

Interpreting the Altimetric Radar Cross Section of Young Sea Ice

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Abstract—Ku-band radar altimeters such as the one onboard CryoSat-2 measure the return-time of backscattered radar pulses to derive sea ice freeboard and thickness. This method relies on the two-way penetration of radar waves through the overlying snow. Early research with a surface-based radar over tank-grown sea ice indicated that most radar power is returned from the sea ice surface rather than the overlying snow, but this has not been replicated by surface-based radars in the field. We present a 24-day timeseries of surface-based radar measurements from young Antarctic sea ice (growing from 20 to 36 cm in thickness), showing a progressive reduction in its normalised radar cross-section (σ^0) over time. This is associated with weakening of very strong, single-peaked waveforms associated with specular reflection from the ice surface; these initially swamp the signal from the snow surface but rapidly diminish with snow and ice ageing. Our observations help resolve the apparent contradiction between previous surface-based studies, and are consistent with an early-winter decline in σ^0 observed by CryoSat-2. Our results also highlight the suppressing effect of frost flowers on σ^0 , and the similarity of σ^0 evolution between the Ku- and Ka-bands. Our findings are relevant to the capability of the planned Ku/Ka-band CRISTAL altimetry mission, which aims to leverage differential radar penetration through snow on sea ice.

Index Terms—Snow, altimetry, cryosphere, radar, sea ice

I. INTRODUCTION

Satellite radar altimetry missions operating in the Ku-band such as Envisat, Sentinel-3 and CryoSat-2 derive sea ice freeboard by measuring the return-time of radar pulses that are assumed to backscatter from the sea ice surface. These freeboard values can then be converted to estimates of sea ice thickness, which is a WMO-defined *essential climate variable* [1]. However, there is a lack of consensus in the scientific literature about the exact location where the Ku-band radar backscatter originates from sea ice (the dominant scattering surface). In particular, a 1992 experiment in a tank facility by Beaven et al. [2] using a surface-based Ku-band radar indicated that - despite a ~ 10 dB reduction in total backscattered power - very little power was backscattered from the upper surface of snow sifted through a screen onto a 12 cm sea ice sheet.

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This provided key support for the design and launch of the CryoSat-2 satellite mission in 2010 [3]. The assumption that the snow surface is not represented in the radar return now underpins most of the publicly available sea ice thickness data products derived from CryoSat-2 data (e.g. [4], [5], [6], [7], [8]). However, subsequent studies with Ku-band surface-based radars on thicker sea ice in both the Southern Ocean [9], [10] and the Arctic Ocean [11] have indicated that a significant amount of power is returned from the snowpack overlying the sea ice. To date, it has been difficult to reconcile these disparate findings.

To complicate the issue, the fraction of backscattered radar power from the sea ice surface (relative to the snow surface) is sensitive to the coherence of the radar return. This is partly a function of target properties such as roughness and Fresnel coefficients, but also of the geometry of the radar system (see [12] for investigation over relatively homogeneous first-year ice). This means that surface-based radar studies of the dominant scattering surface conducted using sled-mounted radars such as those listed above, are not directly applicable to satellite length-scales and beam-geometries. Specifically, the size of satellite altimeter footprints on the ice surface are much larger than those of surface-based radars, even though the angular beam-widths are much narrower.

Despite the key geometrical differences between surface- and satellite-based measurements described above, several airborne studies have suggested that (contrary to the findings of [2]), the snow surface or volume can produce a meaningful radar return at nadir-view. For instance, results from the CryoSat Validation Experiment (CryoVEx) using the Ku-band *Airborne Synthetic Aperture and Interferometric Radar Altimeter System* (ASIRAS) in the Bay of Bothnia suggest that the dominant scattering surface was frequently well above the snow-ice interface [13], with the degree of this effect being associated with air temperature. [14] also used ASIRAS in the Western Arctic as part of CryoVEx, finding an even closer relationship between the dominant radar return and the snow surface in early May with the use of an airborne laser-scanner. [15] combined ASIRAS and CryoSat-2 data with an airborne laser scanner and also found Ku-band freeboards were more closely associated with the laser-derived snow-air interface heights than the snow-ice interface height. [16] analysed airborne Ku-band echograms over sea ice in both hemispheres, and clearly resolved the snow surface, noting the main peak in returned power "meandered" between the air-snow and snow-ice interfaces. Finally, [17] found that the maximum power of Ku-band waveforms occurred within 0.1 m

of the air/snow interface in 33% of waveforms over Antarctic sea ice in the Weddell Sea, although these results come from summer conditions where the potential for wet snow limits direct comparison to [2].

As well as Ku-band investigations of the return produced by the snow surface over sea ice, there have also been some over terrestrial and glacial snow. [18] document results from an airborne Ku-band radar over Byrd Glacier and found a relatively strong return from the air-snow interface (see their Figure 5). In the field of satellite Ku-band altimetry of ice sheet surface elevation, it is generally accepted that radar returns from the snow surface can produce a bias in ranging, and this is operationally mitigated on a seasonal timescale with a correction based on the normalised radar cross-section (σ^0) [19], [20].

Several field campaigns have also documented a return from the snow surface over sea ice at lower frequencies, that should in theory be more ‘penetrating’. For instance, [21] used a C-band surface based radar at nadir-view and found that the dominant scattering horizon was more often closer to the air-snow interface than the snow-ice interface. [22] used a helicopter-mounted radar sweeping 2-8 GHz to resolve the snow and sea ice surfaces and measure snow depth over Antarctic sea ice. [23] took a similar radar that swept from 2 - 8 GHz on a sled to sea ice in East Antarctica, finding again that the snow surface produced a radar return of comparable magnitude to the snow-ice interface. Perhaps most interestingly, a follow-up experiment to the 1992 experiment (described by [2]) was conducted at the CRREL tank in 1994 (described by [24] and [25]). However there were three key differences in the experimental setup: the radar’s frequency sweep was lower (2 - 18 GHz, putting the central frequency in the X-band); the snow accumulated naturally over a longer period, and the ice was older and thicker at around 33 cm. A power peak from the snow surface was clearly visible with a power of around one-sixth that of the snow-ice interface.

As well as surface-based and airborne evidence for a meaningful radar return from the snow surface in the Ku-band, there is also satellite evidence for the phenomenon. [26] examined snow accumulation near drifting buoys in the Arctic during Autumn 2013, and found that CryoSat-2 freeboard estimates were positively correlated with the snow freeboard. [27] applied this approach more widely, finding that snow accumulation on a pan-Arctic scale was frequently associated with a rise in CryoSat-2 derived radar freeboards, suggesting a non-negligible return from the snow surface. [28] then investigated the results of accounting for such a return in CryoSat-2 sea ice processing, finding that the representation of interannual variability in the Beaufort Sea and Fram Strait was marginally improved.

The literature described above illustrates the degree of discrepancy between [2] and subsequent studies on multiple scales. In this paper we investigate this discrepancy, focussing on the very young and thin nature of the ice studied at the CRREL sea ice tank in 1992. We do this by presenting a series of Ku- and Ka-band surface-based radar measurements of young, smooth sea ice near the Antarctic Peninsula. The timeseries was produced over a 24-day period, during which

the ice thickened from 20 to 36 cm and the ice surface transformed from a covering of frost-flowers to a 6 cm snowpack. Our in-situ results indicate a decline in the sea ice σ^0 by approximately a factor of four relative to the first day on which it was measured with a snow cover. This is associated with substantial weakening over time of specular, single-peaked waveforms similar to those observed by [2], while double-peaked waveforms representing the snow and ice surfaces remain roughly consistent in amplitude. We then show that our results are consistent with the evolution of σ^0 as measured by CryoSat-2. We go on to discuss what this might mean for the concept of ‘radar penetration’ over such young ice.

Our use of the term ‘young sea ice’ above is informed by the World Meteorological Organisation’s *Sea Ice Nomenclature* [29]. This defines young ice as ‘ice in the transition stage between nilas and first-year ice, 10-30 cm in thickness’. As such, the sea ice is not strictly ‘young’ towards the end of our in-situ study, having technically thickened to what the WMO terms ‘first-year ice’.

II. FIELD SITE, WEATHER CONDITIONS, AND ICE EVOLUTION

Our field site was located in Hangar Cove, to the North of Rothera Research Station on Adelaide Island, Antarctica (67.564°S, 68.128513°W). The cove underwent intermittent cycles of freezing up and then blowing clear of ice over the course of Austral Winter 2023. The cove was blown clear of ice by a southerly wind on the 13th of August, before freezing over into large nilas sheets on the 14th under quiescent conditions (Supplementary Figure S1. The ice then rapidly grew over a two-week period to a thickness of 19 cm (measured at several points along the transect) by the 29th of August - this was the first time that this floe’s thickness was measured. During this period, the average air temperature was -10.5°C (typical for the time of year). On the night of the 28th August, the ice developed a widespread covering of frost flowers: saline ice crystals that form when relatively warm, young ice is exposed to air temperatures below -15 °C [30], [31]. The frost flowers were flattened slightly by a light wind before our first radar measurements on the 30th (Figure 2a & b). Sea ice thickness was initially measured with 2” diameter auger-holes and a tape-measure, with 9 cm diameter sea ice cores confirming these measurements from the 4th of September onwards.

Shortly after our first measurements (30th August) there was a light snowfall which resulted in 3 cm of low-density snow cover. The density was so low that it was not possible to accurately take a volume-controlled sample with our 3 cm high box-shaped density cutter. This was then scoured and metamorphosed by high winds on the evening of the 31st to a 1 cm thick cover of denser snow. A blocking high-pressure weather system then led to a period of relative geophysical stasis, with the snow cover only thickening to 3 cm again by the 20th of September. Snowfall on the 22nd then led to a doubling of the snow depth (from 3 cm to 6 cm), prior to our final measurements on the 23rd of September. Snow depths

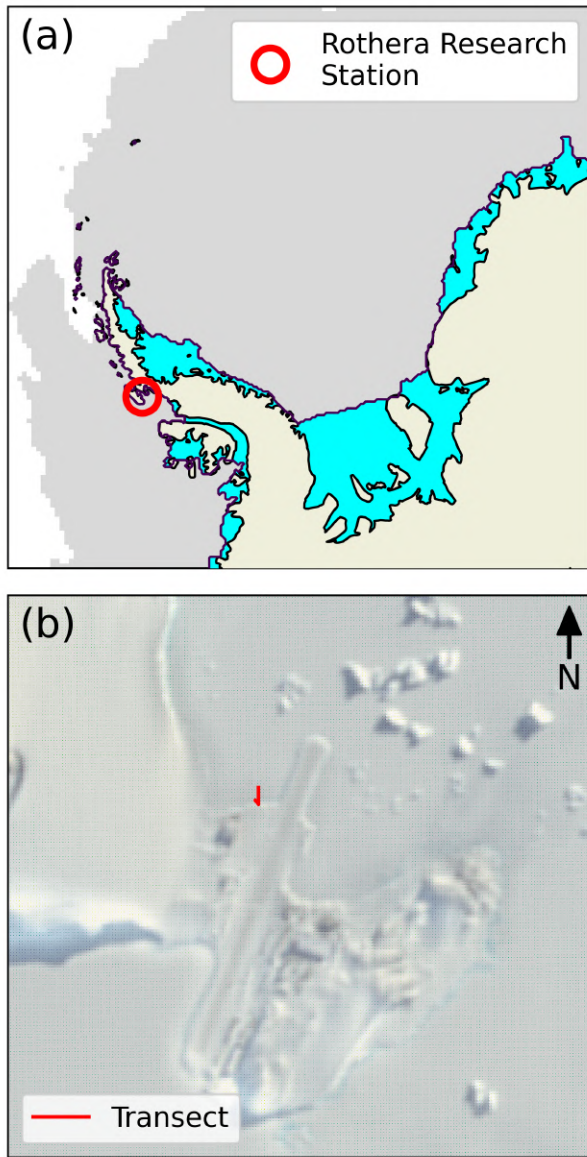


Fig. 1. (a) Location of Rothera research station in West Antarctica, with sea ice concentration on the last day of our study (2023/9/23) of over 50% marked in grey, and regions of floating ice shelf in light blue. Sea ice concentration data taken from [32] (b) location of the radar transect in Hangar Cove to the north of Rothera Research Station shown on enhanced colour Sentinel-2 image.

were measured with a ruler along the top of a snow pit near the centre of the transect.

This low rate of snow accumulation (6 cm between freeze-up on the 14th of August and our final measurement on the 23rd September) coincided with the average sea level pressure over the same period being 1.5 standard deviations above average for that time of year, producing relatively stable weather conditions. The mean 2 m air temperature during our study period (first to last measurement; 30th August - 23rd September) was -10.2°C. The station experienced a low-pressure weather system after our last transect measurement on the 23rd of September, with several weather fronts causing snow dumps and accumulation in excess of 20 cm. Unfor-

tunately, a necessary snow clearing operation on the nearby aircraft runway led to snow being plowed from the land onto our field site, covering much of the transect, thus ending this study.

III. RADAR MEASUREMENTS

A. Transect Measurements

Measurements were collected using the KuKa radar system, operating in stare mode following [33]. We calculate σ^0 by integrating the power over all range-bins between 1.3 and 3 m in radar range; noting that the snow surface generally appears at a range of around 1.5 m. We find that changes to these integration limits do not affect our results, since the returned power outside the limits is negligible.

Because of its relevance to satellite altimetry missions, our investigation focuses only on the co-polarised channels of the KuKa radar system. The KuKa radar system is fully polarimetric, so has two copolarised channels that are analogous to satellite radar altimeters: one with the antennas running along the transect ('HH') and another with antennas running across the transect ('VV'). We find that these two channels produce identical results for the purposes of the analysis in this manuscript - we therefore only display results from the VV channel in the interest of clarity.

Due to safety concerns involving the thickness of the ice and the strength of its attachment to shore, it was initially not possible to accompany the radar onto the sea ice. Instead, a pulley system was set up on the ice such that the radar could be dragged on a sled over a ~ 20 m transect (Figure 2b) while all personnel remained on shore. On its outbound journey the radar could only be dragged towards the pulley; on the return journey towards the shore, we were able to control the path of the radar as we pulled towards our position on the shore (Figure 2c). As such, all measurements presented here correspond to the return (shorebound) journey, where we could ensure that the radar was not sampling snow that had previously been disturbed by the sled. For early transects where the radar was unaccompanied, we mounted a camera on the Ka-band arm of the radar system; this indicated the radar was moving and also what the target was at any given time. A folding ruler was attached to the sled in view of the camera, giving scale to this visual analysis.

B. Frost Flower Measurements

Our first transect occurred prior to snow accumulation, when the ice was covered with frost flowers (Figure 2a). Two samples of these, taken from approximately midway along the transect, indicated salinities of 41.0 & 41.3 ppt. To investigate the role of the frost flowers on radar backscatter, we used a foam-edged squeegee to clear the frost flowers along a strip of ice of around 1 x 3 m at the near end of the transect (Figure 2b). The frost flowers were extremely fragile and so were easily removed, with minimal disturbance to the ice below. The transect was approximately 120 cm wide, and so accommodated the 3 dB footprints of both radar systems. After presenting the results of this investigation, we exclude the squeegeed area from later analysis to avoid biasing our results.

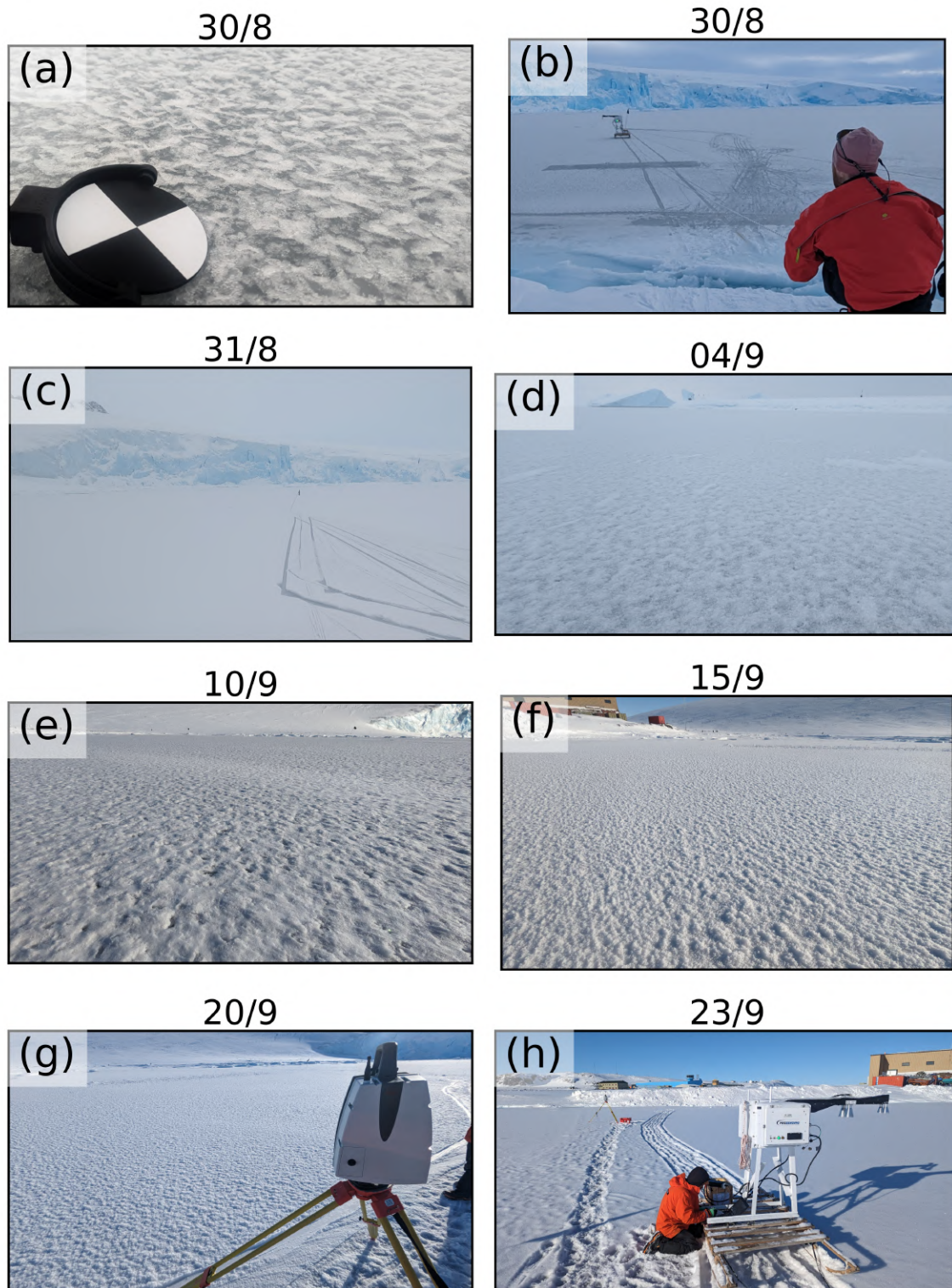


Fig. 2. Photos illustrating the geophysical evolution of the sea ice & snow surface during our study. (a) The frost flowers with a 9 cm diameter disc for scale. (b) The transect, featuring the strip cleared of frost flowers. The distance between the strip and the black flag marks the ~ 21 m transect analysed subsequently. (c) Snowfall on the night of 30th August obscured the cleared strip. Sled tracks illustrate the “outbound” route (right of photo, not analysed) and the “return route” over which we had directional control of the sled. (d) Wind-scouring reduced the snow to a thin covering. (e) a ‘holey’ pattern developed on the ageing, thin snow, where the ice could be seen through small gaps. (f) A rougher small-scale texture develops after light precipitation and wind-driven redistribution, with the ice no longer visible through the holes. (g) Minimal geophysical evolution between 15th & 20th (h) Snowfall leads to a smooth-surfaced, 6 cm deep snowpack.

C. Movement Detection and Duplicate Waveforms

To produce an accurate overall characterization of our transect, we identified and removed data acquired while the sled was stationary. This was done prior to calculating σ^0 values, since including these data would bias the average towards just a few points on the transect where the radar was stationary sometimes for long periods. When the radar is stationary, the radar waveforms (and σ^0 values) are nearly identical (see below for quantitative details). This is in stark contrast to the level of variability in waveform shape and power experienced when the radar is moving over even an apparently homogenous surface. We first removed stationary data by using coordinates from the on-board GPS. However, there is the possibility for false-negatives and false-positives in this approach. False-negatives can occur when a slow, random drift in the GPS signal implies the radar is moving when it isn't. False-positives for stationarity occur when the radar is moved so slightly that it is not detected by the sensitivity of the GPS unit.

To investigate the sources of error described above, we implemented a GPS-independent algorithm to identify waveforms obtained while stationary in the following way: each waveform represents a series of paired power and range values, and all relevant waveforms have a successor waveform taken immediately afterwards. A waveform can be correlated with its successor to produce a Pearson's correlation coefficient, and these coefficients are generally close to unity for stationary waveforms and detectably less for those taken while moving. By analysing the data from the on-board camera in periods when it was mounted, we were able to precisely identify when the sled was moving. The camera recorded at thirty frames per second, which is ten-times that of the Ku-band chirp rate and six times that of the Ka-band rate, allowing each waveform to be easily classified as being taken while the system was moving or while it was stationary.

We find that we can mask out waveforms taken while stationary by identifying waveforms with a Pearson-R correlation score in excess of $> 1 - 1 \times 10^{-5}$. We find that as the exponential term is increased below -10, noise in the system means that inter-waveform variability even while static leads to a false negative for stationarity, where the radar is determined to be moving when it isn't. For an exponential term above -3, the converse is true: waveforms recorded while moving (per the camera) are incorrectly identified as being so similar as to imply the sled is stationary - this is a false negative for stationarity. Given this large range of values which produce acceptable performance (exponential terms between -3 and -10), and noting that the number of waveforms is nearly completely insensitive to variation in the exponential between -3 and -7, we opt for a middle value and use a threshold on the Pearson-R coefficient of $1 - 10^{-5}$. A visual representation of the algorithm is presented in Supplementary Figure S2. We find that false positives and false negatives from the GPS-based method were very infrequent in our study, representing $< 1\%$ of transect waveforms.

D. Range Resolution and Sidelobes

A persistent sidelobe appears in the radar's co-polarised impulse response function around 5 cm in radar range beyond the main peak. We characterised this sidelobe by recording the radar's impulse response function using a 121 x 81 cm metal plate on three dates: before (24th Aug), during (14th Sep) and after (8th Oct) our 24-day timeseries. Further information on some of the drivers of this can be found in [34]. We display the corresponding waveforms in Supplementary Figures S3 & S6, where the sidelobe persistently exists at around 3 cm beyond the main peak's maximum power. To address this, we implement the deconvolution routine described for this radar by [34], using a deconvolution waveforms produced from the metal plate scan on the 14th of September (two-thirds of the way through the experiment). This suppresses but sometimes does not completely remove the sidelobe trailing the main peak (Figure 3 and S4). We use the deconvolved data for the peak-picking and waveform analysis, but the undeconvolved waveforms for the σ^0 analysis. This is because the radar's σ^0 retrieval routine (following [33]) is calibrated against a target of known σ^0 without deconvolution, and so we are only in a position to present σ^0 without deconvolution. Indeed it is not clear to what extent the deconvolution process should conserve σ^0 - such a question is outside the scope of this manuscript. However, it is clear that deconvolved waveforms contain less power than undeconvolved waveforms for the snow-covered transects of interest, but the power that they do contain is very highly correlated with the undeconvolved waveforms. To illustrate this, our timeseries of waveform powers produced with and without deconvolution has a Pearson- R^2 correlation coefficient of 0.986 (Supplementary Figure S5), so the findings in our manuscript would be no different were we to examine the range-integrated power of the deconvolved data instead of σ^0 .

The theoretical range resolution of our radar in the Ku-band is 2.5 cm in free space [33], but the spectral windowing applied also degrades range resolution to 4.1 cm in free space (for case of Hann window). Due to the reduced propagation speed of radar waves in snow, this value is improved (i.e., reduced) by around 25% [11], [35] to approximately 3 cm. The limited range resolution means we cannot always identify waveform peaks corresponding to the air/snow and snow/ice interfaces following the approach of [36], even when power is likely being backscattered from both interfaces.

IV. SATELLITE σ^0 ANALYSIS

As discussed in Section I, a direct quantitative comparison between surface-based radar observations from our instrument and those of CryoSat-2 cannot be assumed due to differences in beam geometry, footprint size of both instruments and length-scale. Specifically, the pulse-limited, Doppler-sharpened footprint of CryoSat-2 is around 0.5 km² whereas the Ku-band 6 dB footprint of KuKa is around 0.18 m² at a range of 1.6 m; this is a difference in footprint area of 2.8×10^6 . The equivalent beamwidth of CryoSat-2 based on its pulse-limited footprint diameter of 0.12°.

We therefore use σ^0 data from CryoSat-2 to show that our surface-based observations appear to be at least consistent with

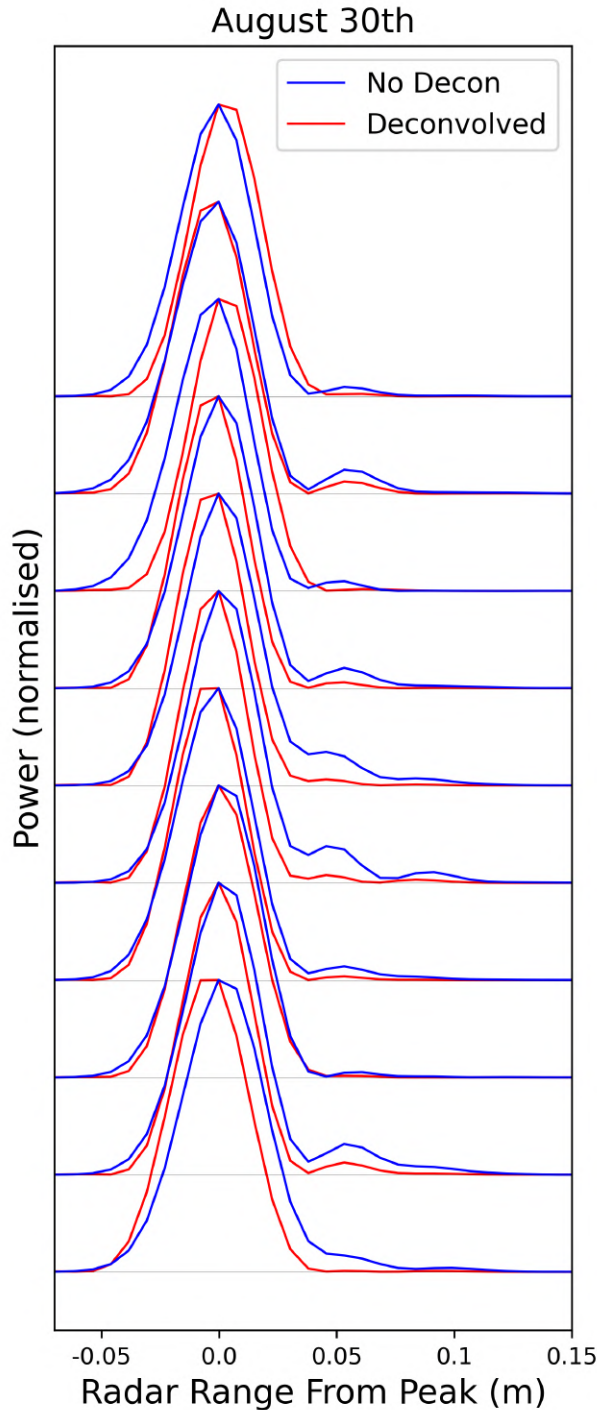


Fig. 3. Ten selected waveforms produced from the transect over frost flowers on August 30th with (red) and without (blue) application of the deconvolution procedure. Peaks are aligned by their range bin of max power, and normalised by amplitude. The sidelobe appearing ~ 5 cm beyond the main peak is suppressed by the deconvolution, but rarely removed. Waveforms not visibly exhibiting the sidelobe have not been displayed. Another version of this plot on a dB scale is displayed in Supplementary Figure S7.

measurements at the satellite's scale and geometry. It is not possible to perform a local analysis of satellite-derived σ^0 evolution to compare to our surface-based results. This is due to contamination effects from the nearby land and icebergs, but also the sparse nature of satellite altimeter data coverage. Instead, we analyse the large-scale seasonal evolution of σ^0 from CryoSat-2 in both hemispheres - this variable is provided with the Baseline-D data product [37] which is produced by the European Space Agency (<ftp://science-pds.cryosat.esa.int>). For completeness, we performed exactly the same analysis with the Saral/AltiKa geophysical data record (version F, downloaded from Aviso; <ftp-access.aviso.altimetry.fr>).

We discard σ^0 data from waveforms that may represent full or partial backscatter from a lead (open water between sea ice floes). To do this, we follow the process and definitions of [6] and impose a pulse-peakiness threshold of 9 and a stack-standard deviation threshold of 6.29 for CryoSat-2 SAR mode waveforms. These values were recently evaluated by [38] and found to work well in the Canadian Arctic. For the AltiKa mission, we apply the thresholds of [39] (pulse peakiness < 5 ; waveform leading edge width < 2 range bins).

We decomposed the data by sea ice type (discriminating first-year ice from multi-year ice) using the OSI-403-d product from the Ocean and Sea Ice Satellite Application Facility [40]. We then applied an additional filter in the Southern Ocean using the product of Melsheimer et al. [41] (M23). M23 does not discriminate effectively between open ocean and FYI, so it was necessary to combine it with the OSI SAF product rather than simply replace the OSI SAF product in the southern hemisphere. Pixels identified by M23 as being $> 50\%$ multi-year ice were classified for the purposes of our analysis as multi-year ice. Pixels identified as FYI by both M23 and OSI-403-d were treated as FYI. All other data was discarded. M23 is only available since 2013, curtailing the length of our CryoSat-2 analysis in the Southern Hemisphere by two seasons.

Our surface-based observations concern the mean σ^0 of a piece of sea ice as it evolves and thickens over time. To compare this to satellite data, it would therefore not be most relevant to investigate the mean σ^0 of the entire Arctic/Antarctic icepack as it evolves and thickens over time, because during this time it also expands laterally. To consider the entire winter Arctic icepack would therefore be to consider σ^0 of existing ice as it ages (this is comparable to our surface-based observations), and to mix that signal with the constant appearance of new, young ice at the ice edge.

In the Arctic, we attempt to monitor changes of roughly the same ice by identifying a region of ice that exists on the 1st of November in each year, and then charting the evolution of σ^0 in this region (as the ice edge expands outside the region). The 1st of November is sufficiently late to allow enough ice to have formed for a meaningful analysis over first-year ice, but is also sufficiently early to allow an accurate analysis in early-winter (October/November/December) period of interest. The annual masks are shown in Supplementary Figure S8. There are clearly some weaknesses of this approach involving ice drifting into and out of these masked regions. This means that the ice within the area on the 1st of January cannot be simply

described as an older version of the ice that was inside the area on the preceding 1st of November; some ice will have drifted out of the area, and some younger ice will have likely drifted into the area (with the exception of landfast ice). To investigate whether the results of this analysis are robust, we first conduct a separate analysis in Hudson Bay, where it is difficult for ice to either escape or be replaced once it has formed.

In the Southern Ocean the masking operation is considerably more difficult because of the more seasonal nature of the icepack [42], and the net northerly ice motion from the coasts [43], [44]. In order to track the same ice over a season in Antarctica, the mask underpinning the algorithm would have to evolve dynamically based on ice-motion vectors. We consider the construction of such a masking algorithm beyond the scope of this manuscript, and as such we do not apply a masking operation to the Southern Ocean, finding ultimately that the results are similar between hemispheres. This masking operation is not necessary to track the seasonal σ^0 evolution of multiyear ice in either hemisphere, since new MYI cannot form in winter. As a final attempt to limit the degree to which we account for new ice formation at the pack-edge, we mask out all sea ice equatorward of 66°N/S.

We process along-track σ^0 measurements from the two satellites, interpolating them at daily time-resolution onto the grids of the ice type products using a drop-in-the-bucket averaging scheme [45]. We then apply the mask in the Northern Hemisphere to limit the influence of new-ice formation. We then calculate two average values on each day: the mean σ^0 of MYI pixels, and FYI pixels. Because the day-to-day values of mean σ^0 exhibit significant variability, we display them in this manuscript in decibel scale by taking $10\log_{10}(\sigma^0)$.

V. RESULTS

A. Radar Signatures of Frost Flowers

Prior to our first radar measurements, we were able to carefully clear the frost flowers from a strip of sea ice and investigate the impact on σ^0 . Figure 4 shows that the strip exhibited a much higher radar cross-section than the part of the transect where the frost flowers were left intact. This is perhaps unsurprising, given that the surface was visually much smoother after the frost flowers were cleared without disturbing the underlying ice surface characteristics. This smoothness produces more coherent backscatter in our nadir viewing geometry. Specifically we note that in the geometrical optics model presented by [46] and used by [2], the coherent power returned by a randomly rough surface (σ_c^0) when illuminated by a spherical wavefront decays exponentially with the square of the root-mean-squared distribution of surface height (σ_h) as follows:

$$\sigma_c^0 \propto \exp(-4k^2\sigma_h^2) \quad (1)$$

Where k is the wavenumber corresponding to the Ku-band in free space. This makes clear that as the roughness induced by the frost flowers is removed, the backscatter will increase significantly (with other things such as dielectrics held equal).

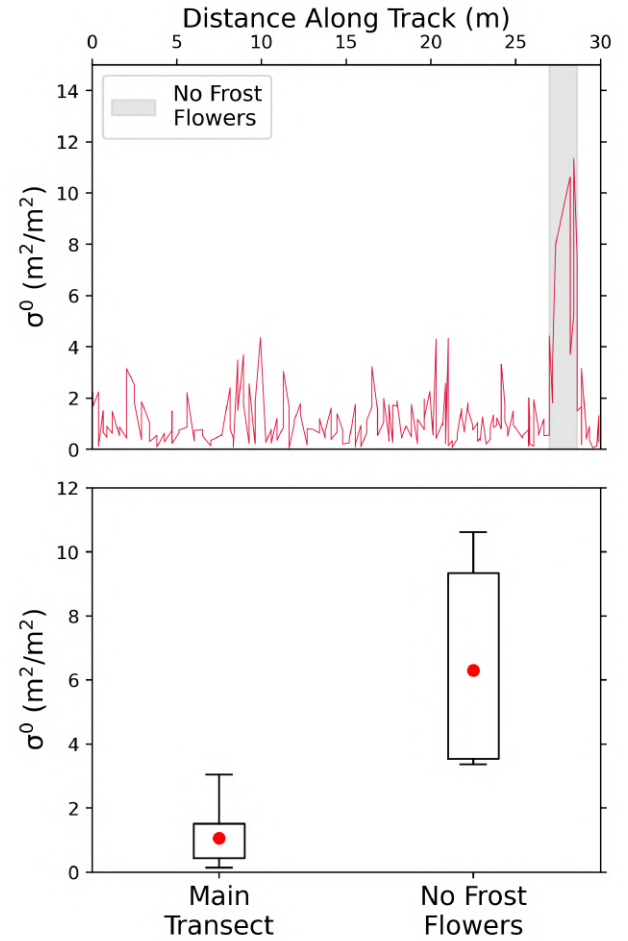


Fig. 4. Waveform σ^0 along the transect on August 30th (prior to snow accumulation). A strip of ice where frost flowers had been cleared from the surface shown shaded in grey. σ^0 values are elevated in this region. Red dots mark the means of the distributions.

The size of the KuKa footprints are evident in Figure 4, as the footprint incrementally overlaps with the cleared strip until it is fully aligned, and then the overlap incrementally reduces again. As such, we cannot be confident that all waveforms with a high σ^0 originate *entirely* from the cleared area - there may be some part of the footprint that lies in the area outside which contained frost flowers. Nonetheless, in pursuit of an indicative value for the frost flowers' effect, we create a simple definition of the strip in our geolocated radar data. We take the first waveform with a σ^0 exceeding any value in the first 0 - 20 m of the transect to mark the "start" of the strip. We take the last waveform with this property to mark the end of the strip. This defines a strip in our radar data with a width of 1.6 m, which is shaded in grey in Figure 4.

B. Time-Evolution of Transect σ^0

The first transect of snow-covered ice occurred on the 31st of August - when analysing σ^0 data from snow-covered ice we excluded the section which was previously cleared. σ^0 values along the transect were higher than they had been the previous day, but lower than the values previously produced by the frost flower-free strip. This may be due to destruction

of the frost-flower structure during snowfall, making the ice surface much smoother and more specular to radar than it was when covered with frost flowers. The second reason that the ice may have backscattered more power is that the frost flowers were measured to be more saline than the ice by 10 ppt. As saltier ice has a higher liquid water content at a given temperature [47], the potential act of flattening them to the ice surface, making this new ice surface more saline, may have therefore increased the dielectric contrast and Fresnel coefficient across the snow/ice interface, increasing backscatter [48]. This is consistent with the marginal increase in salinity by 2 ppt of 5 mm-deep scrapings taken from the ice surface on both days (Figure 5).

After the roughly two-fold increase in transect σ^0 on the first day (driven by the transition from a frost flower cover to a snow cover), σ^0 shows a clear decreasing trend over the remaining timeseries. The mean σ^0 value on the 31st of August (the first day that the transect was surveyed in snow-covered conditions) was 2.06 m²/m². On the final day of the survey (23rd September) the corresponding value was 0.48 m²/m²: a factor of four decrease. When snow covered (i.e. from the 31st onwards), it appears that the spread of σ^0 values along the transect also decreases, with the highest spread being exhibited on the first day of snow cover (31st August).

Our σ^0 timeseries is uncorrelated with the corresponding timeseries of snow depth and air temperature (Supplementary Figure S9). While σ^0 and snow depth generally decrease and increase respectively over the course of the study, jumps in one are not obviously reflected by the other. This is particularly clear between the 20th and 23rd of September, when the snow depth doubled from 3 cm to 6 cm but the mean σ^0 remained almost exactly the same. We stress the need for caution in interpretation of correlations here, due to the limited number of data points.

Statistically significant correlations (at the 5% level) do exist between the timeseries of snow-covered Ku-band σ^0 and both the time elapsed and ice surface salinity (dashed lines of Figure 6). It is subjective as to whether to include the frost flower covered transect in this regression analysis, as it is arguably a special case: we therefore also regress the data including the frost flower transect. We find that including the frost flower data point (marked with a triangle) eliminates statistical significance in the case of the ice surface salinity (regression line not shown in Figure 6), but not when regressing against the time elapsed (dotted line in Figure 6). Once the ice and frost flowers were covered with snow (31st August), our ice surface scrapings became less saline over time, and this is reflected in decreasing σ^0 values. While this suggests that changes to the dielectric properties of the snow/ice interface may be driving σ^0 changes, we do not have proof of a causal link. We discuss the potential for causation in Section VI-A.

As discussed later in Sect. VI-B, our data is uniquely well-positioned for this investigation: the homogenous snow cover allowed the radar system to remain highly level while in motion, generating consistently accurate altimetric σ^0 values over our ~20 m transect. In environments where snow topography typically induces significant tilt, or snow softness causes the sled to sink unevenly, our derived timeseries of σ^0 would be

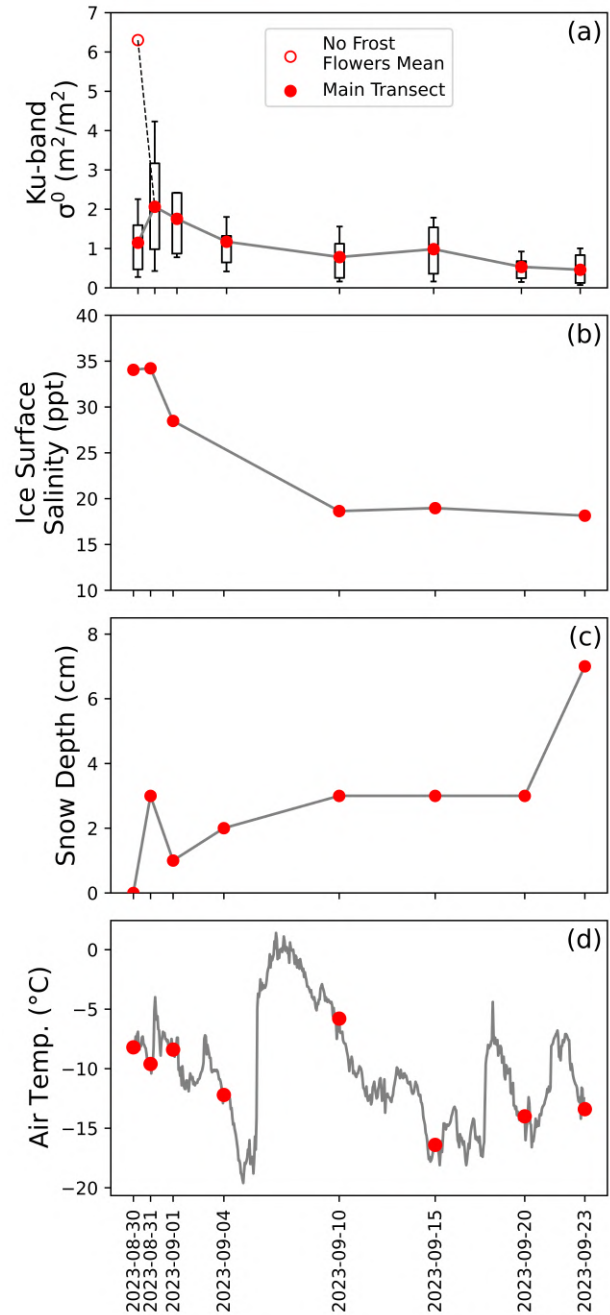


Fig. 5. Timeseries of KuKa-radar derived Ku-band σ^0 and three geophysical variables: ice surface salinity (measured by scraping the ice surface to ~5 mm depth), snow depth (manually measured with a ruler), and air temperature (measured by the Rothera research station automatic weather station). The hollow red marker in top panel indicates mean σ^0 within the grey band of Figure 4, and reflects the surface backscatter in the absence of frost flowers. Red markers in the lower panel show the temperature at the time of the transect.

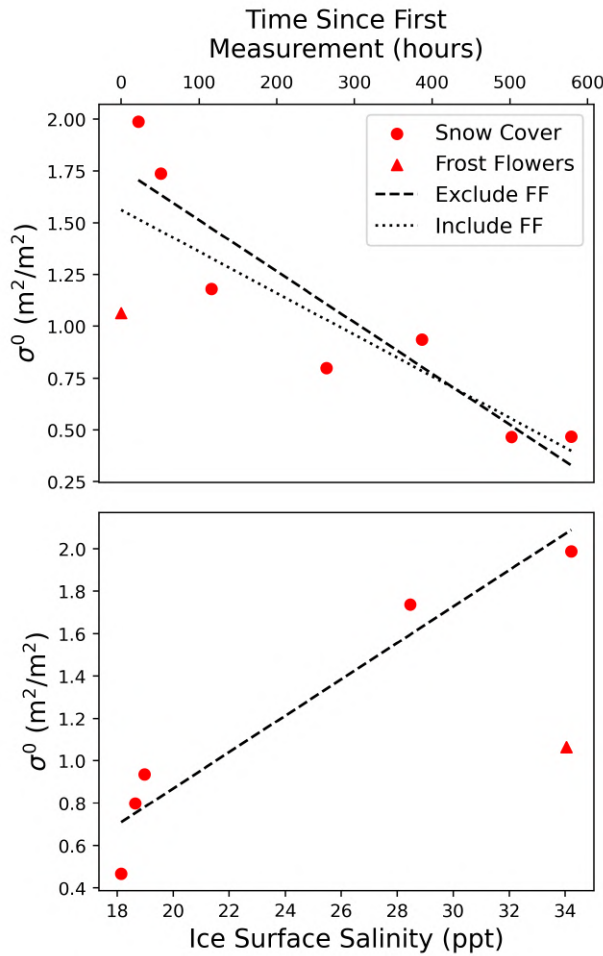


Fig. 6. Relationships between measured σ^0 and two variables: time and ice surface salinity. Correlations were tested at the 5% level, both including and excluding the transect which featured frost flowers (30th August). This (perhaps anomalous) data point is shown with a triangular marker. Where statistically significant relationships exist, they are shown with dashed line where August 30th is included, and a dotted line where August 30th is excluded.

less reliable without tight filtering for instrument tilt. We are fortunate that the level, thin snow conditions did not pose this challenge to our transects.

C. Waveform Analysis

We performed a peak-classification exercise on waveforms from our transects over snow covered sea ice. This was informed by results from the frost-flower covered ice on the 30th of August (Figure 3) which shows that the trailing-edge sidelobe was suppressed but not fully removed by the deconvolution process described by [34]. As such, we classify *single-peaks* as waveforms with either a single maximum in the range window 1.4 - 1.7 m, or where there are two maxima but one is potentially a suppressed sidelobe. Suppressed sidelobes are defined as local maxima trailing the global maximum by between 0 and 10 cm, but being less than 5% of the global maximum's amplitude. Double peaked waveforms are defined as waveforms with two local maxima where neither is a suppressed sidelobe of the other. Waveforms in the analysis

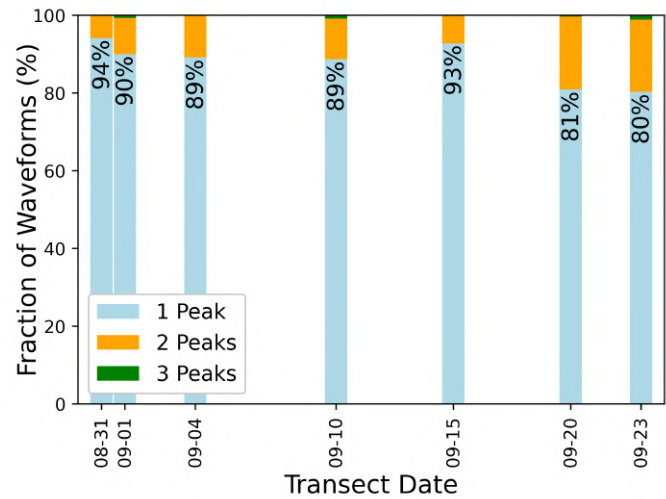


Fig. 7. Waveform peak number classification by transect for all dates with snow-cover, with the relevant fraction of single, double, and triple peaks shown in blue, orange and green respectively. Triple-peaked waveforms were so rare that they are minimally displayed on the plot. White annotations indicate specific percentage of single-peaked waveforms in a transect.

window with three local maxima were extremely rare, even before considering sidelobes. They were not present on four of the dates, represented less than 1% of the waveforms on three other dates, and represented 1.1% of waveforms on the 23rd of September.

When classified as above, the single-peaked waveforms of a given transect represent the large majority of waveforms in a given transect. However, this dominance decreases over time, dropping to around 80% for the last two transect-days of our study (Figure 7). On average, these single peaks with amplitudes between three and ten times the power of peaks that are the first-of-two, and 10 - 40 times that of peaks that are the second-of-two (see Figure 8a & b). Critically, these peaks have ranges corresponding most closely to the ranges *second* peak of double-peaked waveforms, suggesting that they represent highly specular reflections from the ice surface and are simply masking out the return from the snow surface.

We stress that the peak amplitudes shown in Figure 8b should not be interpreted as a quantitative representation of how backscattered power is partitioned between the snow and ice surfaces respectively, because these double-peaked waveforms are an artificially selected subset of all waveforms where peak sizes must be comparable to each other to within an order of magnitude. In the case where backscatter from one interface falls below a value where it can be resolved relative to the adjacent peak, it does not contribute low values to the averages shown in panel b, but rather the whole waveform is reclassified as 'single-peaked' and contributes to panel a.

D. Hudson Bay as a Natural σ^0 Laboratory

In the previous sections, we documented the decreasing Ku-band σ^0 of young sea ice over a 24-day period, when measured with a surface-based Ku-band radar. We attributed this to the decreasing power of backscatter from the ice surface. We now show that a similar pattern emerges from CryoSat-2 data,

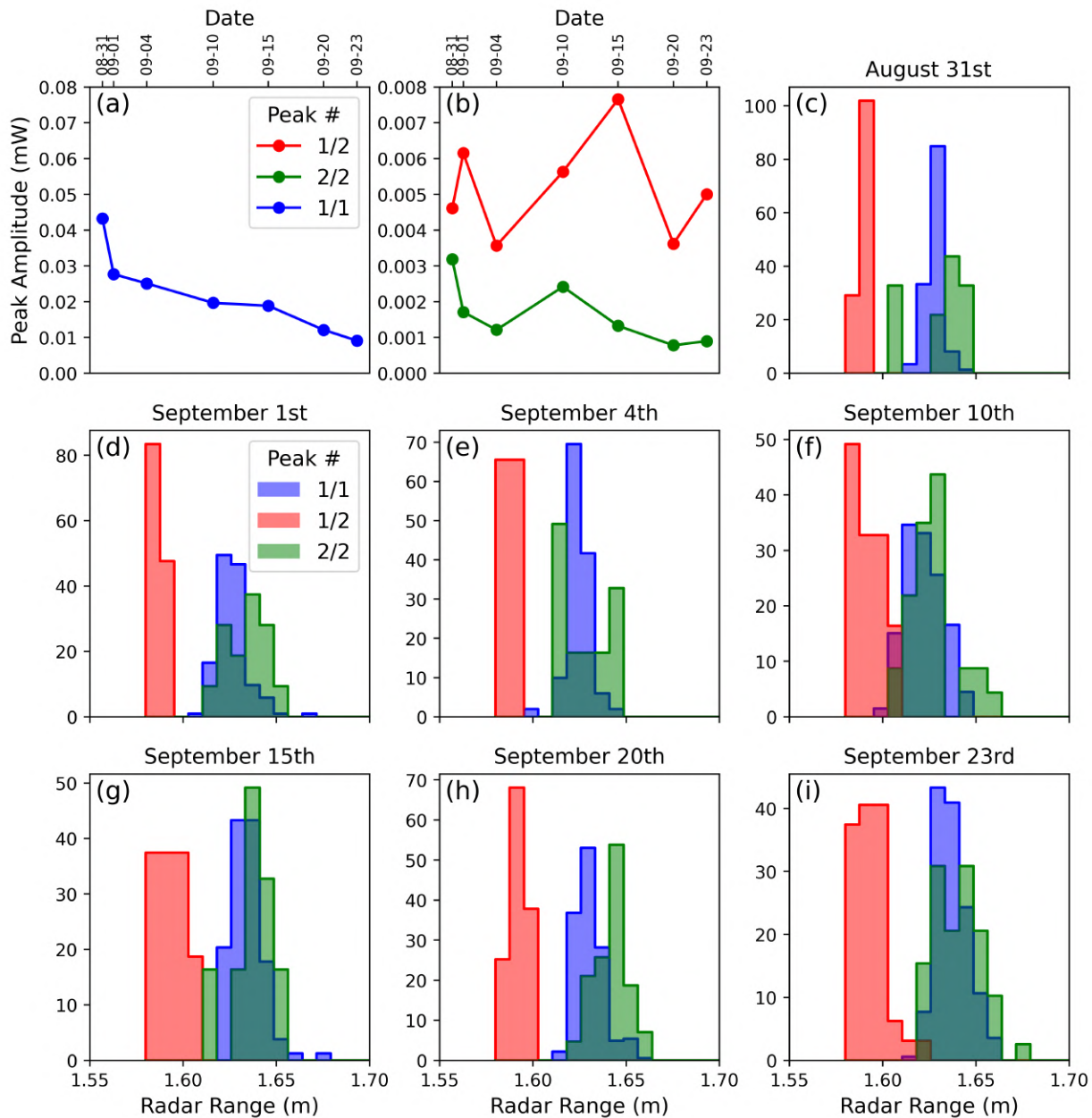


Fig. 8. Positions of classified peaks on days with snow cover. Panels a & b indicate the amplitude of single and double-peaked waveforms respectively. Note the order-of-magnitude difference in y-axis scales, which necessitated display on different panels. Panels c - i indicate the distribution of the radar ranges of these peaks over the course of the study. Areas of histograms are normalised, but it should be considered that as well as being more powerful, single peaks are far more numerous. The powerful single peaks are associated with the second of the double-peaks in range, suggesting that they represent specular reflection from the ice surface. These specular reflections diminish in power over time (panel a).

focusing on an area with limited ice export and import: Hudson Bay.

Hudson Bay is a seasonally ice-free basin in Canada, connected to the global ocean via Foxe Basin to the North and Hudson Strait to the North-East. Mansel, Coats and Southampton Island separate Hudson Bay from Foxe Basin and the Hudson Strait, leading to relatively little ice loss or addition from the region. Because Hudson Bay is ice-free in summer, the ice within it is consistently first-year ice. Critically, Hudson Bay is large enough to be well-sampled by CryoSat-2 every day during the winter season, allowing a

robust characterisation of its σ^0 evolution at appropriate time-resolution. This is not the case for areas of fast-ice such as Belgica Bank (east of Greenland) or Elson Lagoon (northeast of Utqiagvik, Alaska) which were other candidate-sites for this analysis.

Figure 9 illustrates the winter evolution of CryoSat-2 σ^0 in Hudson Bay. There is a clear and decreasing trend in σ^0 from mid-November to the start of March, with σ^0 then remaining stable until the start of May, before increasing again.

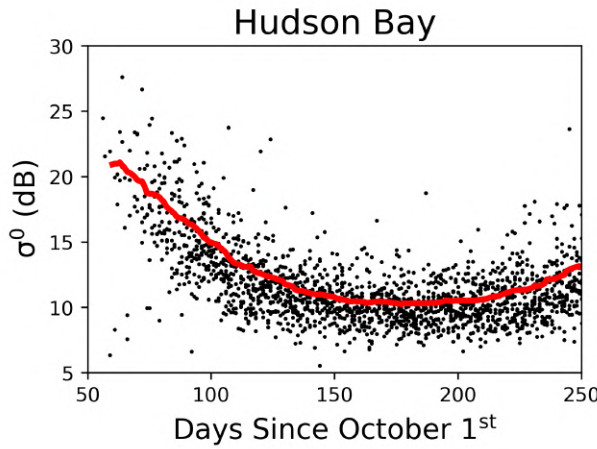


Fig. 9. Seasonal evolution of Ku-band sea ice σ^0 in Hudson Bay, with a 10-day running mean marked in red. Running means do not appear ‘centered’ among the grey scatter points as they are calculated on values prior to conversion to the logarithmic scale.

E. Seasonal Evolution of CryoSat-2 σ^0

We now show that our results hold true in both polar regions, by repeating the analysis from Section V-D over the full Arctic Ocean, masking out regions not covered by sea ice on the 1st of November. The σ^0 data used here are somewhat independent of our previous Hudson Bay analysis for two reasons: the Hudson Bay data are often masked out of a pan-Arctic analysis because Hudson Bay generally freezes up later than our 1st November cutoff date in the Arctic; but also because in any case sea ice south of 66°N was not considered here.

Figure 10 illustrates the winter evolution of CryoSat-2 σ^0 in the Arctic (top row), where we mask out regions not covered by sea ice on the 1st of November. There is a clear and decreasing trend in σ^0 in the months of October and November, with a smaller but still present trend in December. These early-winter σ^0 values are much more elevated over FYI by comparison to MYI, although the MYI does qualitatively show a similar phenomenon.

We find that the σ^0 evolution of ice in the Arctic Ocean is similar to our previous results for Hudson Bay (Figure 9). There are two key differences: σ^0 values are generally slightly lower, and the timeseries is shifted earlier in the season, due to the Hudson Bay’s delayed freeze-up. This similarity indicates that early-winter reductions in σ^0 found here are geophysically real, as opposed to being artefacts of the masking process used in this section.

The early winter evolution of our Arctic values is consistent with Table A2 of [49], which recommends increasingly elevated minimum σ^0 thresholds for lead discrimination of 28, 25.8 and 24.1 dB in October, November and December respectively in CryoSat-2 SAR mode. This is to avoid misidentifying highly specular early-winter returns from sea ice as leads for sea level height interpolation.

As described earlier, our masking operation is harder to apply in the Southern Ocean and as such we forgo it. When performing our analysis in an otherwise similar fashion (Figure 10, bottom row), we find qualitatively the same results for FYI:

σ^0 of the sea ice is maximal at the start of the winter season (in March), and then declines steeply. The data is considerably more scattered for MYI in the Southern Ocean. Some of the weakened “early winter” signal may be attributable to the relative lack of sophistication in the masking process.

VI. DISCUSSION

A. Changes at the Snow/Ice Interface of Young Ice

Our surface-based radar measurements extend the study of [50], which recorded reductions in altimetric σ^0 in the C-, X- and Ka-band between 9 and 62 hours after ice began forming in a sea ice tank facility. While this study was considerably shorter than ours and did not include radar measurements of snow covered ice, declining σ^0 values were observed in all three frequencies over the study period. This was attributed to declining near-surface brine content of the ice.

We also document a declining σ^0 over twenty-four days when measured with a surface-based radar, and notice a correlation with reductions in the salinity of scrapings from the ice surface and snow thickness. As well as being consistent with [50], a desalinating ice surface is consistent with data from both the Arctic [51], [52] and the Southern Ocean [53], [54], which generally exhibit a negative correlation between ice thickness and ice surface salinity. Modelling results also support desalination in sub-zero conditions [55]. Desalination of the sea ice surface over time was also observed in the tank facility which hosted the radar study of [2] (see Figure 9-4 of [56]). Based on its prevalence in the literature, we conclude that the desalinating ice surface that we observed is not a statistical artifact or series of random fluctuations.

Having established a correlation between σ^0 and ice surface salinity, we now point out three ways in which the relationship might be causal. (1) Declining ice salinity would decrease the dielectric permittivity of the sea ice [57], [58]. This would decrease the dielectric contrast, Fresnel coefficient and surface-scattering across the snow/ice interface [48], leading to reductions in backscatter. (2) If some of the ‘lost’ salt from the ice surface migrated upward into the snow (as observed by [59], [60], [54], [56], [61], [62]), then this would increase the dielectric permittivity of the snow above the snow/ice interface by increasing its wetness at a given temperature via freezing point depression - we calculate this effect below. This would further reduce the Fresnel reflection coefficient across the interface, decreasing its backscattering contribution. (3) Increasing snow salinity would do more than just decrease the dielectric contrast across the snow/ice interface; it would also reduce the incident power on the interface by increasing microwave absorption directly above. In this case, even if the dielectric contrast across the interface were to somehow remain constant, its backscattering contribution would still decrease [60], [63].

While scrapings of the ice surface are relatively easy to take consistently (and so are presented here), snow samples very near the base are very difficult to obtain consistently without inadvertently scraping the ice surface, which can substantially contaminate a sample. We find our samples of the lowest 1 cm of the snowpack to vary substantially

CryoSat-2

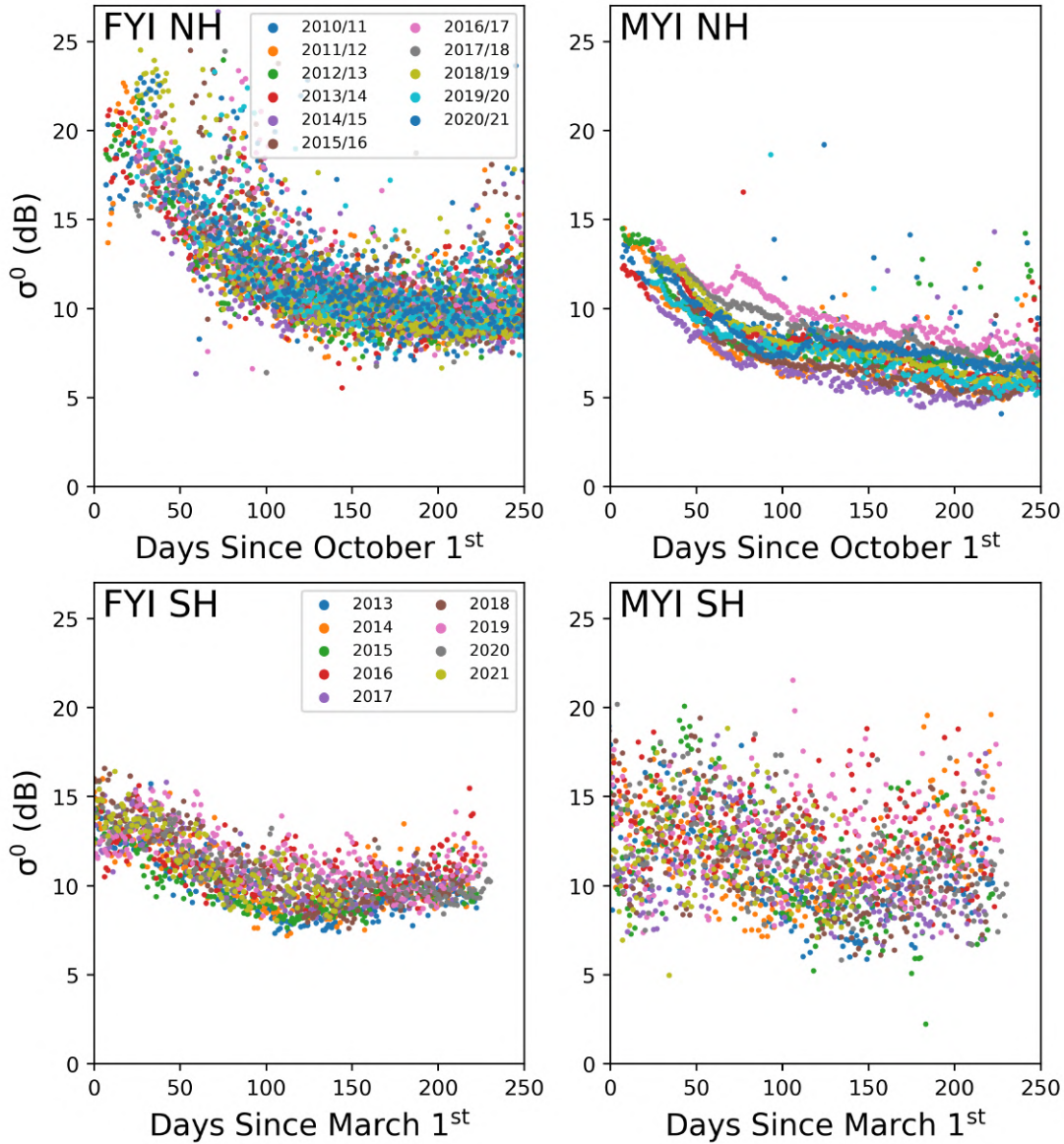


Fig. 10. Seasonal evolution of Arctic (top row) and Antarctic (bottom row) sea ice Ku-band σ^0 from CryoSat-2. Arctic data is shown for areas covered by sea ice on October 1st. For FYI, values are elevated in early winter shortly after freeze-up. This is also the case for Arctic MYI, but a robust signal is not detectable for Antarctic MYI - we discuss potential reasons for this in Sect. VI-E. In the Northern Hemisphere, ice type was classified using the OSISAF 403-d product [40]. In the southern hemisphere, MYI pixels in the OSISAF product were also only analysed for NRCS when flagged as being more than 50% in the MYI product of [41].

from measurement to measurement. It is therefore a challenge to separate methodological variability from true, small-scale spatial variability because of this. Due to these constraints (but also due to logistical reasons) we are not able to present a consistent and representative timeseries of the absolute basal snow salinity for the transect. This means we are not in a position to produce a timeseries of the dielectric contrast across the snow/ice interface, but this should be a priority for future studies of this kind.

While we do not produce a timeseries of dielectric contrast, we are able to show that our hypothesised changes in salinity

could account for the four-fold reduction seen during the 24-day period when the ice was snow-covered. We do this using Equation 7 of Beaven et al. [2] which is reproduced from Equation 6 of Fung and Eom [46]. This describes the coherent component of the backscatter from a randomly rough surface illuminated by a narrow beam with a spherical wavefront as:

$$\sigma_c^0 \cong \frac{\Gamma_p}{B^2} \exp(-4k^2\sigma_h^2) \exp\left(-\frac{\theta^2}{B^2}\right) \quad (2)$$

when

$$B^2 = \frac{1}{(kR_0\beta)^2} + \left(\frac{\beta}{2}\right)^2 \quad (3)$$

and when $\theta = 0$ the reflection coefficient (Γ_p) can be written:

$$\Gamma_p = \left| \frac{n_1 - n_2}{n_1 + n_2} \right|^2 = \left| \frac{\sqrt{\epsilon_{r1}} - \sqrt{\epsilon_{r2}}}{\sqrt{\epsilon_{r1}} + \sqrt{\epsilon_{r2}}} \right|^2 \quad (4)$$

where β_0 is the 3 dB beamwidth of the antenna and R_0 is the range to the scatterer, σ_h is the RMS roughness of the surface, k is the electromagnetic wavenumber, θ is the incidence angle and n_i is the refractive index of the snow ($i = 1$) or ice ($i = 2$), and ϵ_r is the relative dielectric permittivity. This equation (also discussed in [25]) makes clear that σ^0 scales linearly with Γ_p for a fixed radar system (encoded in B) and a surface with given roughness characteristics encoded in σ_h . We can calculate the impact of changes in ice and snow salinity on the dielectrics, and the knock-on impact on Γ_p and then the coherent component of σ^0 coming from the snow/ice interface.

To illustrate the potential impact of a desalination of the ice by 10 ppt (from 30 to 20 as seen in Figure 5), we must first consider where the salt lost from the ice goes. For simplicity, we assume that around 30% of it travels upwards into the basal snow, and the remaining fraction moves downwards and away from the snow/ice interface. This would allow snow that is 30% the density of ice to increase its salinity by 10 ppt in response to sea ice desalination of 10 ppt.

The surface air temperatures of our transects were typically between -5°C and -15°C , and for this investigation we fix the temperature of the snow-ice interface at -10°C . This allows us to model the Ku-band dielectric permittivity (ϵ_r) of the snow and ice using tools from the Snow Microwave Radiative Transfer Model (SMRT) [64]. Specifically, we use the temperature, density and bulk-salinity dependent formula from [65] to calculate the complex dielectric permittivity of the snow. We then use the salinity and temperature-dependent formula from [47] to calculate the brine volume fraction of the ice, and use that to calculate ϵ_r by using the Polder Van Santen mixing model for saline ice from SMRT. If we assume that the basal snow salinity starts at 20 ppt, we find that a reduction (increase) in ice (snow) salinity by 10 ppt reduces the Fresnel reflection coefficient (Γ_p) at the snow/ice interface from 0.018 to 0.002 - a factor of 8.35. If the basal snow salinity begins at 10 ppt, the Γ_p is reduced by a factor of 4.05. This shows that changes to the ice surface salinity of the magnitude measured in this study can reproduce the observed reductions in σ^0 .

We stress however that simply because these salinity reductions *can* produce the observed changes in σ^0 , it does not mean they are solely responsible. For instance, metamorphic changes in the basal snow grains (i.e. hoar formation) could well be relevant in progressively reducing the power that reaches the highly scattering snow-ice interface over time. We also note that our calculations above do not include changes to the two-way loss factor of the snow as it salinifies, which would also decrease the backscatter from the snow-ice interface independently of changes to the right hand side of Equation 2.

There is also a role for temperature in our analysis, which would impact the Fresnel coefficients discussed above. Although Figure 5 makes clear that there is no long-term temperature trend in our data, there are clear daily and hourly fluctuations. We should therefore question whether the signal

from the salinity change could even be discerned above the "noise" of the temperature fluctuations.

To investigate this we once again calculate Fresnel coefficients (Γ_p in Equation 4) which are linearly related to the coherent component of the radar return per Equation 2. We first show the temperature sensitivity of Γ_p for a modelled snow/ice interface where the snow is 300 kg/m^3 and the ice is 30 ppt. Given we do not have robust estimates of the snow salinity, we calculate the sensitivity of the Fresnel coefficients for snow salinities of 10, 20 and 30 ppt (Figure 11 upper). This indicates that the temperature sensitivity of Γ_p is relatively low for temperatures below 265 K (-8°C). Above this, Γ_p increases more quickly for the 10 ppt snow, whereas the Γ_p sensitivity for the 20 and 30 ppt snow remains comparatively stable until temperatures rise above 268 K (-5°C). This relative stability compared to sensitivity to salinity (see above) stems from the brine volume fraction and refractive indices of the snow and ice changing in unison and partly cancelling each other out in the calculation of Γ_p .

Taking the surface air temperature evolution as a rough proxy for the temperature evolution of the snow-ice interface, we can also make an approximate timeseries of Γ_p for our three selected snow salinity values (Figure 11 lower). The timeseries are relatively flat for snow of 20 and 30 ppt. At 10 ppt changes in the snow brine volume fraction do not effectively cancel out the changes in the ice brine volume fraction for a given temperature change, so the sensitivity is larger. For instance the Γ_p value is in excess of 0.05 on the 10th of September, compared to a typical value of between 0.03 and 0.04 on all other snow-covered days. This represents an anomaly of around 60%. This scale of anomaly can be compared to our salinity analysis above, where we found that reductions in ice surface salinity (with some flux up into the snow) could lead to a quadrupling or even an eight-fold decrease in Γ_p values. We therefore suggest that the impact of salinity changes discussed in the manuscript are likely larger (and more systematic over time) than the temperature fluctuations.

B. The Role of Instrument Tilt in Our Surface-Based Data

Our surface-based study was terminated by the destruction of the transect due to unavoidable operations involving the nearby aircraft runway, but it is worth briefly considering how it might have continued. A significant snowfall occurred prior to our final transect measurement, doubling the mean snow depth on our transect from 3 to 6 cm. During our final transect, the snow depth was spatially homogeneous. This meant the sled on which the KuKa radar was mounted remained consistently level, and measurements were taken within a very narrow range of tilts. When the snow was 6 cm deep on the 23rd, the standard deviation of the instrument's tilt was 0.40° and 0.66° in the across- and along-track directions respectively. The along-track tilt is slightly larger due to the jerking motion of the sled-pulling force, which slightly rocks the instrument in the along-track direction.

After the conclusion of our study, the nearby snow properties and topography both changed, as the snow underwent

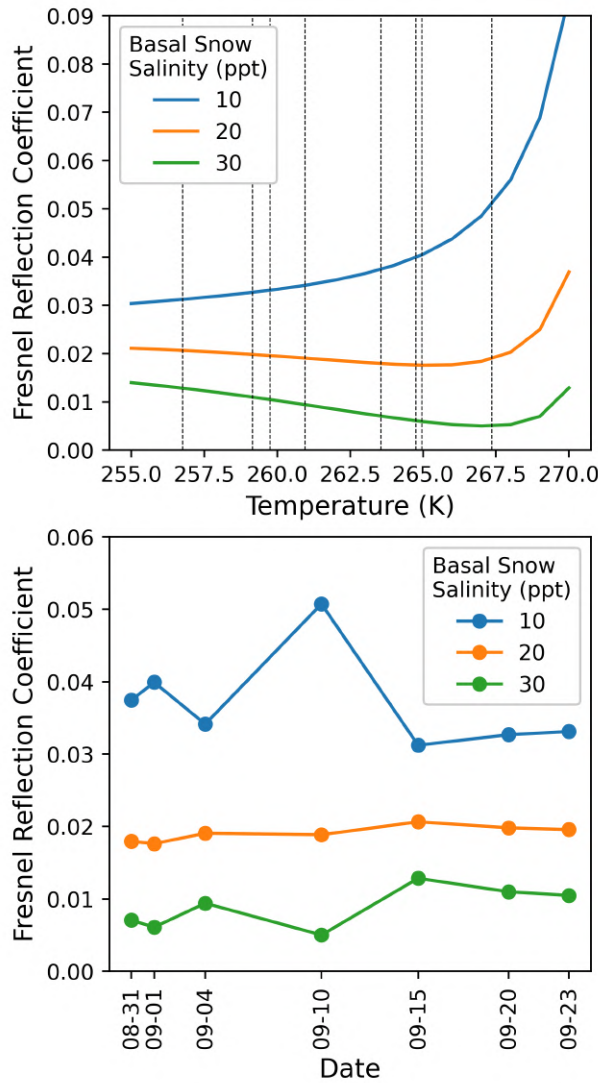


Fig. 11. **Top:** temperature sensitivity of Fresnel coefficient at a modelled snow/ice interface for three different snow salinities (density 300 kg/m³) and an ice salinity of 30 ppt. Because the brine volume fractions of the saline snow and ice respond in unison to temperature changes, there is relatively little sensitivity in the Fresnel coefficients to temperature changes below 265 K. Above this the sensitivity increases with decreasing snow salinity. Vertical dashed lines indicate the surface air temperature measured nearby at the time of radar transects. **Bottom:** modelled timeseries of Fresnel coefficients on all snow-covered days based on nearby measurements of surface air temperature shown in Figure 5. For all salinities the fluctuations in the Fresnel coefficients are small relative to the changes described above in response to salt fluxes.

microstructural metamorphism and was redistributed by the wind. Generally this type of evolution leads to an environment where the sled is typically not exactly level, and assumes a tilt relative to the ice surface due to the modified snow cover. A similar analysis of tilt data from the ‘Northern Loop’ of the MOSAiC expedition (published in [11]) indicates that the standard deviation of along-track (across-track) tilts for a given loop is always in excess of 1.5° (1.8°) and sometimes as high as 2.2° (2.5°). To put this more simply, the tilt-variability of the KuKa instrument during the MOSAiC expedition was between three and five times larger than the tilt-variability of this study’s most variable transect. This did not pose a

problem for the analysis presented by [11] since it was not dependent on σ^0 calculations, but instead addressed a snow-depth determination method. Given our σ^0 calculations are likely sensitive to this tilt, we suggest that without extensive data filtering for tilt we would have seen a degradation in the quality of our data should our experiment have continued much longer, and the snow become deeper and more redistributed.

C. Altimetric σ^0 in the Ka-Band

This manuscript has focused on the Ku-band of the electromagnetic spectrum, in order to address the tension between recent surface-based studies and that of Beaven et al. [2]. However, questions also exist surrounding the degree of snow-pack penetration by radars operating in the Ka-band [39], [66]. For instance, a recent study of AltiKa/SARAL data suggested that the radar retrievals tracked a median elevation 10% below the snow-air interface [67] with [39] having previously put it around 46%. This is relevant to the AltiKa/SARAL mission but also the upcoming CRISTAL mission which will use both the Ku- and Ka-bands.

σ^0 data from our surface-based measurements and the AltiKa mission indicate that the evolution is very similar to that of the Ku-band when the sea ice was snow-covered (Supplementary Figure S10). The only case when the evolution of mean σ^0 diverges significantly is immediately after snow accumulation, when the Ka-band transect-mean σ^0 jumps to ~50% higher than the Ku-band mean. This is largely driven by a few very high values, which give the distribution of σ^0 values along the transect a positive skewness, and drive the mean value above the upper interquartile range.

We now reproduce the method behind Figure 10, but in this case with the AltiKa satellite in the years 2015-2021. As well as having a shorter time range, our analysis is also hampered by the Arctic’s significant ‘pole hole’ of the mission, which precludes measurements above 81.5°N of latitude. Despite these differences, the seasonal evolution of σ^0 over both ice types appears very similar to the CryoSat-2 case (Figure 12). This mirrors the similarity of σ^0 evolution as measured at the surface.

Many dual-frequency snow depth retrieval methods exploit differential penetration of radar waves in the Ka- and Ku-bands, with an assumption that Ka-band ranges broadly reflect the range of the snow surface, and Ku-band signals reflect the range of the ice surface [66], [68], [69], [70]. The degree to which the backscattering properties of these frequencies differ is therefore relevant to these methods, and the operation of the planned dual-frequency CRISTAL mission [71]. From our results, we conclude that either (a) σ^0 of both the snow-ice and snow-air interfaces evolves very similarly (despite the surfaces evolving quite differently in a geophysical sense) or (b) The Ku-band and Ka-band σ^0 evolves similarly because both frequencies are largely backscattered from a shared target (whether that is the snow-ice interface, the snow-air interface, the snow volume, or a mixture of these).

AltiKa

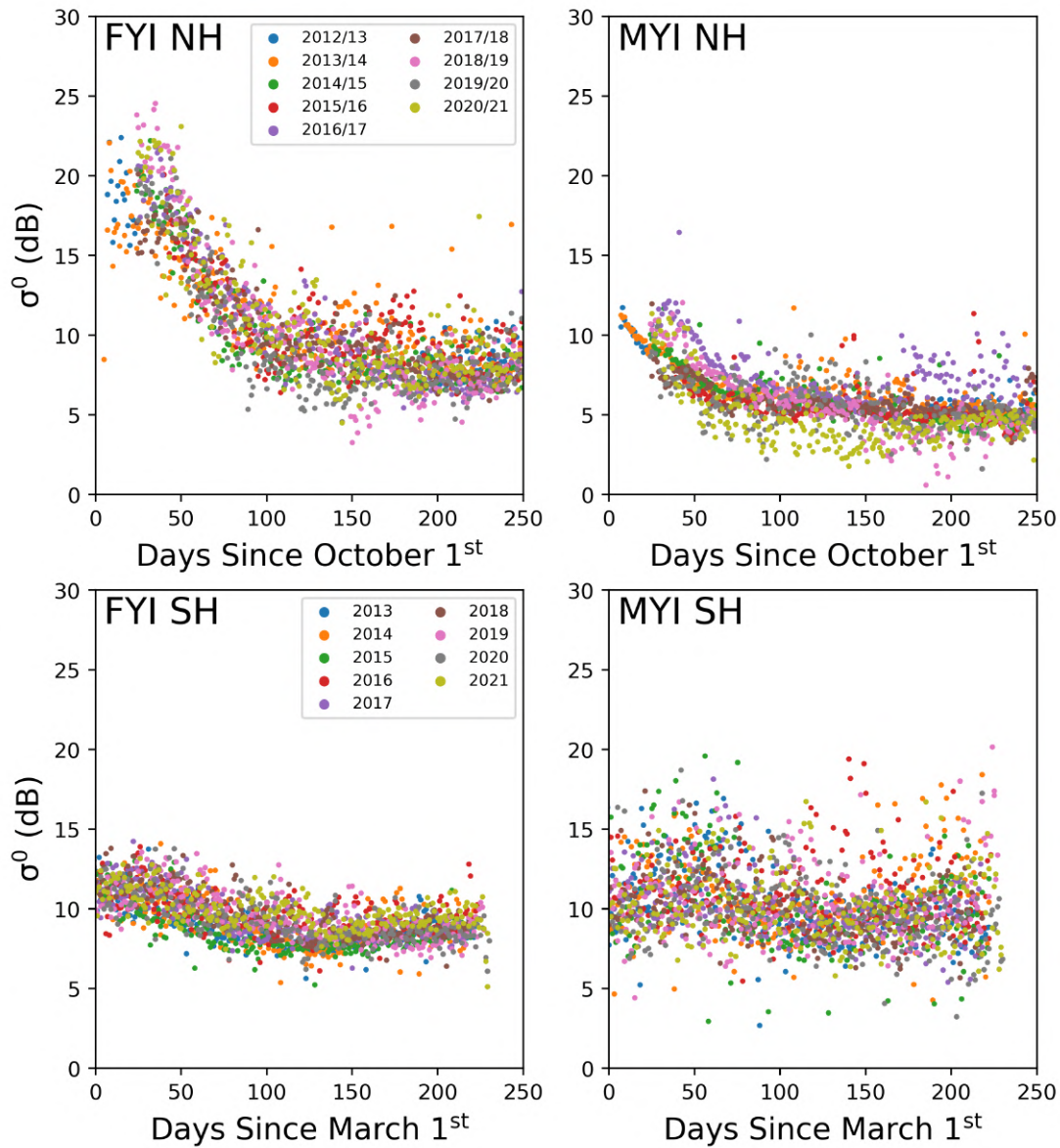


Fig. 12. Seasonal evolution of Arctic (top row) and Antarctic (bottom row) sea ice Ka-band σ^0 . Arctic data is shown for areas covered by sea ice on October 1st. For FYI, values are elevated in early winter shortly after freeze-up. This is also the case for Arctic MYI, but a robust signal is not detectable for Antarctic MYI. A similar pattern is displayed as in Figure 10, indicating that either the air/snow and snow/ice interfaces evolve very similarly on this timescale, or Ka-band and Ku-band radar waves are backscattered from similar targets. In the Northern Hemisphere ice type was classified using the OSISAF 403-d product [40]. In the southern hemisphere, MYI pixels in the OSISAF product were also only analysed for NRCS when flagged as being more than 50% MYI the product of [41].

D. Other Causes of Early-Winter Reductions in Satellite-Derived σ^0

Our surface-based results document a four-fold reduction in σ^0 of young ice over 24-days, and we find this to be largely driven by the weakening of strong specular reflections from the ice surface. Our investigation of satellite-derived σ^0 values indicates that our surface-based results are *consistent* with the satellite data. However, it is important to consider that the simple process of floe-ageing may not be the only contributor to σ^0 reductions seen in the satellite timeseries.

We now discuss other contributors.

Pack ice roughens over the course of a winter season due to ice dynamics (see Figure 6 of [72]). This roughening will act to reduce the altimetric σ^0 of the sea ice cover in a way that is indistinguishable from the ageing-process observed in-situ. However, our surface-based timeseries shows that σ^0 reductions are not driven *solely* by dynamic roughening, since this did not occur on our transect. A similar argument can be made regarding the opening of leads in the ice pack. One may suggest that as first-year ice thickens over the course of the winter, it becomes less prone to lead-formation. While we

have followed standard practice to exclude radar waveforms emanating from leads in this study, it is possible that this is not 100% effective; this raises the possibility that the higher satellite-derived σ^0 values in early winter simply reflect the higher presence of unidentified leads in a thinner, more dynamic ice-pack. We are not in a position to exclude this effect, but instead again point to our (ice-only) in-situ data. We can conclude that the simple ageing process (observed in-situ) does contribute to the satellite-derived σ^0 reductions, but is not the only contributor.

E. What Might Cause Early-Winter σ^0 Reductions over Multiyear Ice?

In Sect. VI-D we identified the processes of roughening and reduced lead-forming as possible other contributors to early-winter reductions in σ^0 of young-ice as measured by satellite altimeters, while highlighting that they cannot contribute to our in-situ results. It is possible that these processes may be contributing to the more limited early-winter reductions in multiyear-ice σ^0 , as seen in Figures 10 & 12. In addition, MYI also undergoes an ageing process in the early winter. This is particularly the case in the Arctic, where bare sea ice and melt ponds are often exposed in summer after the snow cover entirely melts back, before accumulating snow cover. As well as roughening and decreased lead-forming, we can therefore not discount the possibility that rapid ageing-related evolution in the backscattering properties of the snow-ice interface of multiyear sea ice is a contributor to σ^0 reductions seen in Figures 10 & 12.

What then might cause the strong interhemispheric differences between the early-winter evolution of multi-year ice σ^0 ? The Southern Ocean timeseries of MYI in both Figures 10 & 12 are noisier, and while they do suggest early-winter reductions, such reductions are much less clear. We suggest four contributing factors: (1) Even prior to post-2016 reductions, the annual minimum extent of Southern Ocean sea ice is lower than that of the Arctic Ocean [73], and as such there is reduced MYI coverage. This leads the daily average of pixel- σ^0 values to have fewer constituents. Averaging fewer data-points may lead to higher day-to-day noise. (2) The Southern Ocean MYI pack has (until recently) been noticeably non-contiguous, dominated by two separate components in the Weddell Sea and the Ross Sea on opposite sides of the continent [41]. The meteorological experience of these two components will be disparate, compared to the multiyear ice of the Arctic which is largely contiguous. This may lead to more short timescale variability in the aggregated timeseries. (3) Satellite identification of MYI cover in Antarctica is less well advanced [41]. This led to the use of two ice-type products in our analysis, and potentially introduces noise by the misidentification of MYI pixels from day-to-day as pixels are included then excluded from classification based on small geophysical changes. (4) The geophysical conditions at the snow-ice interface are potentially very different over Antarctic MYI, relative to FYI [73]. Antarctic MYI is known to retain more summer snow cover through the sea ice minimum with the ice surface accumulating a remnant snow cover and superimposed ice [74]. The increased prevalence of flooding events

in the Southern Ocean [75] may also alter σ^0 of the snow-ice interface on short timescales, unlike in the Arctic. The impact of slush on C-band signatures from FYI was noted by [76], albeit at incidence angles between 30 and 75°. However, this may explain some of the interhemispheric differences between FYI σ^0 . It remains unclear whether the Southern Ocean's limited MYI is more prone to flooding than its FYI, but it cannot be ruled out as a driver of the ice-type disparity in Southern Hemisphere σ^0 evolution. While it is beyond the scope of this manuscript to go further, future investigation into ice-type dependent differences in σ^0 is warranted.

F. Resolving Disparate Surface Based Measurements of Ku-band Radar Penetration through Snow Cover

The study by Beaven et al. [2] has had a profound impact on the field of sea ice altimetry, providing motivation for the launch of the CryoSat-2 satellite [3]. The study is frequently referenced in support of the assumption that Ku-band radar waves from CryoSat-2 and similar altimeters penetrate the snow overlying polar sea ice (e.g. [77], [6], [78], [79], [4], [80], [81], [70], [82], [83]). This is despite later in-situ work [9] indicating a relatively substantial radar return from the snow surface. Previously published waveform data from the KuKa radar system featured here [33], [11], [10] suggest a similar conclusion: some Ku-band radar power is returned from the snow/ice interface, but a significant amount (relative to the return from the ice) is backscattered from above. We suggest that the relative contribution from the snow surface increases over time (due to the absolute contribution from the ice surface dropping), helping to explain the discrepancy between Beaven et al. [2] and later studies.

Although we have discussed some potential drivers of reducing σ^0 involving snow and ice salinity, knowledge of the exact drivers is not central to our key argument which is as follows: if the snow/ice interface of young, thin sea ice has a substantially higher σ^0 than that of thicker, older sea ice, then it may have produced a very strong return and dominated that from the snow surface in [2], but not in subsequent studies such as [9], [11], [10] where the snow makes a larger *relative* contribution to the total backscatter.

There are three other key geophysical differences between the tank-based experiment of [2] and subsequent field studies of radar penetration (e.g. [9], [11]). One is the salinity of the seawater from which the sea ice was formed: the water in [2] had a salinity of 24 parts per thousand, whereas the sea ice studied here and in the research cited above was formed from "natural" seawater of around 35 parts per thousand. If the lower ice salinity led to a less saline 'brine skim' [30], then this leaves open the possibility that lower salinity basal snow gave the snow-ice interface a greater dielectric contrast and Fresnel coefficient, increasing the contribution from the ice surface (thus diminishing the relative contribution from the snow). The second difference is that the tank-grown sea ice studied by [2] was smooth by comparison to pancake ice of the same age formed from a wavy sea surface in some studies. This is potentially a separate explanatory variable for the discrepancy between [2] and subsequent in-situ characterisations of radar

penetration of snow. The third salient difference is the mode of snow accumulation. In the study of [2], non-saline snow was applied in four one-inch layers, and potentially measured before there was any brine transfer from the ice to the snow. In field studies, basal snow has generally been in-situ for enough time to absorb some salt, reducing dielectric contrast and Fresnel coefficient across the snow/ice interface and diminishing the corresponding backscatter (see Sect VI-A). Despite these differences, our field data from Rothera highlight the extremely thin and young nature of the sea ice studied by [2] as a plausible driver of the subsequent discrepancy with field studies.

Our analysis above begs the question: if the 12 cm thick ice floe grown in 1992 and measured by [2] had been allowed to age, would its backscatter diminish to the point where backscatter from the snow surface would become significant in a relative sense? To provide an insight, we highlight the follow-up experiment that was conducted at the same facility in 1994, where a 7 cm snow cover was accumulated naturally over ice which was allowed to grow longer and become thicker, to around 33 cm [84], [23]. Measurements were made with an ultrawideband radar (2 - 18 GHz), which would be expected to show a weaker return from a given snow surface relative to the 1992 Ku-band measurements [85] which are at a higher average frequency. Unlike the 1992 Ku-band measurements of 12 cm thick ice, the snow surface is clearly visible in the 1994 ultra-wideband measurements of >30 cm thick ice [86], [24], consistent with our findings.

VII. SUMMARY, FUTURE DIRECTIONS AND CONCLUSIONS

We have presented a 24-day timeseries of surface-based radar backscatter from young sea ice, beginning one day before snow accumulated. Measurements from the first day illustrate the profound power of frost flowers to suppress altimetric radar backscatter in both bands. Subsequently collected data indicate that the normalised radar cross-section (σ^0) declines by approximately a factor of four in the first few weeks following initial snow accumulation when measured in-situ. These declines were associated with the weakening of strong, specular reflections from the snow/ice interface. These σ^0 reductions are consistent with the early-winter evolution of CryoSat-2 derived σ^0 of sea ice in Hudson Bay where we are able to track the ageing of ice without signal contamination from ice dynamics, drifting and new-ice formation. However the signal is also consistent with that of the wider Arctic, and to a lesser extent Antarctica (where it is harder to track the ice).

In-situ, these reductions reflect reduced backscatter from the sea ice surface over time, driven by ageing of the snow/ice interface. This seasonal signature of ice and snow ageing resolves an apparent contradiction in the scientific literature, where radar measurements by Beaven et al. [2] showed very powerful backscatter from the surface of 12 cm thick ice relative to the snow/air interface above, but later studies over thicker, older ice [86], [9], [11], [10] showed a more complex partitioning of returned power between the ice and the overlying snow.

This phenomenon could be better investigated with a deeper snowpack, perhaps with a multi-week experiment in a sea ice tank facility where sufficiently high snowfall over young ice would lead to these interfaces being resolved without the confounding influence of sea ice flooding, since sea ice is known to resist flooding in these facilities by adhering to the tank walls (e.g. [56]).

Both our in-situ and satellite data indicate a Ku- and Ka-band decline in σ^0 in the early stages of ice growth. This suggests that the two frequencies may be backscattering from the same surfaces to at least some degree. The alternative explanation (in the case that the frequencies are backscattering from the snow/ice and snow/air interfaces exclusively), is that these two surfaces display a remarkably similar seasonal evolution of σ^0 . The degree to which these frequencies behave similarly or differently will affect the success of the planned CRISTAL radar altimetry mission [71], which aims to use the Ku- and Ka-bands to distinguish the sea ice and snow surfaces respectively.

OPEN RESEARCH

The KuKa data used in this study can be found at the Canadian Watershed Information Network (DOI: <https://doi.org/10.34992/pqvf-q516>). Analysis code and auxiliary data can be found at https://github.com/robbiemallett/Interpreting_NRCS. The CryoSat-2 data can be found at <ftp://science-pds.cryoSat.esa.int>, and the AltiKa data at <ftp-access.aviso.altimetry.fr>.

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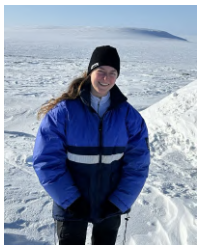
data. Her work includes studies at a range of scales from microstructure, surface-based, airborne, and satellite studies to explore how physical snow characteristics affect radar scattering. She has also led a number of equality, diversity, and inclusion initiatives in science.

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Jeremy Wilkinson is an expert on sea ice dynamics, thermodynamics and mechanics, ocean wave propagation through sea ice, and deep-convection and water mass modification. He represents the UK on the Arctic Ocean Science Board and the International Arctic Science Committee's (ISAC) Marine Science Working Group and is a member of the Programme Advisory Board for Arctic Science for the UK funding agency NERC.

Dr Wilkinson uses a broad range of techniques, from the remote sensing and in-situ monitoring of sea ice, through to its visualisation through the use of upward-looking sonars, and the modelling of sea ice and the flow of oil spilled under sea ice. He has participated in more than twenty polar field campaigns, and has experience in large multidisciplinary programmes e.g. ESOP-1, ESOP-2, SITHOS, CONVECTION, and DAMOCLES.

Dr Wilkinson has also been awarded a number of major grants: These include the EU FP7 Arctic Climate Change, Economy and Society project (ACCESS: 2011-15), the European Science Foundation funded programme Arctic Ocean Sea-ice and Ocean Circulation Changes Using Satellite Methods: (SATICE: 2010-13), the NERC funded Sea ice Processes and Mass Balance in the Bellingshausen Sea (IceBell: 2010-13), the ONR funded MIZ programme (2012-15). Jeremy was coordinator of the €12 million FP7 programme ICE-ARC (2014-17), and is currently coordinator of the NERC DEFIANT project.



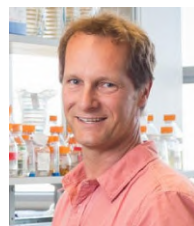
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