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Sentinel-1 Detection of Ice Slabs on the Greenland Ice Sheet

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Abstract. Ice slabs are multi-meter thick layers of refrozen ice that limit meltwater storage in firn, leading to enhanced surface runoff and ice sheet mass loss. To date, ice slabs have largely been mapped using airborne ice-penetrating radar, which has limited spatial and temporal coverage. This makes it difficult to fully assess the current extent and continuity of ice slabs or to validate predictive models of ice slab evolution that are key to understanding their impact on Greenland's surface mass balance. Here, for the first time, we map the extent of ice slabs and similar superimposed ice facies across the entire Greenland Ice Sheet at 500 m resolution using dual-polarization Sentinel-1 (S-1) synthetic aperture radar data collected in winter 2016-2017. The S-1 inferred ice slab extent is in excellent agreement with ice-penetrating radar ice slab detections from spring 2017, as well as the extent of the visible runoff zone as mapped from optical imagery. Our results show that ice slabs are nearly continuous around the entire margin of the ice sheet, including regions in Southwest Greenland where ice slabs have not been previously identified. The algorithm developed here also lays the groundwork for long-term monitoring of ice slab expansion with current and future C-band satellite systems and highlights the added value of future L-band missions for near-surface studies in Greenland.

1 Introduction

Over the last two decades, more than half of mass loss from the Greenland Ice Sheet (GrIS) has come from the runoff of surface meltwater (Van Den Broeke et al., 2009; Enderlin et al., 2014; Mouginot et al., 2019), and surface processes are projected to remain the dominant contributor to Greenland's sea level contribution over the next century (Fox-Kemper et al., 2021). By extension, much of the uncertainty in future mass loss from the ice sheet can also be ascribed to uncertainty in surface processes (Fox-Kemper et al., 2021). One process that remains poorly constrained is the development and expansion of ice slabs in firn near the equilibrium line. Ice slabs are multi-meter thick layers of refrozen ice that form just below the surface (Machguth et al., 2016) and can be horizontally continuous over tens of kilometers (MacFerrin et al., 2019). As a result, ice slabs are largely impermeable and limit the vertical percolation of meltwater into the underlying relict firn, leading to a rapid transition from retention to runoff as they form (Machguth et al., 2016; MacFerrin et al., 2019; Tedstone and Machguth, 2022). To date, ice slabs have primarily been mapped using Operation IceBridge (OIB) airborne ice-penetrating radar surveys, as these data directly resolve the vertical structure of the subsurface and can distinguish homogeneous refrozen ice bodies from

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lower density firn (MacFerrin et al., 2019; Jullien et al., 2023). These data have shown that ice slabs dominate the wet snow zone along the western, northern, and northeastern coasts of Greenland. The southeast basin is the only major region where no ice slabs have been detected, due to the high snow accumulation rate that insulates subsurface liquid water from refreezing and preferences the formation of perennial firn aquifers (Forster et al., 2014; Munneke et al., 2014).

While the OIB data have provided critical insights into ice slab extent across the GrIS, these data are significantly limited in both space and time. Data are only available directly beneath the aircraft track, and collection was limited to a moderate number of flight lines in spring (typically April or May) each year from 2011-2014 and 2017-2018, along with a few additional flights over the wet snow zone in 2010. These gaps in coverage lead to a number of issues. In many regions, the upper limit of the ice slabs is poorly defined, due to a lack of flights perpendicular to the coastline, and there are some areas, most notably in southern Greenland and on peripheral ice caps, where there is insufficient flight coverage to assess whether ice slabs are even present. Even in regions of good coverage, there are typically 5-20 km gaps between flight lines. As a result, the full extent of ice slabs on the GrIS remains poorly defined and it has been difficult to fully assess the km-scale continuity of this facie. Additionally, there are very few repeated flights that were flown perpendicular to the coastline, which are required to robustly assess the inland expansion of ice slabs from year to year. Jullien et al. (2023) showed that some growth occurred between the period from 2010-2012 to 2017-2018, but the spatial resolution of that analysis was coarse and limited by the need to aggregate multiple years of data to achieve reasonable coverage of the whole ice sheet. With the end of the OIB mission in 2019, there is no new ice-penetrating radar data to improve these time series or assess the impact of more recent heavy melt seasons, which included the first high elevation rain event, such as 2019, 2021, and 2023 (Tedesco and Fettweis, 2020; Harper et al., 2023; Box et al., 2023).

These spatial and temporal gaps significantly impede our ability to assess the impact of ice slab development and expansion on the current and future mass balance of the GrIS. For example, MacFerrin et al. (2019) parameterized ice slab extent as a function of the ten-year running mean of local excess melt and applied this parameterization to an ensemble of regional climate models to predict that ice slab expansion would add 7-74 mm of additional sea level rise by 2100. However, this excess melt threshold was tuned by matching the modeled ice slab extent to the aggregate observed extent from 2010-2014 (MacFerrin et al., 2019). As a result, it remains unclear whether the temporal evolution of ice slabs in this model accurately captures the true pace of ice slab growth. As firn models continue to improve, there are many opportunities to implement more physics-based estimates of ice slab expansion and runoff contributions, but in the absence of validating data, significant uncertainties in future projections will remain.

The only clear mechanism for mapping ice slab extent across the entire ice sheet at high resolution ($\sim$1 km or better) on an annual or better basis is to use satellite microwave remote sensing systems. In fact, ice slabs have been mapped from space using the L-band radiometer onboard the Soil Moisture Active Passive (SMAP) mission by Miller et al. (2022a). However, there are significant limitations to this approach. In particular, the instrument resolution is approximately 30 km (Miller et al., 2022a), making it difficult to clearly define the inland extent of the ice slabs or capture expansion on the order of a few kilometers or less per a year. Additionally, although rough estimates of the interannual variability are given, this algorithm aggregates $\sim$5 years of radar data to create a single estimate of ice slab extent (Miller et al., 2022a), which limits its use for generating long

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time series. There are also notable discrepancies between the SMAP and OIB ice slab extents, particularly in the Northwest where SMAP fails to detect large swaths of the OIB-detected ice slabs, and in the North and Northeast where SMAP places the ice slabs at higher elevations than the OIB data (see Figure 11).

An alternate approach is to use active synthetic aperture radar systems such as the European Space Agency's (ESA) Sentinel-1 (S-1) series satellites. Since C-band radio waves penetrate roughly 5-15 meters into snow, firn, and ice, depending on the local physical and dieletric properties (Rignot et al., 2001; Hoen, 2001; Fischer et al., 2019), the depth-integrated surface echo measured by the instrument contains information about the near-surface structure. In Extra Wide Swath mode, Sentinel-1 covers the entire GrIS approximately every 10 days with a spatial resolution of 20 x 40 meters and a full catalog of data available from late 2014 to the present day. With the anticipated launches of Sentinel-1C & D, the data record is projected to continue uninterrupted through at least the early 2030s. Therefore, Sentinel-1 could not only provide the first pan-Greenland mapping of ice slabs, but such an algorithm would open the door to long-term monitoring of ice slab expansion, potentially covering close to two decades of observations. Here, we develop an algorithm to map refrozen ice facies on the Greenland Ice Sheet using dual polarization Extra Wide swath Sentinel-1 measurements of radar backscatter in conjunction with calibration data from ice-penetrating radar observations.

2 Electromagnetic Interactions in Firn

On ice sheets, mean firn density increases exponentially with depth as it compacts under its own weight (Bader, 1954; Herron and Langway, 1980). In the percolation zone, the structure is further modified by the infiltration and refreezing of surface meltwater that forms ice lenses and ice pipes (Benson, 1962). Ice lenses are horizontal sheets of refrozen solid ice that may be up to a few tens of cms thick and extend laterally for a few meters (Benson, 1962; MacFerrin et al., 2019), while ice pipes are vertical refrozen conduits that represent preferential infiltration pathways connecting these ice lenses (Marsh and Woo, 1984; Pfeffer and Humphrey, 1998; Humphrey et al., 2012). The proportion of the firn occupied by these refreeze features generally increases with decreasing elevation and increasing melt-to-accumulation ratio (Harper et al., 2012; Machguth et al., 2016). In the extreme, consistent excess melting may anneal these ice lenses together into multi-meter thick ice slabs that form in the wet snow zone (MacFerrin et al., 2019; Machguth et al., 2016). The wet snow facies eventually transition to the ablation zone via a region of superimposed ice facies, where the near-surface ice is formed by refreezing within the annual accumulation (Benson, 1962). At the lowest elevations, where annual melting consistently exceeds accumulation, the ice sheet transitions to the bare ice ablation zone composed of homogeneous meteoric ice that is exposed at the surface via horizontal advection and ablation.

These near-surface structural variations with elevation lead to commensurate changes in the dominant electromagnetic scattering mechanisms. In the percolation zone, radar echoes are thought to be dominated by volume scattering from embedded ice features on the scale of a few wavelengths (Fahnestock et al., 1993; Jezek et al., 1994; Rignot, 1995; Baumgartner et al., 1999; Langley et al., 2009), making the GrIS percolation zone one of the most radio bright regions on Earth (Swift et al., 1985; Rignot et al., 1993; Jezek et al., 1994). Past work has successfully modeled the observed percolation zone backscatter at C-band as volume scattering from randomly oriented cylinders (Rignot, 1995). This volume scattering dominated regime

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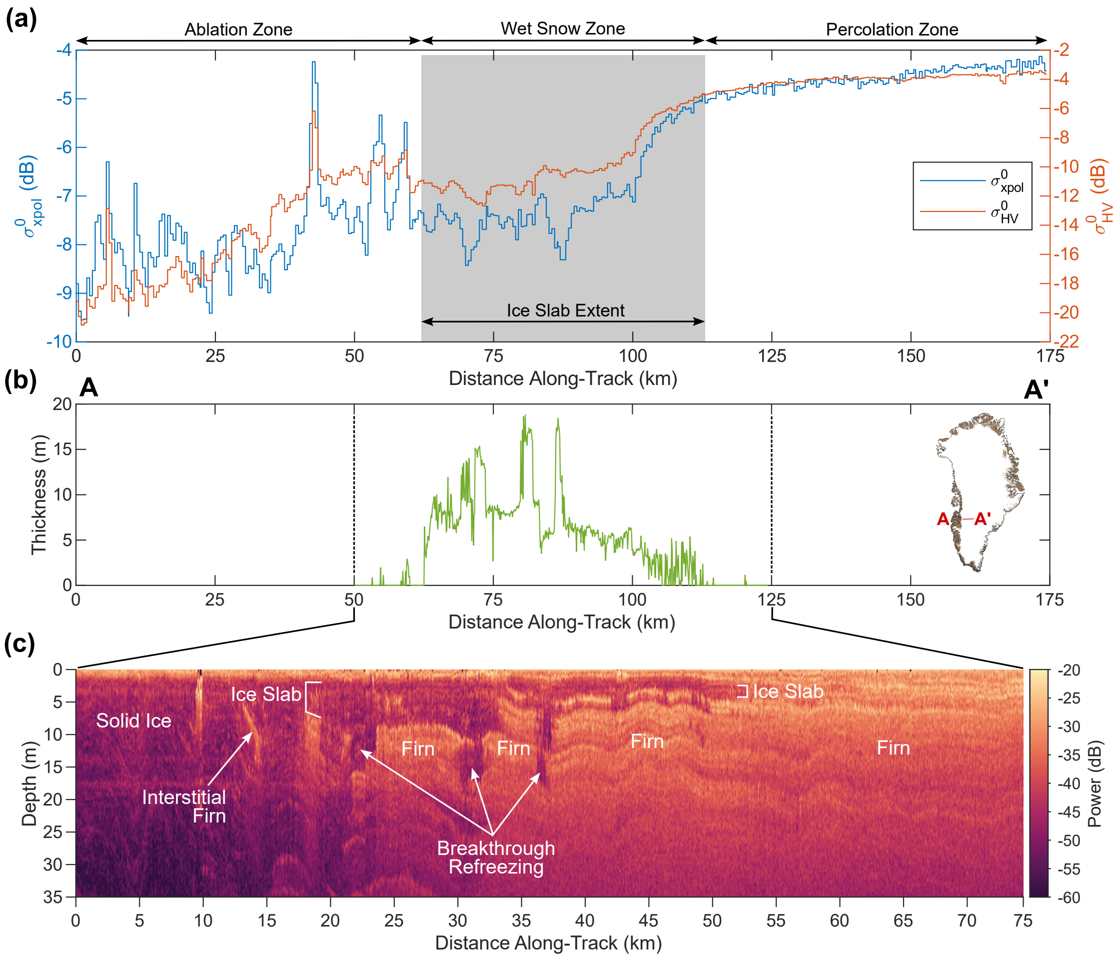

Figure 1. Radar signatures of ice slabs along a transect in Southwest Greenland. (Contains modified Copernicus Sentinel data 2016-2017, processed by ESA.) a) Sentinel-1 σ^0_{HV} is shown in red and the cross-pol backscatter ratio, $\sigma^0_{xpol} = \sigma^0_{HV} - \sigma^0_{HH}$ is shown in blue. The gray region denotes where ice slabs have been detected with ice-penetrating radar (Culberg et al., 2022b). b) Ice slab thickness along the transect as measured with ice-penetrating radar (Culberg et al., 2022b). There is a rapid, down-flow decrease in σ^0_{xpol} as the ice slab thickens, with the backscatter saturating once the ice slab reaches a thickness of around 7 m. The inset map (Gerrish, 2020; Morlighem et al., 2017) shows the location of this transect in Southwest Greenland. c) Radargram from April 2015 collected by the Ultrawideband MCoRDS system (Paden et al., 2014a) showing the subsurface structure in the region where ice slabs have been detected. In the percolation zone, the structure is dominated by layered firn with strong scattering from small embedded ice features. In the wet snow zone, a thick layer of homogeneous refrozen ice with low backscatter overlies relict firn. In the ablation zone, only solid ice remains and there is relatively low backscatter at all depths due to the absence of density constrasts in the subsurface.

also leads to significant depolarization of the incident wave and a large radar cross-section in the cross-polarized (HV or VH) channels (Jezek et al., 1993; Rignot, 1995; Langley et al., 2007; Barzycka et al., 2019). By contrast, scattering in the bare ice

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ablation zone is dominated by rough surface scattering at the air-ice interface, with relatively little volume scattering since heterogeneities such as air bubbles are significantly smaller than the C-band wavelength (Langley et al., 2007, 2009; Barzycka et al., 2019). As a result, the radar cross section of the ablation zone is relatively small and little depolarization occurs, so the echoes are dominated by co-polarized (HH or VV) returns (Langley et al., 2007, 2009). Numerous papers have mapped glacier facies on Arctic ice caps and mountain glaciers based on these characteristic changes in backscatter (Partington, 1998; Long and Drinkwater, 1994; Barzycka et al., 2019). For example, Langley et al. (2008) demonstrated that on Kongsvegen Glacier in Svalbard, the boundaries between firn, superimposed ice, and glacier ice could be mapped in C-Band ENVISAT SAR data from the ∼5 dB change in backscatter between each region, with ground-penetrating radar used to validate the mapping.

Ice slab regions likely represent an intermediate scattering regime between the percolation zone and superimposed ice or ablation zones, with a balance of both surface and volume scattering. Nadir-looking airborne radar sounding measurements show that ice slabs are characterized by strong reflections from their upper and lower interfaces, but very low backscatter within the refrozen ice itself (MacFerrin et al., 2019; Jullien et al., 2023). However, the presence of remnant interstitial firn layers does lead to overall higher backscatter in these refrozen ice facies than in meteoric ice (Figure 1c). Side-looking synthetic aperture radar returns from ice slabs display greater surface scattering and lower volume scattering than the percolation zone, but higher volume scattering than meteoric ice in the lower ablation zone. Figure 1a-b shows an example of this effect along a transect from the ice margin to shallow percolation zone in Southwest Greenland. The percolation zone HV backscatter (σ^0_{HV}) is consistently about -2 dB, but decays at lower elevations as ice slabs begin to form and thicken, eventually reaching a new plateau around -11 dB. The ratio of the HV to HH backscatter ($\sigma^0_{xpol} = \sigma^0_{HV} - \sigma^0_{HH}$ (in dB)), known at the cross-polarized backscatter ratio (Ulaby and Long, 2014) or linear backscatter ratio (Rignot, 1995), has been used as a proxy for the ratio of volume to surface scattering in the Greenland percolation zone (Rignot, 1995) and is also responsive to this change in subsurface structure, decreasing from -4.5 dB to -7.5 dB as ice slabs develop. In this paper, we exploit this reduction in volume scattering that occurs as ice slabs form to map ice slabs from S-1 C-band winter backscatter measurements.

3 Methods

3.1 Sentinel-1 Backscatter Mosaics

For this analysis, we use Extra Wide Swath (EW) ground range detected (GRD) Sentinel-1A & B data collected in HH and HV polarizations at a center frequency of 5.405 GHz (Agency) over the GrIS from 1 October 2016 to 30 April 2017. Only ∼10 days of data are needed to fully cover the entire ice sheet, but we choose to use the full winter period because the extra observations allow us to develop a robust mean backscatter map that reduces the influence of temporal variability in scattering properties, speckle, and variable incidence angles. We expect ice slab extent to be stable during this period since there is no melt infiltration. We only use winter data because the presence of surface meltwater enhances both the surface dielectric contrast and the near-surface attenuation in water-saturated layers, obscuring the subsurface structure. Due to the huge data volume, we process these data in Google Earth Engine (GEE) (Gorelick et al., 2017). Data in the GEE S-1 GRD data collection have undergone thermal noise removal, radiometric calibration, geometric terrain correction, and conversion to dB values in

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the Sentinel-1 Toolbox before being posted to the cloud. Unfortunately, these data have not undergone radiometric terrain correction, and it is impossible to fully implement this algorithm in GEE since it requires access to the data in the original radar coordinates. We experimented with applying an angle-based radiometric terrain correction method designed for GEE (Vollrath et al., 2020), but found that it produced little to no change in the backscatter values due to the extremely low surface slopes on most of the ice sheet. Therefore, we do not implement this correction in our final workflow.

With both Sentinel-1A and Sentinel-1B in operation, the exact repeat interval for any point on the ice sheet is 6 days. However, because the EW swath width is 410 km and Greenland is at high latitudes, the coverage is often more frequent. During our 7 month study period, the average number of observations per pixel was 190, or almost one observation per day, with a minimum of 29 and a maximum of 571 observations. The number of observations is highest in the north and around the margins and lowest in the interior southern saddle. Within each observing pass, the incidence angle varies from 18.9° to 47° across swath (Agency), which creates a significant challenge for generating a consistent backscatter mosaic for the entire ice sheet. Particularly in the percolation zone, backscatter varies strongly with incidence angle, which leads to obvious seams between overlapping swaths and spatial variations in backscatter that are attributable to observation geometry rather than physical properties of the ice sheet. Studies using C- and L-band satellite radar scatterometery data often exploit a linear relationship between incidence angle and backscatter to correct for these incidence angle variations across the swath (Long and Drinkwater, 1994; Ashcraft and Long, 2005; Lindsley and Long, 2016; Long and Miller, 2023). However, the coefficients of this linear fit vary with region and time (Lindsley and Long, 2016; Long and Miller, 2023). Therefore to correct for the effects on incidence angle in our mosaic, we fit a linear function to incidence angle vs backscatter on a per-pixel basis using all available images in our study period and then use this relationship to calculate the theoretical backscatter at an incidence angle of 35°. Scatterometer studies have typically corrected their data to an incidence angle of 40°, but here we choose a to correct the data to an incidence angle close to the middle of the S-1 scene. We combine ascending and descending orbits from both satellites to maximize the angular diversity in each pixel for the most robust fit and apply a separate empirical linear to correction to the σ^0_{HH} and σ^0_{HV} measurements. In this way, we form a consistent mean winter backscatter image for the entire ice sheet. We then calculate the σ^0_{xpol} map by subtracting the σ^0_{HH} map from the σ^0_{HV} map. Before further analysis, we multi-look each mosaic to 500m square pixels with a boxcar filter and export the data as unsigned 16 bit integers. We also use the BedMachinev3 ice mask to remove pixels in regions without ice (Morlighem et al., 2017).

Figure 2a-b shows the mean winter σ^0_{HV} and σ^0_{xpol} mosaics for Greenland in winter 2016-2017. Regions with ice slabs clearly show greater σ^0_{HV} than the lower ablation zone, but reduced σ^0_{HV} compared with the percolation zone. Similarly, ice slabs show a lower σ^0_{xpol} than the percolation zone.

3.2 Excluding the Dry Snow Zone and Firn Aquifer Regions

In order to reduce false positive detection of ice slabs, we exclude regions of the ice sheet that a) experience little to no melting or b) are already known to host firn aquifers. This step is critical because, as can be seen in Figure 2b, both of these regions exhibit low σ^0_{xpol} values that are on par with what is observed in known ice slab regions. In the dry snow zone, this occurs because the subsurface is dominated by smooth depositional snow layers with little heterogeneity beyond the ice grain scale.

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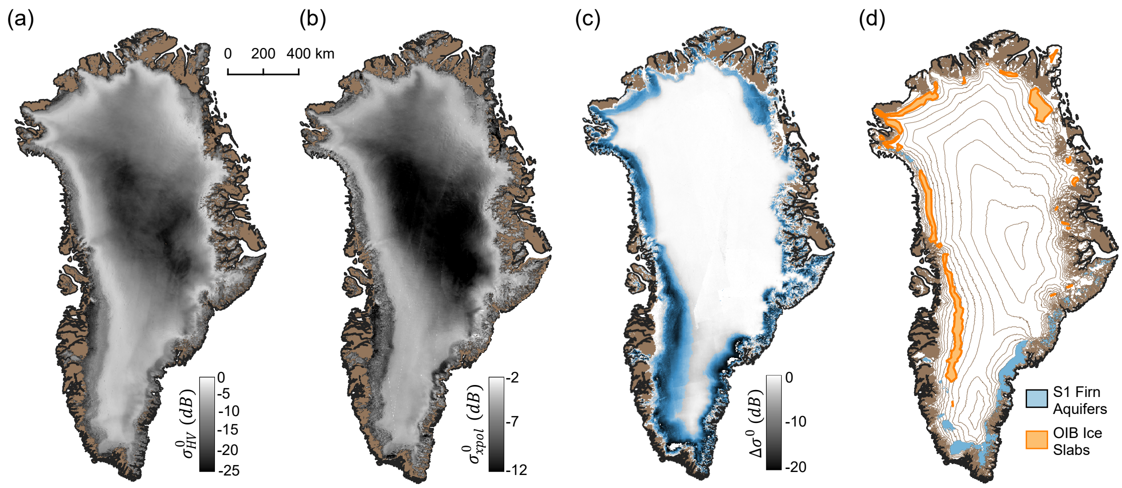

Figure 2. a) Average winter σ^0_{HV} map at 35^o incidence angle covering 1 Oct 2016 - 30 April 2017. b) Average winter σ^0_{xpol} map at 35^o incidence angle covering 1 Oct 2016 - 30 April 2017. c) Difference between summer and winter HH backscatter ($\Delta\sigma^0$), averaged over 1 Nov 2014 - 31 Aug 2020 . We exclude all regions outside the blue overlay from our ice slab analysis, since the minimal change in backscatter between seasons indicates that there is relatively little surface melting in these areas. d) Locations of firn aquifers (blue) detected using Sentinel-1 data from 2014-2019 as published in Brangers et al. (2020). Regions detected as ice slabs with ice-penetrating radar data (Jullien, 2023) are shown in orange for reference. In all panels, the Greenland coastline was produced by the British Antarctic Survey (Gerrish, 2020), the ice mask as part of BedMachine v3 (Morlighem et al., 2017), and the 200 m contours are derived from ArcticDEM (Porter et al., 2018). Panels a-c contain modified Copernicus Sentinel data 2016-2017, originally processed by ESA.

Firn aquifer regions retain liquid meltwater through the winter which leads to increased subsurface absorption and therefore a relatively greater degree of surface scattering, since subsurface volume scattering is suppressed (Brangers et al., 2020; Miller et al., 2022a).

To exclude regions with minimal surface melting, we adapt an existing method for mapping wet snow facies in Greenland based on the change in S-1 σ^0_{HH} between winter and summer (Hu et al., 2022). Much like the classic scatterometer and radiometer algorithms for estimating firn saturation from VV backscatter (Wismann, 2000; Ashcraft and Long, 2006; Hicks and Long, 2011; Miller et al., 2022b), this approach exploits the fact that the enhanced microwave absorption in wet snow leads to a significant reduction in backscatter during the summer when surface melting occurs. We first calculate an average winter σ^0_{HH} map at 35° incidence angle by applying the linear correction method described in Section 3.1 to aggregated data from 1 Nov - 31 March each year between 2014 and 2020. We calculate an average summer σ^0_{HH} map at 35° using all observations between 1 July and 31 Aug from 2015-2020 and then calculate the difference between the summer and winter backscatter as $\Delta\sigma^0 = \sigma^0_{summer} - \sigma^0_{winter}$. We aggregate data over these five years because melt extent varies significantly from year to year and from region to region. This extended time series prevents us from inadvertently excluding areas from analysis due to anomalously low melt extent in any given year. We then choose an empirical threshold to discriminate regions

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with consistent surface melting. Hu et al. (2022) derived a threshold of -7 dB to discriminate between wet snow facies and percolation zone in the $\Delta\sigma^0$ image, based on the distribution of backscatter values observed in Northeast Greenland. However, we find that this threshold is overly aggressive when applied to our average $\Delta\sigma^0$ map and excludes some regions in North Greenland where ice slabs have been observed with ice-penetrating radar. Therefore, we use a threshold of $\Delta\sigma^0$ < -4.7 dB, which is the minimum value that produces a melt region mask which encompasses all OIB ice slab observations from spring 2017. This threshold value falls midway between the Hu et al. (2022) threshold of -7 dB for discriminating wet snow facies and the common threshold of -3 dB for discriminating regions of surface melting (Nagler and Rott, 2000; Liang et al., 2021; Li et al., 2023), suggesting that this is a reasonable empirical choice that is consistent with prior work on wet snow mapping with S-1. Figure 2c shows the five-year melt extent mosaic, with the region we consider for ice slab detection ($\Delta\sigma^0$ < -4.7 dB) highlighted in the blue.

To exclude firn aquifer regions, we use the Sentinel-1 firn aquifer map originally published in Brangers et al. (2020). These firn aquifer areas were detected by identifying pixels where the mean April σ^0_{HV} exceeded the mean September σ^0_{HV} by 9.4 dB or more, using mean monthly values aggregated over 2014-2019, similar to our firn saturation map. Figure 2d shows the locations of these firn aquifers in relation to previous OIB ice slab detections.

3.3 Threshold Optimization and Uncertainty Analysis

To map ice slabs with our S-1 mosaics, we optimize independent backscatter thresholds that demarcate the upper and lower limit of the ice slabs. For the upper boundary, we first take the σ^0_{HV} and σ^0_{xpol} mosaics and mask out the dry snow zone and firn aquifer regions. We then define ice slabs as covering the region where $\sigma^0_{HV} < \alpha$ and $\sigma^0_{xpol} < \beta$, where α and β are independent empirical thresholds. We search for the optimal values of α and β that maximize the agreement between the upper limit of the ice slabs as detected by airborne ice-penetrating radar, and the upper limit of the ice slabs as estimated by S-1.

To optimize α and β, we create a training data set using the Jullien et al. (2023) high-end estimate of ice slab extent derived from OIB flight lines surveyed in March-May 2017. For each flight line that passes through an ice slab area, we extract the portion of the flight line that overflies the ice slabs, as well as an additional 50 km buffer that extends inland of the upper limit of the ice slabs. We discretize these lines into points every 50 m and assign each point a value of 1 if an ice slab was detected in the OIB data at that location or 0 if no ice slab was detected. Similarly, we binarize our S-1 ice slab detections where a pixel value of 1 means an ice slab was detected and 0 means no ice slab was detected. We then test all combinations of thresholds for -7.12 dB $< \beta <$ -2.37 dB and -13.6 dB $< \alpha <$ -2.1 dB, calculate the F1 score for each combination, and choose the threshold values that give the highest F1 score. The F1 score is a measure of the accuracy of a binary classification and is calculated following Equation 1. Figure 3 shows this optimization trade space with the optimal threshold combination shown in the white dot. We find that using both σ^0_{xpol} and σ^0_{HV} thresholds together leads to modestly better agreement with the OIB detections, compared to using only σ^0_{xpol}. The optimal ice slab extent estimated using only σ^0_{xpol} has an F1 score of 0.787, compared to an F1 score of 0.811 for the joint optimization. The optimal ice slab extent estimated using only σ^0_{HV} has an F1 score of only

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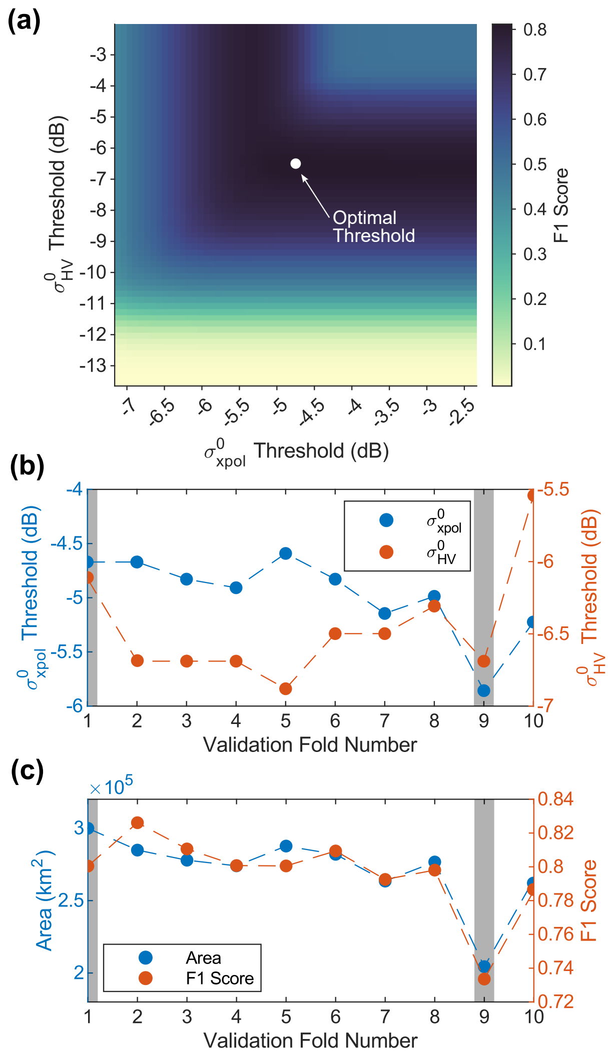

Figure 3. Optimization of the detection thresholds for the upper limit of ice slabs. a) F1 score as a function of σ^0_{xpol} and σ^0_{HV} thresholds. The optimal threshold combination (maximum F1 score) is shown in the white dot. b) Optimal thresholds from each iteration of the ten-fold cross validation scheme. The thresholds that give the minimum and maximum total ice slab extent are marked in the grey bars. We use these two thresholds to quantify uncertainty in the upper limit of the ice slabs. c) The total ice slab area and F1 score on the withheld validation set for each iteration of the ten-fold cross-validation.

0.674, so it is clear that the cross-polarized backscatter ratio provides significant additional information.

$$\mathrm{F1} = \frac{2 * \text{true positive}}{2 * \text{true positive} + \text{false positive} + \text{false negative}} \quad (1)$$

To quantify uncertainty in our detection of the upper limit of the ice slabs, we use a ten-fold cross-validation scheme. We divide our training dataset into ten even subsets, rerun our optimization routine on each data subset, and calculate the F1 score

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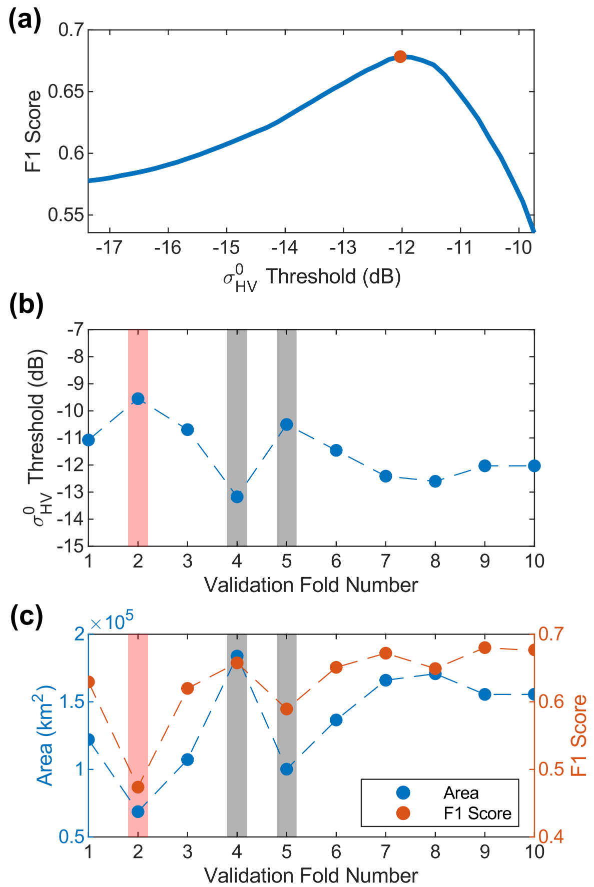

Figure 4. Optimization of the detection threshold, ϕ, for the lower limit of ice slabs. a) F1 score as a function of the σ^0_{HV} threshold. The optimal threshold (maximum F1 score) is shown in the red dot. b) Optimal thresholds from each iteration of the ten-fold cross validation scheme. The thresholds that give the maximum and minimum total ice slab extent are marked in the grey bars. We use these two thresholds to quantify uncertainty in the lower limit of the ice slabs. c) The total ice slab area and F1 score on the withheld validation set for each iteration of the ten-fold cross-validation. We discard the iteration marked with the red bar to due to the anomalously poor F1 score and choose the maximum and minimum ice slab extents from the remaining nine iterations (marked in the two gray bars).

on the 90% of the data that was withheld. This yields ten separate estimates of the optimal backscatter thresholds, tuned to different regions of the ice sheet. We take the thresholds that yield the maximum and minimum total ice slab area and use these limits to bound the plausible extent of the ice slabs as estimated by S-1. Figure 3c-d shows the result of this analysis.

Next we separately optimize the threshold for detecting the lower boundary of the ice slabs using the same method as described above, but now using a training dataset that covers the ice slab region and a 50 km buffer down-flow into the ablation zone. Initial analysis of the backscatter mosaics suggests that σ^0_{xpol} does not display any distinct change in behavior associated

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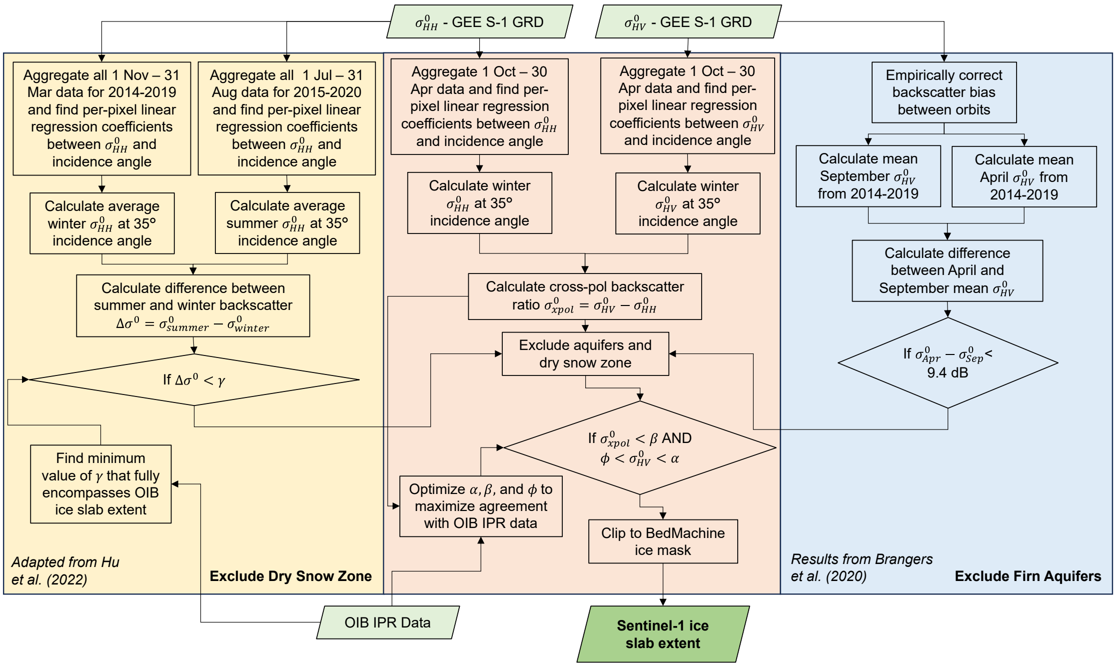

Figure 5. Sentinel-1 ice slab detection algorithm flowchart.

with the lower boundary (see Figure 1a), so we optimize a single threshold, $\sigma^0_{HV} > \phi$, with uncertainty determined using the same ten-fold cross-validation scheme described above.

Finally, we use our optimized thresholds to map the minimum, most likely, and maximum extent of ice slabs from the S-1 data following the complete workflow shown in Figure 5. To quantify the accuracy of these mapped extents, we calculate the confusion matrix, F1 score, and Cohen's κ for each limit.

4 Results and Discussion

4.1 Sentinel-1 Map of Ice Slab Extent

Figure 6 shows the S-1 estimated ice slab extent in winter 2016-2017, compared with the OIB ice slab detections. We find excellent agreement between the upper limit of the ice slabs as identified by OIB and the S-1 estimated upper limit. Figure 7 shows the confusion matrices, F1 scores, and Cohen's κ for the minimum, most likely, and maximum S-1 estimated ice slab extent that quantify this agreement. The most likely ice slab extent has an F1 score of 0.811 with a true positive rate of 94% when detecting the upper limit of the ice slabs. The S-1 estimates at 500 m are also able to capture much of the km-scale variability along the upper limit, including regions of discontinuous ice slabs (see Figure 6f for an example). The fingering

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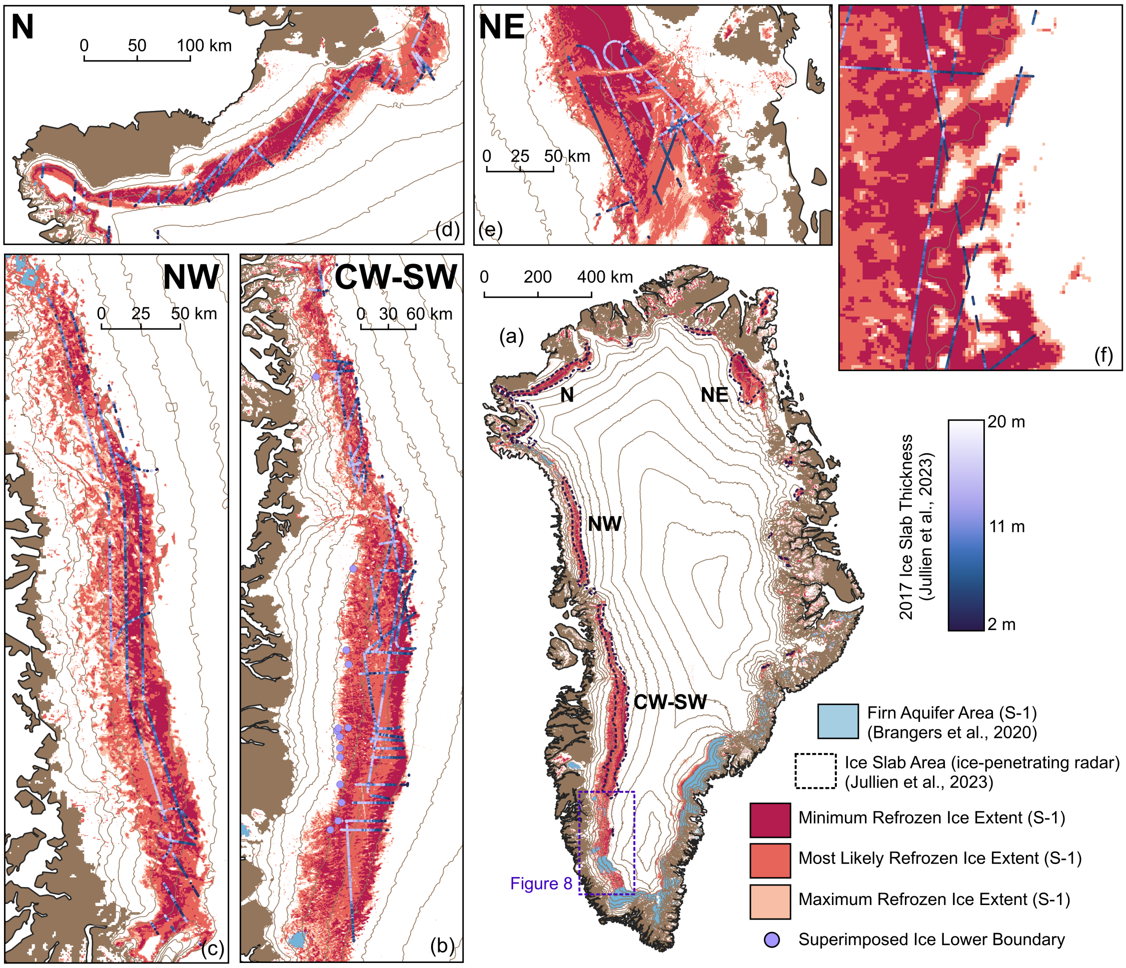

Figure 6. Sentinel-1 mapping of ice slabs in winter 2016-2017. a) S-1 detected ice slabs are shown in red, with the outline of the OIB detected ice slabs in the dashed black line (Jullien, 2023). We find overall excellent agreement between the S-1 and OIB mapping, although S-1 detects significant additional ice slab area in Southwest Greenland, along the Central East margin, and on peripheral ice caps. b) Zoom-in of the Central and Southwest regions. OIB ice slab detections are overlaid in the purple dots (Jullien, 2023), where darker colors indicated thinner ice slabs. There is a significant gap between the lower limit of the OIB ice slab detections and the lower limit of the S-1 mapping. The lower limit from S-1 is better aligned with the lower limit of superimposed ice as mapped from ice-penetrating radar in this paper (large purple dots). c) Zoom-in of Northwest region. d) Zoom-in of Northern region. e) Zoom-in of Northeast region. f) Zoom-in from Southwest Greenland showing details of the upper boundary. We find excellent agreement between the OIB and S-1 detections even where ice slabs are discontinuous due to preferential expansion in topographic lows. In all panels, the Greenland coastline was produced by the British Antarctic Survey (Gerrish, 2020), the ice mask as part of BedMachine v3 (Morlighem et al., 2017), and the 200 m contours are derived from ArcticDEM (Porter et al., 2018).

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Upper Boundary

Minimum Extent

OIB \ Sentinel-1	Yes	No
Yes	60,090 (True Positive Rate: **72%**)	23,792 (False Negative Rate: **28%**)
No	17,872 (False Positive Rate: **8%**)	208,542 (True Negative Rate: **92%**)

F1 Score: 0.742
κ: 0.652

Most Likely Extent

OIB \ Sentinel-1	Yes	No
Yes	78,705 (True Positive Rate: **94%**)	5,177 (False Negative Rate: **6%**)
No	31,411 (False Positive Rate: **14%**)	195,003 (True Negative Rate: **86%**)

F1 Score: 0.811
κ: 0.727

Maximum Extent

OIB \ Sentinel-1	Yes	No
Yes	80,396 (True Positive Rate: **96%**)	3,486 (False Negative Rate: **4%**)
No	36,395 (False Positive Rate: **16%**)	190,019 (True Negative Rate: **84%**)

F1 Score: 0.801
κ: 0.710

Lower Boundary

Minimum Extent

OIB \ Sentinel-1	Yes	No
Yes	38,035 (True Positive Rate: **61%**)	24,159 (False Negative Rate: **39%**)
No	24,454 (False Positive Rate: **16%**)	130,447 (True Negative Rate: **84%**)

F1 Score: 0.610
κ: 0.453

Most Likely Extent

OIB \ Sentinel-1	Yes	No
Yes	58,747 (True Positive Rate: **94%**)	3,447 (False Negative Rate: **6%**)
No	53,211 (False Positive Rate: **34%**)	101,690 (True Negative Rate: **66%**)

F1 Score: 0.674
κ: 0.485

Maximum Extent

OIB \ Sentinel-1	Yes	No
Yes	61,741 (True Positive Rate: **99%**)	453 (False Negative Rate: **1%**)
No	66,979 (False Positive Rate: **43%**)	87,922 (True Negative Rate: **57%**)

F1 Score: 0.647
κ: 0.424

Figure 7. Confusion matrices quantifying the agreement between the OIB and S-1 ice slab detections for the minimum, maximum, and most likely ice slab extents. We quantify the fit for the upper boundary (top row) and lower boundary (bottom row) separately, since these thresholds were optimized separately. The most likely extent does an excellent job of detecting the upper limit of the ice slabs, with an F1 score of 0.811 and Cohen's κ of 0.727, but the lower boundary is much more uncertain, with a best F1 score of only 0.674 and Cohen's κ of 0.485, likely due to the consistent overestimation of ice slab extent in Southwest Greenland.

structures that we map in many regions are consistent with preferential expansion through topographic lows where water collects as it flows laterally through saturated firn layers.

Our mapping identifies new ice slab regions in Southwest Greenland that have not been previously classified as such, likely due to a lack of comprehensive airborne radar coverage in this region. These newly-identified ice slab areas are highly consistent with the extent of the visible runoff zone mapped from Landsat imagery in Tedstone and Machguth (2022), confirming that vertical percolation is limited in these areas (Figure 8). However, the S-1 ice slab extent is often patchy and discontinuous in this region, likely due to the high prevalence of buried surface lakes and isolated aquifer regions that limit detection of ice slabs due to the presence of liquid water in the subsurface.

There are also a number of discrepancies between the OIB and S-1 mapping. In the northwest, S-1 appears to slightly underestimate the upper elevation of the ice slabs, particularly where they transition to firn aquifers, and the ice slab extent is fairly discontinuous in this region. The S-1 algorithm generally fails to detect ice slabs in basins with persistent buried supraglacial lakes because surface scattering from the water table dominates the return, likely contributing to this discontinuous mapping in the Northwest where buried lakes are common (Koenig et al., 2015). In the Northeast, the S-1 algorithm fails to detect gaps in the ice slabs in the shear margins of the Northeast Greenland Ice Stream that are present in the OIB data. This

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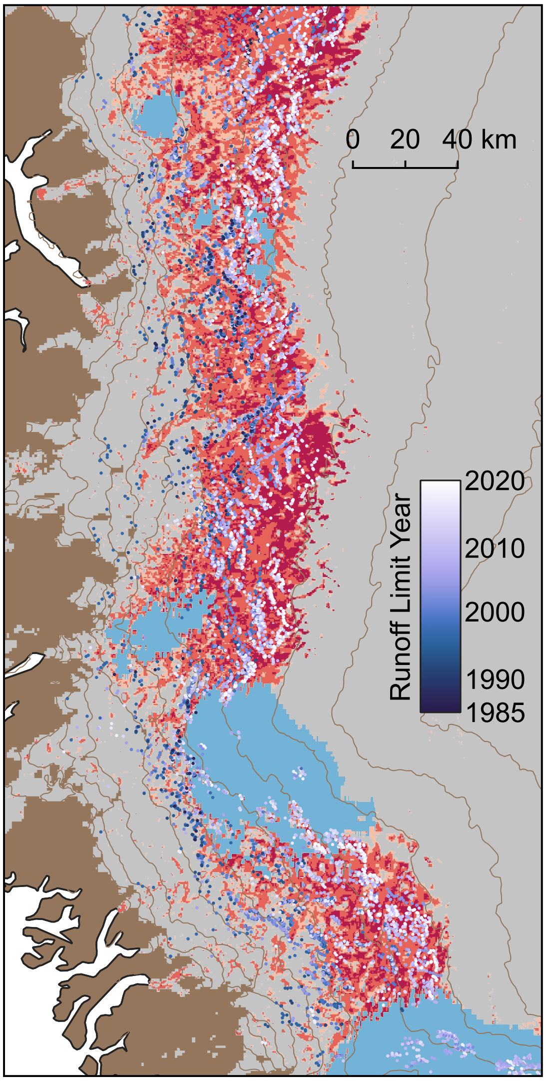

Figure 8. Comparison of the maximum visible runoff line from 1985-2020 with the newly mapped ice slabs regions in Southwest Greenland. The ice slab regions are marked with the same orange and red color scheme as Figure 6 and firn aquifers are shown in light blue (Brangers et al., 2020). Points marking the visible runoff limit in each sector, color-coded by year of observation, are overlaid in purple (Tedstone, 2022). There is a clear correspondence between the newly mapped ice slab regions and the runoff limit, confirming that vertical percolation is limited in these areas. The Greenland coastline was produced by the British Antarctic Survey (Gerrish, 2020), the ice mask as part of BedMachine v3 (Morlighem et al., 2017), and the 200 m contours are derived from ArcticDEM (Porter et al., 2018).

highlights that regions with significant surface crevassing are challenging for both OIB and S-1 detection of ice slabs. S-1 will tend to overestimate ice slab extent in crevassed regions, due to enhanced σ^0_{HV} that our algorithm ascribes to volume scattering from firn, but actually results from rough surface and multi-bounce effects within the crevasses. On the other hand,

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surface crevasse clutter in the OIB data can prevent definitive classification of the near-surface structure, particularly when using radiometric metrics that assume relatively homogeneous planar structures. The S-1 algorithm also fails to detect some isolated ice slab segments identified at anomalously high elevations in the OIB data in the North and Northwest. Manual review of the radargrams in these areas shows that most fall in high melt, high accumulation areas where a thick layer of relatively transparent winter snow overlying a strong reflector at the previous summer surface may have been misclassified as an ice slab.

We also consistently map ice slabs along the upper boundaries of firn aquifer, both in the Northwest and Southeast, that are not identified in the OIB data. It is possible that these areas represent aquifer regions with low volumetric water content where the seasonal backscatter variability does not meet the threshold for aquifer detection, but surface scattering is still enhanced by partial winter meltwater retention. Time series of σ^0_{HV} from these aquifer-marginal areas in the Southeast show an intermediate scattering regime, with slower backscatter recovery than the percolation zone, but more rapid recovery than the well-defined aquifer regions. Alternately, there is ice-penetrating radar evidence for near-surface refreezing in continuous ice layers less than 1 m thick following both the 2012 and 2015 melt seasons (Culberg et al., 2021; Miller et al., 2022a) that extend to the upper limit of the southeastern firn aquifers. Similar shallow ice layers might also contribute to enhanced surface scattering and lead to erroneous ice slab detections in the Southeast.

Overall, we estimate a most likely ice slab extent of 104,020 km^2, compared to previous estimates of 60,400 - 73,500 km^2 from OIB data processed in Jullien et al. (2023) and 76,000 km^2 from SMAP data processed in Miller et al. (2022a). Much of this additional area comes from the newly detected regions in Southwest Greenland, as well as smaller contributions from narrow regions along the periphery, peripheral ice caps including Flade Isblink, and some misclassified regions at lower elevations and in fast-flowing glacier tongues in the mountainous eastern basins. Difficulty in accurately mapping the lower boundary of the ice slabs, further discussed in Section 4.2, also adds to the discrepancy in total extent.

4.2 Uncertainty in the Lower Boundary of Ice Slabs

Mapping the lower elevation limit of ice slabs is significantly more challenging than mapping the upper limit, as evidenced by the large uncertainty and apparently poor fit with the OIB detections. Our best estimate of the lower limit of the ice slabs has an F1 score of 0.674, compared to 0.811 for the upper boundary. There are two major sources of uncertainty which may contribute to this poor fit. First, it is likely that the limited penetration depth of S-1 prevents a clear delineation between regions where ice slabs are simply thicker than the system depth sensitivity and regions with a solid ice column. Figure 9 shows two-dimensional histograms of S-1 backscatter versus OIB-detected ice slab thickness. Both σ^0_{HV} and σ^0_{xpol} show little to no relationship with ice slab thickness beyond ∼7 m, suggesting that S-1 is largely insensitive to scattering structure below that depth. Since well-developed ice slabs in regions such as Southwest Greenland are often 8-10 m thick, it is unsurprising that S-1 struggles to clearly detect the transition from ice slabs to solid ice. Second, the lower limit of the ice slabs in the airborne ice-penetrating radar dataset is not actually a data-driven boundary. Jullien et al. (2023) used the RACMOv2.3p regional climate model to exclude any regions below the long-term equilibrium line from their analysis, so the lower boundary is actually set by the model results. Given the simple snow model coupled to RACMO, the model may not accurately capture the true extent of ice slabs. Additionally, it almost certainly does not capture complex regions where near-surface ice may have been formed by

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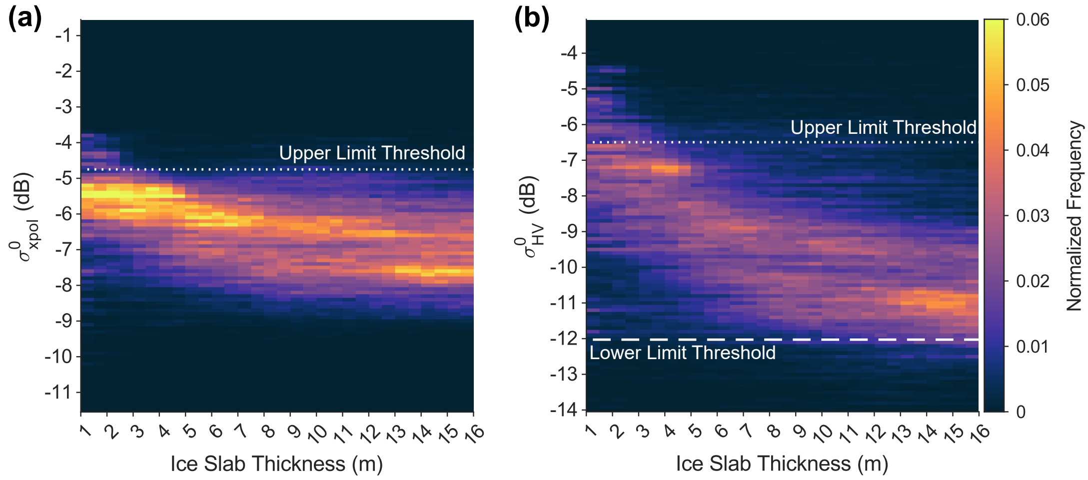

Figure 9. Sentinel-1 backscatter sensitivity to subsurface structure. a) Normalized two-dimensional histogram of ice slab thickness from ice-penetrating radar versus Sentinel-1 σ^0_{xpol}. b) Normalized two-dimensional histogram of ice slab thickness from ice-penetrating radar versus Sentinel-1 σ^0_{HV}. In both cases, the change in backscatter saturates around an ice slab thickness of $\sim$ 7 m, suggesting that the S-1 penetration depth is limited to approximately that depth. The optimal thresholds for the upper and lower limit of the ice slabs are shown in dashed white lines on each plot. This figure also demonstrates that the σ^0_{xpol} metric improves detection of the ice slab upper limit because the spread of backscatter values that map to an ice slab thickness of 1-2 m is significantly reduced compared to σ^0_{HV}.

refreezing, even if firn is no longer present. This includes historical ice slab regions, superimposed ice regions, where meltwater fully saturates the annual accumulation and refreezes to form surface ice layers (Benson, 1962), areas where the firn column may have been completely filled by surface meltwater draining through surface crevasses (Culberg et al., 2022a), or regions where refrozen ice was advected in from higher elevations.

We hypothesize that any ice formed by refreezing induces significant volume scattering due to trapped air bubbles, interstitial firn pockets, and other heterogeneities in density, leading to a σ^0_{HV} signature that is more similar to ice slabs than meteoric ice. This is consistent with previous work which showed clear differences in C-band polarimetric backscatter between glacier ice, superimposed ice, and firn regions (Langley et al., 2008, 2009; Barzycka et al., 2019). To test this hypothesis, we reanalyze 14 airborne radar data flights from 2017 in Central West and Southwest Greenland that are approximately parallel to ice flow. Both the IMAU Firn Densification model (Brils et al., 2022) and the maximum depth of ice blobs observed in the Jakobshavn catchment (Culberg et al., 2022a) suggest that pore close-off occurs at around 30 m depth in this region. Therefore, in each radargram, we identify an englacial layer that is approximately 30 m below the surface near the upper limit of the ice slabs and assume it represents the bottom of the firn column. We trace this layer downstream until it outcrops at the surface due to ablation. Where surface sidelobes obscure the radiostratigraphy or there are significant stratigraphic disturbances near the surface, we estimate the maximum outcropping elevation as the last point where the layer can be clearly traced, and the minimum elevation as the point where we would extrapolate the layer outcropping to occur if the layer slope remained the

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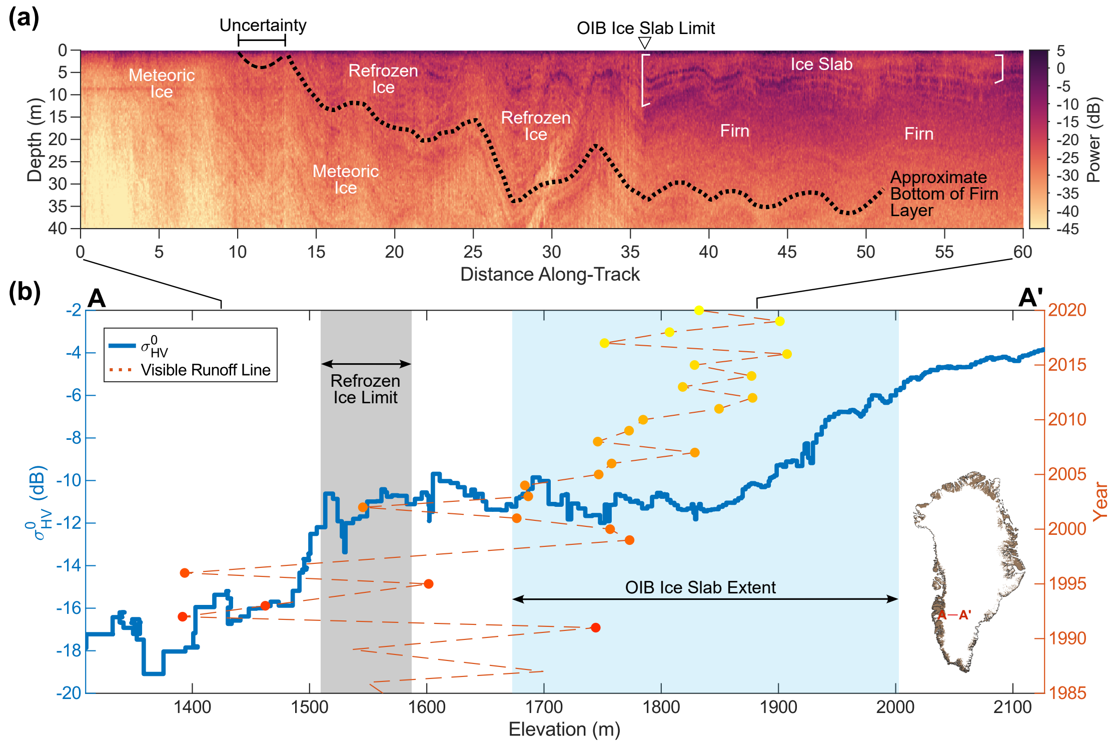

Figure 10. Sentinel-1 detects the lower limit of refrozen ice facies. (Contains modified Copernicus Sentinel data 2016-2017, processed by ESA.) a) Accumulation Radar transect from April 2017 (Paden et al., 2014b) showing the inferred transition from ice slabs, to superimposed ice facies, to solid meteoric ice. The dashed black line shows the englacial layer that we trace from the bottom of the firn until it outcrops at the surface in order to define the lower limit of the refrozen ice facies. b) Comparison of σ^0_{HV} (blue line) as a function of elevation with the OIB ice slab extent (blue patch) (Jullien, 2023), estimated lower boundary of superimposed ice facies (grey patch, this paper), and the elevation of the visible runoff line between 1985 and 2020 (dashed red line with dots at annual measurement points) (Tedstone, 2022). The region where we infer that surface ice was formed by refreezing is marked by a plateau in σ^0_{HV} around -11 dB and is also the region over which the visible runoff zone has retreated in the last two decades, supporting the idea that this region may have been near or above the firn-line in the recent past. The inset map in panel b (Gerrish, 2020; Morlighem et al., 2017) shows the location of this transect in Southwest Greenland.

same. Figure 10a shows an example of this layer tracing process. We infer that ice at depths shallower than the traced layer was likely formed by refreezing, rather than compaction.

In Figure 6b, the large purple dots mark the minimum elevations of these outcropping points, showing strong agreement between the S-1 inferred lower boundary of ice slabs and this new OIB-inferred limit of refrozen surface ice facies. This region between the boundary of refrozen ice facies and the lower limit of the OIB-mapped ice slabs corresponds to the area over which the visible runoff line has retreated since the mid-1980s (see Figure 10) (Tedstone and Machguth, 2022), with significant interannual variability in runoff extent. This suggests that the S-1 mapping in part captures the historical equilibrium

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zone, which would have been in positive mass balance prior to the 1980s and may have still experienced intermittent years of positive mass balance into the early 1990s. Given the slow ice flow in the Southwest ($\sim$40 ma^{-1}), this contributes to a wide zone where surface ice consists of historical ice slabs that have not yet fully ablated, further modified by intermittent superimposed ice formation, and ongoing downstream advection of other refrozen ice. Therefore, we infer that our S-1 mapping captures not only ice slabs, but all regions where the near-surface ice was formed predominantly by refreezing.

This conclusion is consistent with some of the regional differences in the mismatch between the S-1 and OIB-inferred lower exent of the ice slabs. In the Southwest, there is a 20-35 km gap between the bottom of the OIB-detected ice slabs and S-1 mapped ice slabs. This is consistent with the low surface slopes, long history of melt, and slow and variable retreat of the snowline and expansion of the visible runoff zone in this region (Ryan et al., 2019; Tedstone and Machguth, 2022). In contrast, the two mappings agree fairly well in the North which has seen consistent expansion of the runoff zone and retreat of the snowline since 1990 (Ryan et al., 2019; Noël et al., 2019), suggesting that the formation of refrozen ice facies in this region is a more recent and rapid phenomenon. However, some of the discrepancies in the lower limit are likely attributable to other complex surface scattering mechanisms rather than an extended superimposed ice zone. For example, in the Northwest, the S-1 lower limit is particularly diffuse, with complicated and disconnected regions identified as potential refrozen ice all the way to the ice sheet margin. We hypothesize that this is due to a propensity for regions of heavy crevasses to be misclassified as refrozen ice, an issue which is more pronounced in the fast-flowing Northwest where surface strain rates are high and crevassing is prevalent.

4.3 Comparison with SMAP-Derived Extent

Figure 11 compares our S-1 derived refrozen ice facie extent with the ice slab extent derived from SMAP in Miller et al. (2022a). Overall, S-1 offers a significant improvement in both accuracy and resolution, particularly capturing regions in Northwest Greenland that SMAP failed to classify as ice slabs and accurately capturing the elevation bands where ice slabs form in the North and Northeast. However, SMAP does a somewhat better job of capturing the lower limit of the ice slabs in Southwest Greenland, in large part because the lower limit of the SMAP-inferred percolation zone (dark purple dashed outline) is much more consistent with MODIS-inferred estimates of the summer snowline (Ryan et al., 2019) than S-1 (lilac region), which maps wet snow well into the ablation zone in some regions. SMAP also maps melt significantly further inland on the ice sheet than S-1, in part due to the comparatively coarse effective resolution, all of which contributes to different areas in which ice slabs are assumed to be viable.

The upcoming launch of the joint NASA-ISRO NISAR mission and eventual launch of ESA's Radar Observing System for Europe-L-Band (ROSE-L) mission will soon provide L-band synthetic aperture radar data over Greenland, which has the potential to offer the best of both these products. The enhanced penetration depth at L-band may particularly enable a better delineation of the low-elevation transition from ice slabs to superimposed ice and enable more robust constraints on ice slab thickness from space. The longer wavelength will also significantly improve interferometric coherence over the ice sheet and potentially enable ice slab mapping based on volume decorrelation (Rizzoli et al., 2017) or other coherence-derived metrics. This will be a particularly important avenue of investigation given NISAR is expected to primarily collect data in

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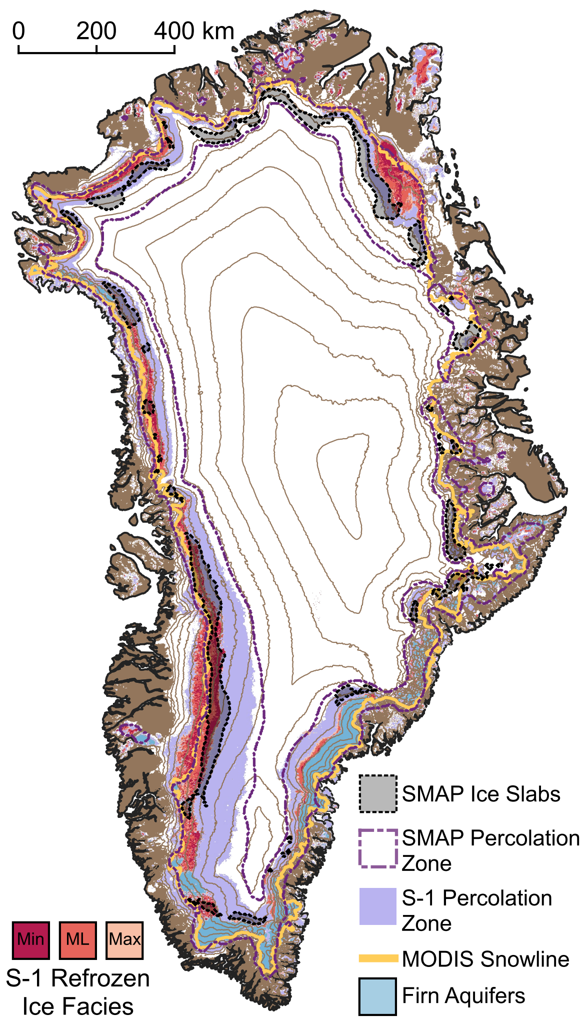

Figure 11. Comparison of S-1 inferred refrozen ice facies in winter 2016-2017 (this paper) and SMAP-inferred average ice slab extent from 2015-2019 (Miller, 2021). S-1 shows a significant improvement in resolution and accuracy over SMAP. However, SMAP is able to better capture the true extent of the percolation zone, and hence the lower limit of the ice slabs, as demonstrated by the better match between the lower limit of the SMAP-derived percolation zone (Miller, 2021) and a MODIS-derived estimate of the average summer snowline (Ryan et al., 2019). Firn aquifers are shown in light blue (Brangers et al., 2020).

single-polarization mode over Greenland. Additionally, NISAR will not collect data above 77.5° north, unfortunately limiting future capacity to study the rapidly changing northern basins. However, where data are collected, these complementary L-band observations have the potential to significantly improve our capacity to study the near-surface of Greenland from space, and our C-band algorithm development will provide an important bridge between the historical OIB data and future L-band data, which will not overlap in time with OIB.

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5 Conclusions

We have shown that Sentinel-1 winter σ^0_{xpol} and σ^0_{HV} signatures can be used to map the extent of Greenland's refrozen ice facies from space at 500 m spatial resolution. Our mapping is in excellent agreement with both subsurface observations from the OIB ice-penetrating radar data and remote sensing observations of visible surface runoff. We identify new ice slab regions in Southwest Greenland and our mapping suggests that ice slabs are largely ubiquitous in the wet snow zone in all regions besides Southeast Greenland. Given the radiometric stability and consistent calibration efforts for Sentinel-1, we expect that it may be possible to apply the optimized thresholds we derive here for winter 2016-2017 to data collected in other years. However, there is still significant work to be done to assess the interannual radiometric stability of S-1 across the GrIS at various signal-to-noise ratios and to characterize other forms of instrumental uncertainty, particularly due to the evolving observation strategy of S-1 and missing measurements from either S-1A or S-1B in various years. Additionally, evolving conditions on the GrIS, particularly in response to extreme melt (Culberg et al., 2021) and increasing rainfall (Box et al., 2022, 2023), may significantly alter the subsurface stratigraphy, and therefore the observed backscatter, in ways that are not yet well-understood. Further work in required to fully characterize the physical and dielectric mechanisms that drive C-band sensitivity to firn, ice slabs, and superimposed ice structure and how their radiometric signatures may change with time. Future work might also focus on improving the discrimination of crevasses and buried or drained lakes, which can currently lead to misclassifications in ice slab regions. Regardless, the algorithm we develop here lays the groundwork for generating long time series of ice slab expansion from C-band SAR observations with sufficient spatial coverage and resolution to enable long-term monitoring and validation of predictive numerical models.

Data availability. Final Sentinel-1 mosaics (shown in Figure 2) and the final ice slab extent in winter 2016-2017 will be deposited at the NSF Arctic Data Center with a permanent DOI at the time of manuscript acceptance. For the purposes of peer review, the current data sets are temporarily available at: https://drive.google.com/drive/folders/17NzLt60h8iWwQ8JP72mBsYaNFjLhOesx?usp=sharing [Last Access: 2023-11-08]. All Sentinel-1 data were accessed and processed through Google Earth Engine. The data catalog entry can be found at https://developers.google.com/earth-engine/datasets/catalog/COPERNICUS_S1_GRD [Last Access: 2023-11-08]. Ice-penetrating radar detections of slabs are available at https://doi.org/10.5281/zenodo. 7505426 [Last Access: 2023-11-08] (Jullien, 2023). Ice-penetrating radar survey lines and the radargrams shown in Figures 1 and 10 are available from the Center for Remote Sensing and Integrated Systems at https://data.cresis.ku.edu/data/accum/ or through the National Snow and Ice Data Center at https://nsidc.org /data/iracc1b/versions/2 [Last Access: 2023-11-08] (Paden et al., 2014a, b). The elevation of the visible runoff line as a function of time is available at https://zenodo.org /records/6472348 [Last Access: 2023-11-08] (Tedstone, 2022). Sentinel-1 firn aquifer detections are available at https://arcticdata.io/catalog/ view/doi%3A10.18739%2FA2HD7NS8N [Last Access: 2023-11-08] (Brangers et al., 2020). The data used for throughout this paper for basemaps of Greenland are available as follows. The Greenland coastline is available from the British Antarctic Survey at https://data.bas.ac.uk/ full-record.php?id=GB/NERC/BAS/PDC/01439 [Last Access: 2023-11-08] (Gerrish, 2020). The ice mask is available through BedMachine Greenland v4 https://sites.ps.uci.edu/morlighem/dataproducts/bedmachine-greenland/ [Last Access: 2023-11-08] (Morlighem et al., 2017). The 200 m elevation contours are derived from ArcticDEM and available at https://www.pgc.umn.edu/data/arcticdem/ [Last Access: 2023-11-08] (Porter et al., 2018).

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Author contributions. RC conceived the study, designed and implemented the processing algorithms, and wrote the initial paper draft. RJM and JZM provided guidance and suggestions on the Sentinel-1 processing and ice slab detection workflow. All authors contributed to the 380 scientific analysis of results and writing of the final manuscript.

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