at May 21, 9 P.M. The sensitivity is calculated as a partial derivative of HH and VV NRCS with respect to the snow moisture parameter. It is seen that at low snow moisture values (at or below than $\sim 1.5\%$), the sensitivity is negative, meaning that NRCS decreases with increasing snow moisture, which is particularly pronounced at high incidence angles. This effect is explained by the fact that the volume scattering is suppressed by relatively small amounts of moisture in snow, but the surface scattering did not increase enough to increase the total NRCS. Such a decrease in NRCS associated with an increase in snow moisture (associated with high air temperatures) is observed on May 24, just before the storm event. Fig. 13 also shows that at higher snow moisture content (higher than $\sim 1.5\%$), the sensitivity is positive, meaning that the total NRCS tends to increase with increasing snow moisture. Such a situation is observed after the storm event, when the surface scattering becomes dominant, mainly due to high levels of snow moisture leading to a high dielectric contrast between the air and snow. From Fig. 13, it is also seen that at high values of snow moisture, NRCS becomes less sensitive to snow moisture; therefore, three snow moisture values after the storm estimated to be higher than 13% (Fig. 12) could be slightly overestimated.

Warmer temperatures and increased moisture content in snow followed by the storm also led to decreasing salinity of the snow basal layer (0–2 cm) from 9.7±1.2 ppt (average value and standard deviation over 12 sampling sessions conducted before the storm) to 7.7±0.9 ppt (over 8 sampling sessions after the storm). Bulk salinity of the upper ice layer (0–5 cm) had also decreased from 5.1±0.5 ppt (before the storm) to 4.1±1.2 ppt (after the storm), likely due to percolation of fresh snowmelt into the upper portion of the sea ice. However, these changes did not noticeably modify the scattering contribution from the snow–ice interface as shown in Section V-E and the corresponding figure Fig. 16.

Table IV demonstrates average optimum parameters minimizing the cost function between the model and experimental data before and after the storm. Most importantly, the snow moisture content and its variability substantially increased between the two periods (as shown in Fig. 12). This is in agreement with observed changes in the snowpack energy budget over the experiment (Fig. 5). Higher air temperatures, variable radiative fluxes, and likely also an enhanced turbulent heat flux, during the poststorm period, meant that more energy was available within the snow for diurnal melting and subsequent refreezing.

It is noticeable that relative changes in both small- and large-scale optimal roughness parameters at the air–snow interface are consistent with our LiDAR observations (Table I). For instance, the rms heights of both small- and large-scale roughness did not change considerably, whereas the correlation lengths of both roughness components decreased. The percentage change in the average optimal small-scale correlation length was $-10\%$ compared with an observed change of $-34\%$ and the same for the optimal large-scale correlation length was $-10\%$ versus an observed change of $-13\%$. The optimal correlation length and rms height of the snow-ice interface decreased. Absolute values for the optimal rms height and correlation length compare less well with our LiDAR observations. One possible reason for

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Before storm</th>
<th>After storm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow moisture [%]</td>
<td>0.62</td>
<td>12.03</td>
</tr>
<tr>
<td>Small-scale snow roughness RMS height [cm]</td>
<td>0.29</td>
<td>0.3</td>
</tr>
<tr>
<td>Small-scale snow roughness correlation length [cm]</td>
<td>1.98</td>
<td>1.79</td>
</tr>
<tr>
<td>Large-scale snow roughness RMS height [cm]</td>
<td>5.3</td>
<td>5.6</td>
</tr>
<tr>
<td>Large-scale snow roughness correlation length [cm]</td>
<td>385.8</td>
<td>347.5</td>
</tr>
<tr>
<td>Ice roughness RMS height [cm]</td>
<td>0.29</td>
<td>0.15</td>
</tr>
<tr>
<td>Ice roughness correlation length [cm]</td>
<td>2.2</td>
<td>1.71</td>
</tr>
</tbody>
</table>
Fig. 14. Model and experimental angular dependencies of HH NRCS. (a) and (b) Two cases before the storm. (c) and (d) Two cases after the storm. Error bars denote one standard deviation based on ten consecutive scans. Volume and snow-ice interface scattering components are not shown in (c) and (d) as they are below than $-50$ dB.

This minor inconsistency is that both measured roughness parameters and the model assume an isotropic correlation function, which our observations demonstrate was not the case, particularly after the storm (Fig. 7). Although our scatterometer measurements are averaged over $60^\circ$ arc in the azimuth direction, scanning over $360^\circ$ would potentially better compensate for the anisotropy of the rough snow surface.

These results suggest that a general increase in roughness of the snow cover system caused by the storm event, and changes in snow moisture caused by increased poststorm air temperatures and the precipitation event on May 26, caused an overall increase in C-band radar backscatter and also significantly affected diurnal variations in scattering.

E. Scattering Mechanisms During Prestorm and Poststorm Periods

Four examples of measured and simulated HH and VV NRCS as the functions of the incidence angle are shown in Fig. 14 (HH) and Fig. 15 (VV). Figs. 14(a) and (b) and 15(a) and (b) represent two cases before the storm, while Figs. 14(c) and (d) and 15(c) and (d) represent two cases after the storm. All four scattering components from (1) are plotted for each case. Good agreement between model and experimental angular dependencies is found for all four cases at both HH and VV polarizations. One may observe that the volume scattering component is significantly higher before the storm compared with the poststorm period. This component does not strongly depend on the incidence angle, and mainly contributes to the total backscatter at high incidence angles. The nature of volume scattering from the basal layer of snow and its dependence on temperature, snow thickness, and brine content is documented in [26]. Note that the volume and snow-ice interface scattering components are not shown in Figs. 14(c), 14(d), 15(c), and 15(d) as they are below than $-50$ dB. It is also seen that scattering from the large-scale component of the air–snow interface becomes more significant after the storm. We also note that HH NRCS from the large-scale roughness is greater than corresponding VV NRCS; this effect is particularly pronounced at higher incidence angles. Larger HH NRCS values compared with VV NRCS represent a property of the Kirchhoff model. The opposite situation is observed for HH and VV NRCS from the small-scale roughness.

To further investigate scattering mechanisms during the prestorm and poststorm phases of our experiment, we calculated average relative power contributions HH + VV (and standard deviations) of each of the four model scattering components before and after the storm as the functions of incidence angle. These results are presented in Fig. 16.

During the prestorm phase of our experiment, the large-scale scattering component is dominant at low incidence
angles and becomes less significant at high incidence angles [Fig. 16(a)]. The volume scattering component is significant at high incidence angles and becomes less significant at low incidence angles [as shown in Fig. 16(b)]. Contribution from the small-scale roughness component of the air-snow interface is moderate at all incidence angles. Snow-ice scattering component is not very significant at all incidence angles.

During the poststorm period, the contribution from the large-scale component of the air–snow interface becomes significantly higher compared with the prestorm period at all incidence angles. Volume scattering does not play a significant role after the storm at all incidence angles due to the increased amount of moisture in snow. However, the significance of the small-scale air-snow scattering becomes higher, particularly at high incidence angles [as shown in Fig. 16(c)]. Contribution from the snow–ice interface remains low after the storm event [Fig. 16(d)]. These results can be associated with the increase in large-scale roughness and the increase in the water content in the snowpack (after storm), which could suppress the intensity of volume scattering primarily coming from the basal layer of snow. The masking effect of wet snow and the importance of the air–snow interface when the snow is wet could be found in [58].

VI. CONCLUSION

This paper presents the measurement and modeling results of time-series angular dependencies of C-band NRCS over the FY snow-covered sea ice during the winter–spring transition period. The scatterometer data were collected from a stationary platform near Cambridge Bay, Nunavut, Canada, over the period of time between May 20 and May 28, 2014, covering a storm event on May 25. The prestorm period is characterized by stable, relatively cold environmental conditions, while the poststorm phase of the experiment was accompanied by large
variability of environmental parameters (such as air and snow temperatures).

Throughout the experiment, 20 physical sampling sessions of snow and ice were conducted to capture temperature and salinity profiles of snow and ice, density of snow, as well as sizes of snow grains. Our LiDAR measurements of surface roughness in the vicinity of the experimental site revealed that two scales of roughness at the air–snow interface are present: small scale (~0.001–0.01 m) and large scale (~1–10 m). To simulate changes in NRCS, we proposed a model integrating the following scattering mechanisms: 1) surface scattering from the small-scale roughness component of the air–snow interface; 2) surface scattering from the large-scale roughness component of the air–snow interface; 3) surface scattering from small-scale roughness of the snow–ice interface; and 4) volume scattering from snow grains (with different sizes at different depths) within the snowpack. Collected physical data were used to calculate dielectric profiles of both snow and ice. Snow moisture content as well as snow and ice roughness parameters varied in our simulations.

We observed good agreement between measurement and model NRCS during both the prestorm and poststorm phases of the experiment. Before the storm, $R^2$ between model and experimental NRCS was 0.88 and 0.82 for VV and HH, respectively. After the storm, $R^2$ was slightly lower 0.81 and 0.78 for VV and HH, respectively. Available RADARSAT-2 measurements also reasonably agree with our scatterometer data. Our model results showed that the water content in snow and its variability significantly increased after the storm. This is in agreement with our field measurements of other relevant parameters (e.g., snow and air temperatures, and surface energy fluxes) and qualitative observations. Furthermore, the obtained model results demonstrate that the correlation length of both the small- and large-scale roughness components of the air–snow interface decreased while the rms height of both the small- and large-scale roughness components did not change significantly after the storm. Therefore, our model results suggest that the overall roughness of the snow–ice interface increased after the storm, in agreement with LiDAR observations collected in the vicinity of the scatterometer site.

Our model analysis of scattering mechanisms revealed that large-scale and small-scale surface scattering from the air–snow interface as well as volume scattering are the dominant scattering mechanisms during the prestorm period. The significance of the large-scale scattering decreases with increasing incidence angle, so that the contribution from volume scattering increases with incidence angle. After the storm, the volume scattering contribution significantly dropped, while the large-scale surface scattering contribution increased. The contribution of small-scale roughness after the storm becomes more significant at high incidence angles. We attribute these effects to increase in the air–snow surface roughness (due to the storm) along with, importantly, an overall increase in snow moisture. These effects led to decreasing volume scattering and increasing surface scattering at the air–snow interface after the storm.

A significant increase in water content in the snowpack after the storm suppressed volume scattering from the basal layer of snow and also enhanced the role of surface scattering from both the small- and large-scale components of the air–snow interface. Short-term variations of the snow moisture led to significant variability of the NRCS after the storm at all incidence angles and polarizations.

Presented results on modeling and observation of time-series changes in C-band NRCS with respect to physical properties of snow-covered sea ice provide in-depth understanding of scattering processes under changing environmental conditions during the winter–spring transition period. With the increased number of C-band SAR observations over the Arctic region, it becomes feasible to obtain and analyze time-series changes of radar backscatter from sea ice at high spatial resolution. The results presented in this paper could aid in interpretation of time-series SAR observations with respect to physical properties of snow and ice during the winter–spring transition period in the Canadian Arctic Archipelago. This paper is particularly important in light of the upcoming Canadian RADARSAT Constellation mission that will dramatically increase the temporal resolution of C-band SAR data over the Arctic Ocean. For the future research, it would be interesting to investigate seasonal continuous time-series changes in radar backscatter and scattering mechanisms and corresponding physical properties (including roughness) of snow-covered Arctic sea ice. The sensitivity of active microwave remote sensing to the subtle thermophysical changes in snow-covered sea ice is an important evolving tool for a variety of thermally related changes to snow-covered sea ice and related consequences on physical, biological, and geochemical fluxes across the ocean–sea ice–atmosphere interface.

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Microwave remote sensing, sea ice and Arctic climate


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