



Icebergs Modify Antarctic Landfast Sea Ice Mass Balance

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Abstract. Icebergs locked in landfast sea ice strongly modify snow redistribution on Antarctic landfast sea ice, with important but still poorly constrained consequences for local ice and snow mass balance and ecology. Here, we revisit ice and snow accumulation around icebergs in Atka Bay (eastern Weddell Sea) using a combination of a digital elevation model (DEM) generated from drone photogrammetry, ground-based electromagnetic (EM) induction sounding, manual drilling, snow pits, and ice core analyses. The DEM reveals a characteristic snowdrift morphology consisting of a thick windward drift, two lateral drifts, and a snow-free tail, consistent with previous airborne observations. Snow depths derived from the DEM agree well with manual measurements. Snow properties in the drifts are dominated by dense, wind-compacted slabs. Despite substantial snow loading in drift areas, only limited slush was observed at the snow–sea ice interface, with maximum slush thicknesses of 20 cm. Instead, heavily snow-loaded areas coincide with anomalously thick sea ice (7.8 m), compared to 1.6 m in areas with undisturbed snow cover. Ice core analyses indicate extensive snow-ice formation, suggesting efficient refreezing of flooded snow and a substantial contribution to sea ice mass balance. Basal-origin ice is inferred to be thinner beneath the drifts, while thicker consolidated ice was observed in the snow-free tail. In contrast to previous airborne and modeling studies that limited mass-balance assessments to the snow distribution, this study provides direct, co-located observations of snow, slush, and ice properties, providing evidence that iceberg-induced snowdrifts in Atka Bay can promote substantial snow-ice formation and upward ice growth.



1 Introduction

Snow cover exerts a first-order control on the thermodynamic and mass balance of Antarctic landfast sea ice by various mechanisms: snow insulating the underlying sea ice, hydrostatic loading reducing sea ice freeboard, preconditioning for flooding, and snow-to-ice conversion (Arndt et al., 2020; Wever et al., 2021; Langhorne et al., 2023). Snow redistribution and accumulation strongly influence seasonal ice growth, internal structure, and the stability of landfast ice (Lei et al., 2010; Arndt et al., 2020; Massom et al., 2010; Zhao et al., 2022), with implications for ecosystems and nearshore operations (Fraser et al., 2021, 2023).

Snow accumulation on landfast sea ice is highly heterogeneous (Wever et al., 2021; Fraser et al., 2023), as wind-driven redistribution often dominates over direct precipitation, producing strong spatial gradients in snow thickness (Pomeroy and Gray, 1990; Doorschot and Lehning, 2002; Hames et al., 2022). In Antarctic coastal environments, where winds are persistent, drifting snow is a primary mechanism shaping the snow cover (Fraser et al., 2023), making its representation critical for assessing local ice and snow mass balance.

Icebergs embedded within landfast sea ice act as large, persistent obstacles that substantially modify near-surface airflow and snow transport (Franke et al., 2025; Hames et al., 2026). By altering wind speed, turbulence, and flow separation, icebergs generate characteristic patterns of snow deposition and erosion that differ fundamentally from surrounding undeformed fast ice (Franke et al., 2025; Hames et al., 2026). These iceberg-induced snowdrifts can increase local snow loading, depressing the ice surface below sea level and allowing seawater to flood the lower snowpack and form slush that may subsequently refreeze as snow-ice (Jeffries et al., 2001). This process can enhance snow-ice formation, linking atmospheric forcing to sea ice mass balance in coastal Antarctica (Arndt et al., 2020; Fraser et al., 2023). In Atka Bay, the meter-thick sub-ice platelet layer (SIPL) adds buoyancy which can reduce flooding despite substantial snow loading (Arndt et al., 2020).

Recent airborne multi-sensor observations from Atka Bay have provided the first spatially extensive morphology of iceberg-induced snow redistribution on Antarctic landfast sea ice (Franke et al., 2025). Franke et al. (2025) documented a recurring snowdrift morphology consisting of thick windward drifts, elongated lateral drifts aligned with the prevailing wind direction, and largely snow-free tails. Using microwave radar signatures, they further inferred flooding and slush formation beneath snowdrifts. These mass-balance impacts were inferred from radar signatures.

Building on these observations, Hames et al. (2026) combined numerical snow transport modeling with field data to investigate the physical controls on snowdrift formation around Antarctic icebergs. Their results demonstrated that persistent wind directions produce stable drift morphologies and they revealed a robust sublinear scaling between iceberg size and associated snowdrift area, indicating reduced snow accumulation efficiency for larger icebergs. The study also emphasized the importance of wind-direction and maximum wind speed variability in shaping lateral and windward snowdrifts. While this work advanced the mechanistic understanding of snow redistribution, its implications for snow-ice formation and associated sea ice mass balance have not yet been quantified. Here, we build on these observations by directly investigating the structure of the underlying ice and assessing the potential season-integrated consequences of iceberg-induced snow redistribution.

While the role of grounded icebergs in stabilizing landfast sea ice has become increasingly recognized (Fraser et al., 2021; Nihashi and Ohshima, 2015), existing thermodynamic studies have largely focused on iceberg melt and its effects on ocean



stratification and basal sea ice growth (Jongma et al., 2009; Merino et al., 2016; Lucas et al., 2025), whereas the role of iceberg-induced snow redistribution and snow-ice formation at the sea ice surface remains largely unexplored. In particular, it is unclear whether heavy snow loading primarily leads to sustained flooding and slush persistence, or whether cold conditions and efficient refreezing convert flooded snow into snow-ice, contributing positively to sea ice thickness. Addressing this gap requires co-located observations of snow and sea ice thickness and properties, and slush occurrence around icebergs.

Here, we revisit ice and snow mass balance around icebergs in Atka Bay using an integrated observational approach combining UAV-based photogrammetric snow thickness mapping, electromagnetic (EM) induction measurements, manual drilling, snow pits, and ice core analyses. By directly quantifying snow properties, slush presence, and ice thickness beneath iceberg-induced snowdrifts, this study provides new observational constraints on the processes linking snow redistribution to sea ice mass balance and snow-ice formation in coastal Antarctica.



2 Data and Methods

2.1 Atka Bay sea ice and snow conditions in 2025 and study site selection

All snow and ice data analyzed in this study were collected in November and December 2025, with the study site accessed from Neumayer III Station (Wesche et al., 2016). The fast ice present in Atka Bay during the survey formed progressively from west to east between early April and the end of May 2025. Bay-wide, long-term records on four transects were established in 2018, 2022 (Arndt et al., 2020; Neudert et al., 2024) and 2023 (unpublished), which were revisited for drilling in 2025 (locations in Fig. 1 A). On these transects, the consolidated ice thickness in 2025 was 160 ± 22 cm (mean $\pm$ standard deviation). This is lower than in 2022 (198 ± 31 cm) and 2023 (192 ± 52 cm), but similar to 2018 (168 ± 21 cm) (Arndt et al., 2020). The mean snow thickness in 2025 was 65 ± 27 cm, and the mean SIPL thickness was 472 ± 176 cm.

Approximately 90 icebergs were frozen into the fast ice of Atka Bay in 2025, from manually delineating them in SAR imagery (polygons in Fig. 1 A). The largest had an extent of about 2 km, while the smallest berg identifiable in SAR was around 60 m. Icebergs occupied approximately 1.5% of the bay area. Because the SAR inventory detects icebergs only above an approximately 60 m size threshold, smaller icebergs are likely underrepresented, introducing uncertainty into the inventory. In addition, extensive bare-ice zones with little or no snow accumulation were identified during on-ice travel. These zones extended for more than 1 km in the lee of some icebergs.

We selected one iceberg that measured 170 m in the north–south and 125 m in the east–west direction. It was chosen to provide a reasonable size for logistical purposes during the survey. The iceberg was positioned east of most other icebergs, suggesting that it experienced a less disturbed wind field, as the dominant wind direction is from east to west (Klöver et al., 2013), interspersed with occasional events from the southwest to southeast.

Wind conditions have been monitored continuously at Neumayer III Station and its predecessor stations since 1981 (König-Langlo and Loose, 2007). The mean wind speed 7.88 m s^{-1} during the 2025 sea ice season (averaged over 30 April till 11 December) with East as the prevailing direction (windrose in Fig. 3). Wind events with 10-m wind speeds exceeding 10 m s^{-1} accounted for 26% of the observations, with 96.5% of these events occurring under ENE–ESE winds. This is above the 6–8 m s^{-1} threshold at which drifting snow has been observed to initiate in Antarctica (Leonard et al., 2012).

2.2 Drone photogrammetry derived DEM

A systematic drone survey was conducted on 26 November 2025 centered on the iceberg using a DJI Mavic 3. Images were acquired over a 1000×1300 m area in a nadir grid pattern with 80/60% front/side overlap at an average flying altitude of 136 m, resulting in a ground sampling distance of 4.75 cm pix^{-1} . A total of 16 ground control points (GCPs; 50×50 cm) were distributed across the survey area (see positions in Fig. 3) and their position measured with a Novatel GNSS antenna. Snow thickness was additionally measured at each GCP using an avalanche probe.

Images were processed in Agisoft Metashape. Initial image alignment was based on the drones onboard GNSS. All 16 GCPs were then manually marked in the images and used to optimize the camera model and georeference the photogrammetric model. The model was vertically referenced using the manually measured snow thickness at the GCP locations. Internal GCP residuals

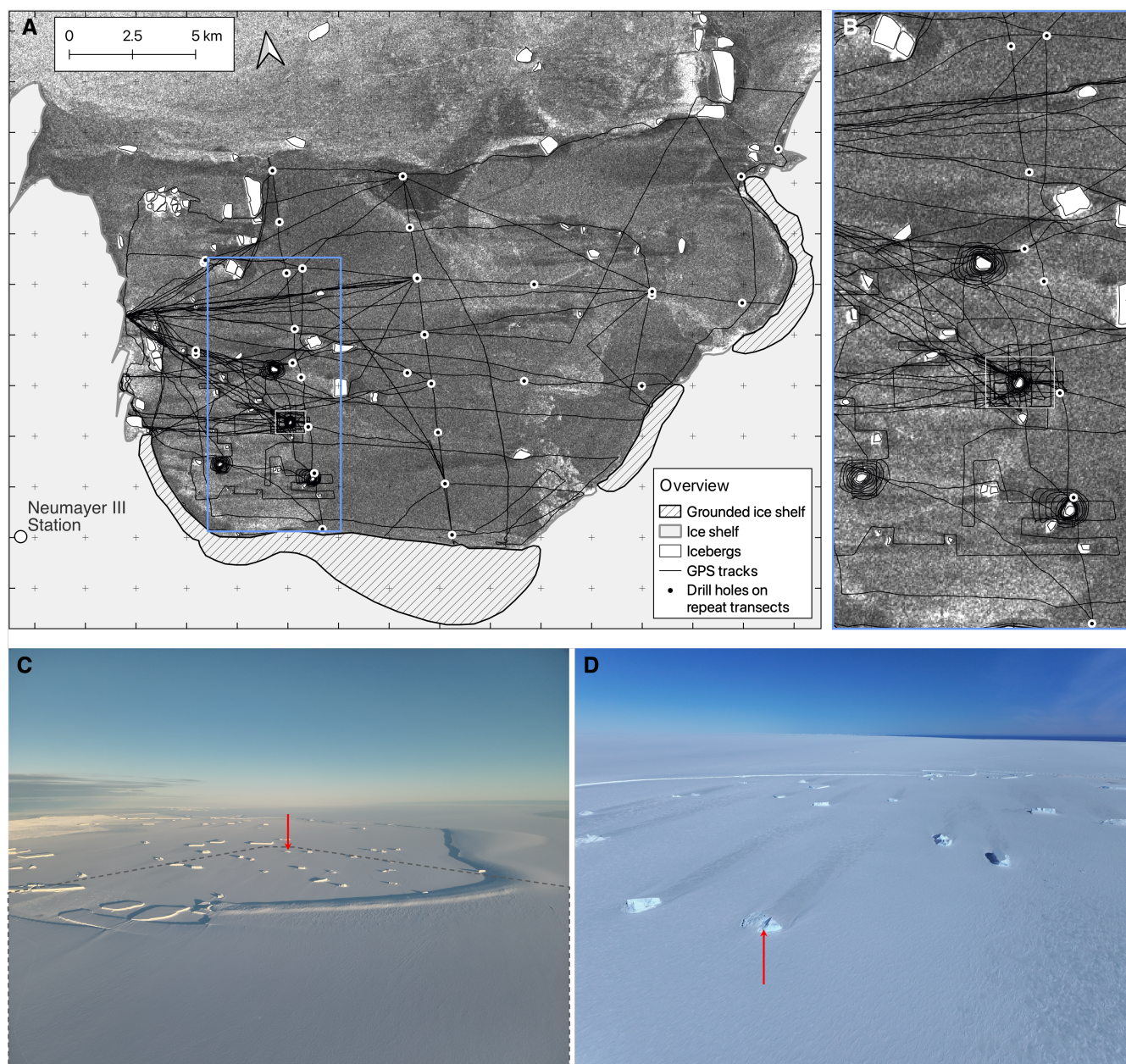


Figure 1. A) Overview of the study area at Atka Bay showing the grounded ice shelf (hatched), icebergs (white polygons), GPS tracks of the GEM-2 surveys conducted during the field campaign, and the location of Neumayer III station. The white rectangle outlines the location of the drone photogrammetry DEM in Fig. 3. The circles indicate the locations of the drill holes on repeat transects. The blue rectangle marks the area enlarged in B). Background show a Sentinel-1 SAR image on 30 November 2025. B) Close-up of the study area. C) Aerial photo of Atka Bay (9 November 2025). The stippled outline indicates the field of view in D). The red arrow points to the studied iceberg. D) Aerial photo of the vicinity of the studied iceberg (3 December 2025).



were 28.0 cm horizontally and 1.8 cm vertically. Because no GCPs were withheld as checkpoints, these values describe internal georeferencing consistency rather than independent model accuracy. After sparse-cloud cleaning, depth maps were generated and used to reconstruct the snow-surface elevation model. The final model had a grid resolution of 19.0 cm pix^{-1} . The DEM was subset to a 700x1000 m area (Fig. 3), excluding areas near the survey boundaries where reduced image overlap resulted in visually lower reconstruction accuracy.

Because the underlying surface (the sea ice freeboard) was approximately level across the area of the GCPs, variations in the DEM surface elevation are interpreted as variations in snow thickness. To assess the accuracy of this snow thickness model, it was compared with manual snow thickness measurements at the drill holes and along two transects: one on the windward side and one lateral transect (grey lines with color-coded measurements in Fig. 3). For a detailed comparison of the DEM snow thickness and manual snow thickness see Appendix A.

2.3 Snow thickness and thickness drilling in the iceberg vicinity

Thickness drilling was conducted between 26 November and 3 December. The locations were selected to be representative of the broader area influenced by the iceberg on a grid centered at the iceberg, consisting of an inner square with 8 drill holes (side length 370 m) and an outer rectangle (side length 700x1000 m) with 11 drill holes.

At each drill-hole location, we first excavated a snow pit to measure snow thickness and characterize the snowpack, including the presence of ice layers and slush at the snow–sea ice interface. A slush sample was taken for salinity analysis if the slush depth was 5 cm or more. Interface and snow temperatures were measured, and an auger drill together with a modified ice thickness tape (Crocker and Wadhams, 1989; Gough et al., 2012) was used to measure sea ice thickness and sub-ice platelet layer thickness.

Additionally, snow thickness was measured along two profiles (Fig. 3): one extending eastward from the iceberg starting at the thick windward drift and another crossing the lateral snowdrift. Georeferenced measurements were acquired manually with a target spacing of approximately 5 m using a MagnaProbe (Sturm and Holmgren, 2018) and, where snow thickness exceeded the instrument's measurement range, an avalanche probe and a handheld GPS. In addition, holes were drilled every 250 m along these snow thickness profiles ($n=7$).

2.4 Snow pits and thickness drilling in the main snowdrifts

Two snow pit sites were established on 27 November to characterize the snowpack of the drifts: one in the windward drift and a second in the northern lateral drift. The sites were selected to capture the thickest snow accumulations, measuring 2.38 and 2.22 m, respectively. The windward site was chosen with some distance to the iceberg (75 m) to remain hydrostatically connected to the surrounding floating fast ice, which responds to the tidal cycle, and to maintain sufficient distance from the potentially grounded iceberg to avoid inadvertently drilling into any submerged iceberg extension. Excavation required substantial manual effort due to the highly compacted snowpack; cutting and removing blocks took approximately half a day with a four-person team.



At the windward pit, snow stratigraphy was logged in detail and bulk density, salinity, and temperature were measured at 10 cm vertical resolution. At the snow–ice interface, slush and basal snow were sampled for subsequent salinity analyses. At the lateral pit, stratigraphy was recorded and basal slush and snow were collected for salinity measurements. For regional reference conditions, 9 additional snow pits are available across Atka Bay in November and December 2025.

130 Thickness drilling through the snow pit floor to measure consolidated ice (CI) thickness and SIPL thickness was attempted on the same day in both snow pits using an auger and a coring system. However, with both systems only a total depth of 6 m was possible, and the full depth was not reached. Drilling in these main pits was repeated with a steam drill on 1 December.

Two key observations guided our subsequent analysis. First, flooding at both snow pits with 10 and 13 cm of slush and the corresponding negative freeboard, indicated that much of the seasonal flooding had likely been transformed into snow-ice.
 135 However, this interpretation is not unique, as prior basal ice growth and non-local buoyancy effects may also contribute to the observed freeboard. The presented explanation is further evaluated using an ice core. Second, the negative freeboard after drilling was the same as the depth of flooding before drilling. This limited hydraulic head suggested that the ice–snow system was close to hydrostatic equilibrium, i.e., there was no sustained excess hydraulic pressure maintained within the ice at these locations.

140 2.5 Ice cores

To support our hypothesis of thick snow-ice below the snowdrifts, coring was attempted at the two snow pits using a Kovacs ice corer; however, the available extension sections were insufficient to recover full-length cores. The entire 6 m of the extracted cores from both pits were soft when drilling and appeared to be snow-ice (granular texture with rounded ice grains packed together, see Fig. B1 A and B).

145 A complete ice core was successfully obtained on the windward side of the iceberg on 29 November, approximately 50 m east of the main snow pit. Manual texture analysis and visual inspection at the time of coring indicated that the core consisted mainly of opaque, soft, easy-to-drill snow ice. Only the lowermost 0.65 m section of the 4.54 m core was clear sea ice (Fig. B1 C). In a cold room, vertical thick-sections were cut for sea ice structural analysis, and 10-cm horizontal core slices were melted for bulk salinity measurements.

150 Additional ice cores were collected across Atka Bay along the transects described by Arndt et al. (2020). An ice core 6 km East of the studied iceberg, designated ATKASNE04, is presented here.

2.6 Electromagnetic induction sounding surveys

Systematic ground-based electromagnetic (EM) induction sounding surveys were conducted on 24 and 29 November and opportunistic underway surveys on 25, 26, and 29 November and 15 December with a GEM-2 instrument (Geophex Ltd.). The
 155 five operating frequencies spanned 1530 to 93,090 Hz. Higher frequencies are more sensitive to conductivity variations near the surface and are therefore more responsive to the presence of saline, conductive slush associated with flooding (Neudert et al., 2026).



We applied two multifrequency inversion approaches: (i) a two-layer model combining ice and snow into a single, very low-conductivity layer above a SIPL layer of intermediate conductivity, as detailed in (Neudert et al., 2024); and (ii) an alternative scheme that includes slush as a separate highly conductive layer as described in Neudert et al. (2026), yielding a four-layer model, that better represents the sub-surface conductivity structure in the presence of saline slush.

The accuracy of the slush thickness retrieval is evaluated against the manual slush observations by aggregating all EM data within 5 m of a manual measurement and computing the mean absolute error and bias.

2.7 Snow-ice budget estimation from the DEM

The underlying ice thickness could be inferred from the snow surface elevations derived from a DEM, if the ice freeboard was known. One approach would be to convert surface elevation to ice thickness via Archimedes' principle, using snow and ice densities to compute the ice thickness required to support the snow load in hydrostatic equilibrium. However, this correction would require accurate snow density measurements, which are not available. The mean snow density derived from nine additional snow pits across Atka Bay was 390 kg m^{-3} , whereas the density at the main snow pit was 455 kg m^{-3} . Moreover, snow density varied considerably with depth. In addition, we lack reliable values for snow-ice density. The SIPL is expected to contribute approximately 12 cm of additional freeboard in the area, for a mean SIPL thickness of 5.25 m from drill holes and an estimated solid fraction of 0.22 (Neudert et al., 2024).

We therefore pursued an alternative approach that additionally allows separation of bottom-grown sea ice from snow-ice. We assume that up to a certain surface height, ice growth over the season was solely by thermodynamic basal accretion. We define this cutoff as a maximum of 75 cm, based on a linear regression between freeboard and snow thickness from drill holes across the bay (Fig. 2 A). We derive an empirical linear relationship between surface elevation and ice thickness from all drill hole measurements across the bay (excluding those at a section of younger ice but including those around the iceberg).

Above this surface elevation, the observed ice thickness is likely to exceed the thermodynamically bottom-grown ice thickness and to include an additional snow-ice contribution generated by flooding during the season. Using the drill holes around the iceberg, we model the resulting total consolidated ice thickness with a power-law relationship (Fig. 2 B, upper branch of the curve) to account for increasing snow compaction with increasing snow thickness.

The lower fraction of this total consolidated ice thickness must be bottom-grown sea ice. For the thickness of the bottom-grown ice fraction, we have one direct constraint from the windward ice core, where the lowest 0.65 m of the 4.54 m ice core were bottom-grown ice. We therefore complement this with an additional assumed bottom-grown ice thickness of 0.30 m beneath the main windward snow pit with a total consolidated ice thickness of 7.84 m to set up an exponentially decaying model (Fig. 2 B, lower branch of the curve). These curves represent a conceptual interpretation quantitatively constrained by observations.

These empirical snow surface elevation–thickness relationships are applied to the DEM snow surface elevation and the resulting snow–ice and bottom-grown sea ice thicknesses distributions are presented in Fig.2 C. One source of uncertainty in the thickness retrieval arises from potential systematic offsets in the measured surface elevation. To quantify the sensitivity of

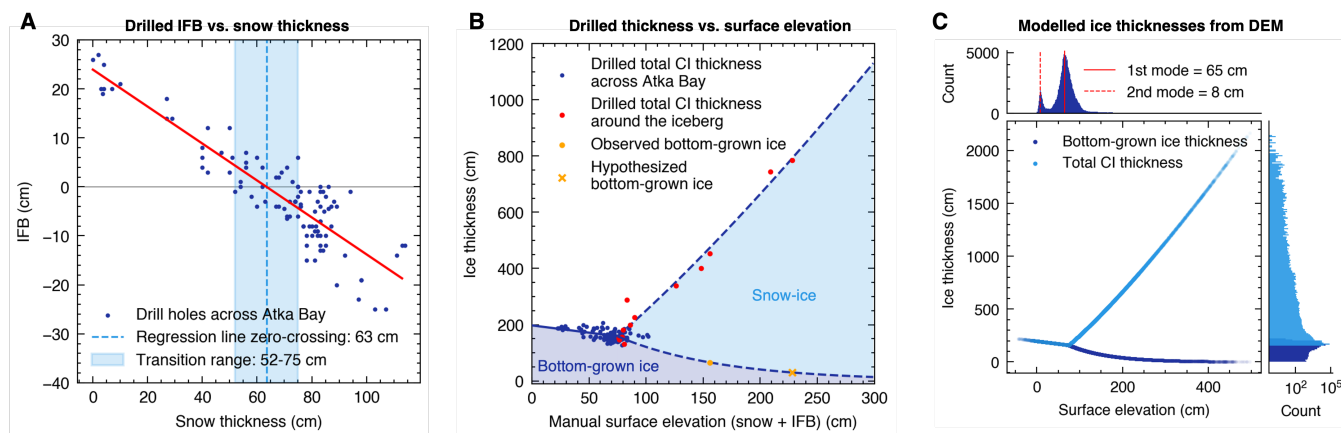


Figure 2. A) Regression line for ice freeboard (IFB) versus snow thickness for all drill holes across Atka Bay in November and December 2025. IFB was always negative for snow thickness over 75 cm. B) Drilled ice thickness versus manually measured surface elevation (sum of snow thickness and IFB) for all drill holes across Atka Bay (blue markers) and for the drill holes around the iceberg where surface elevation was >75 cm (red markers). The one measured bottom-grown ice thickness from the ice core is shown as a yellow circle, while the hypothetical bottom-grown ice thickness is shown as a yellow cross. These data points form the basis for our empirical piecewise regression of ice thickness versus snow surface elevation representing bottom-grown ice (=total consolidated ice thickness) up to 75 cm surface elevation. Above, the relationship is split into total consolidated ice thickness and bottom-grown ice thickness (blue area). Snow-ice thickness (light-blue) is calculated as the residual between the modeled total consolidated ice thickness curve and the modeled bottom-grown ice thickness. C) Main: Ice thickness and ice types derived from the DEM snow surface elevation using the curves in B). Top: Distribution of surface elevations in the DEM. Right: Resulting log-normal distributions of total consolidated ice thickness (light blue) and bottom-grown ice thickness (dark blue).

the derived thickness to such biases, uniform elevation offsets of ± 0.10 m were applied, and the corresponding thickness fields were recalculated.



3 Results

3.1 Snow thickness

195 The DEM in Fig. 3 and the snow thickness profiles in Fig. 4 show a distinct snow drift morphology associated with the iceberg. On the windward side of the iceberg, a dominant snow accumulation extended approximately 270 m in the east–west direction and 270 m in the north–south direction. The feature has a maximum elevation of 4.86 m. The two lateral drifts are oriented approximately parallel to the wind direction, with peak surface elevations of 2.48 m (northern drift) and 2.08 m (southern drift), and lengths and widths of roughly 300x100 m.

200 Downwind of the iceberg, a distinct zone of reduced snow accumulation and low surface roughness is evident. This low-accumulation region extends for at least 690 m and continues beyond the DEM domain with mean surface elevation of only 0.15 m. In comparison, the surroundings of the iceberg within the DEM that are not attributed to any of the features have a mean snow surface elevation of 0.68 m.

The snow thickness derived from the DEM agree with in situ measurements from the transects and manual drilling. Across all
 205 co-located points within, the DEM–snow profile difference is -5.3 cm with RMSE of 14 cm, and the DEM–drilling difference is -10.8 cm with RMSE of 17 cm (Fig. A1 B and C). Manually measured snow thickness spans 0–2.38 m, and the spatial pattern of maximum snow thickness in the windward and lateral drift and reduced thickness outside the drifts, is consistently reproduced.

3.2 Snow properties

210 The snowpack of 2.22 and 2.38 m thickness in the two main snow pits consisted entirely of wind-compacted snow (wind slab) with pen-to-knife hardness, indicating extensive wind-driven metamorphism. In the windward snow pit, at the base of the snowpack, the lowest 10 cm comprised saline slush, with an abrupt transition from the overlying hard wind slab and no intervening layer of progressively softer snow. Snow densities were consistently high, ranging from 450 to 550 kg m⁻³, with a mean density of 455 kg m⁻³, except for the surface layer, which had a density of approximately 300 kg m⁻³ but had been
 215 disturbed by the snow pit excavation work. These density measurements represent the lower bound because the exceptional hardness of the snow made it difficult to completely fill the density cutter. Salinity measurements indicated that the upper 2.20 m of the snowpack were fresh (0 ppt), whereas the basal slush layer had a salinity of 13.4 ppt and the 3 cm immediately above it exhibited an intermediate salinity of 4.7 ppt, indicating upward salt infiltration from the flooded snow–ice interface.

3.3 Sea ice thickness in the iceberg vicinity

220 In the entire study area surrounding the iceberg, sea ice thickness from drill holes (which is equivalent to the DEM-derived total CI thickness) ranges from 1.32 to 7.84 m (see Fig. 5 and Fig. 4). The thicknesses in drill holes at distances larger than 320 m (n=16) have an average thickness of 1.60±0.20 m which is comparable to the bay-wide average thickness of 1.63±0.17 m. The largest thicknesses of 7.84 and 7.45 m are found in the heavy snow loaded areas of the main snow pits. The DEM-derived

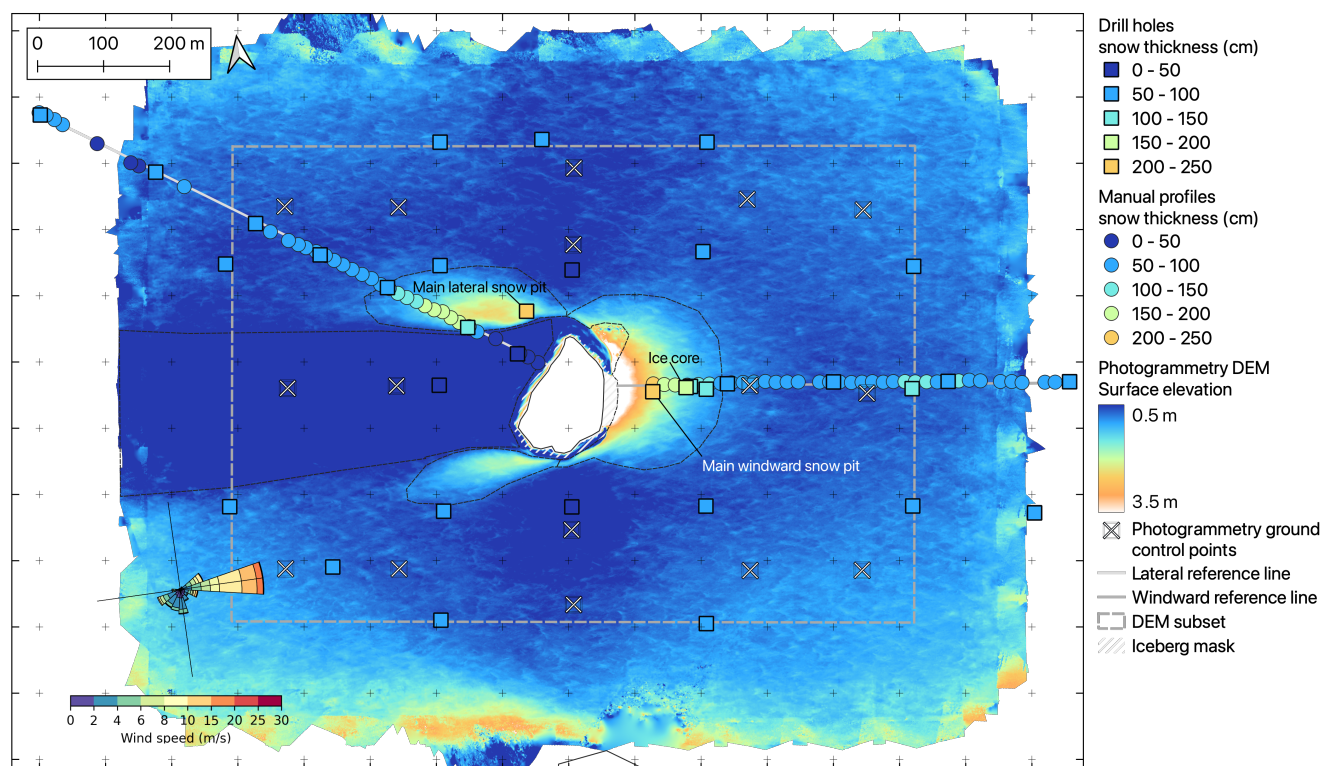


Figure 3. Map of surface elevation from drone photogrammetry (DEM). Snow thicknesses manually measured in pits at drill holes are shown as squares. Snow thicknesses measured along a snow-thickness profile (no coincident drill holes) are shown as circles. Marker colors denote the measured snow thickness. DEM Background color corresponds to the surface elevation. The DEM-derived surface elevation agrees well with the manual snow thickness measurements across the mapped area. The white dashed outline marks the DEM subset used for subsequent analysis to exclude photogrammetry edge effects. The black dashed outlines mark the windward and lateral snowdrifts and the snow-free tail. The wind rose shows wind speed and direction for April to December 2025 from the Neumayer III) weather station in 2025.

total CI thickness exhibits the same spatial distribution, with enhanced thickness co-located with the windward and lateral snow drifts.

The EM snow+consolidated ice thickness from the GEM-2 surveys reproduces this spatial organization. In the tail, no snow was present; thus, the EM ice+snow thickness is expected to correspond to the drilled thickness. Here, the mean EM thickness is 1.99 ± 0.23 m (min/max: 1.40/3.78 m). This compares well with the mean thickness from three drill holes of 1.94 m (1.85 to 2.08 m), with a snow thickness of 0.00 to 0.04 m. Within the windward snowdrift the EM thickness mean with 3.97 ± 0.81 is about 2 m higher than in the tail, a trend that is consistent with higher drilled snow+ice thickness, although the thicknesses are severely underestimated by the EM method. The large total thickness, for example, 10.22 m of snow and ice at the windward snow pit, reduces instrument sensitivity and limits the ability to accurately retrieve variations, especially since the sub-surface model also includes a SIPL with variable conductivity.

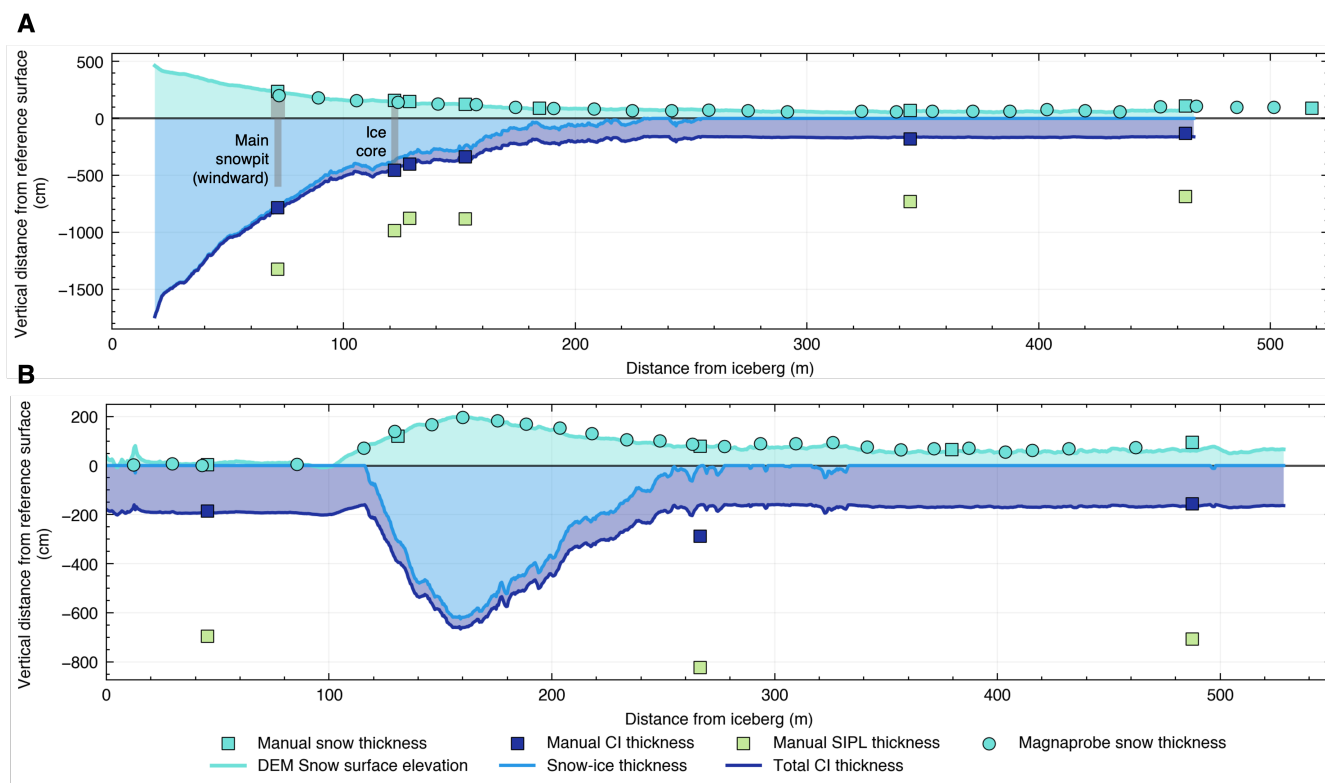


Figure 4. Thickness profiles along A) the windward reference line, and B) the lateral reference line. The locations of the lines are also shown in Figs. 3 and 6. The markers indicate the manual snow and ice thickness measurements. The thick lines are the snow surface elevation, total consolidated ice thickness, and bottom-grown ice thickness derived from the DEM using the surface-elevation-to-ice-thickness relationship.

3.4 Sea ice composition

235 The partial cores retrieved from the main snow pits revealed that the entire upper 6 m of retrieved material consisted of granular ice consistent with snow-ice formation. A complete core at a distance of 122 m from the iceberg along the windward profile yielded a total ice thickness of 4.54 meters. The recovered core length (3.90 m) was shorter than the measured ice thickness (4.54 m). The difference of 0.64 m is attributed to core loss during drilling and handling, as well as potential compaction and fragmentation of the porous upper ice layers (Fig. B1 A and B) that were particularly susceptible to fracture during coring.

240 Notably, only the lowermost 0.65 m of this core exhibited characteristics of sea ice, which is indicative of congelation growth (Fig. B1 C). The lowermost section of the core was recovered as a single intact piece, preserving the location of the transition.

The upper part of the core has an orbicular granular (og) ice texture (thick section photo in Fig. B1 D), characterized by coarse grains and abundant pores, which is consistent with a significant contribution from snow-ice. The transition from granular to columnar texture (thick section photo Fig. B1 E) with incorporated platelet ice (PI) indicates congelation growth.

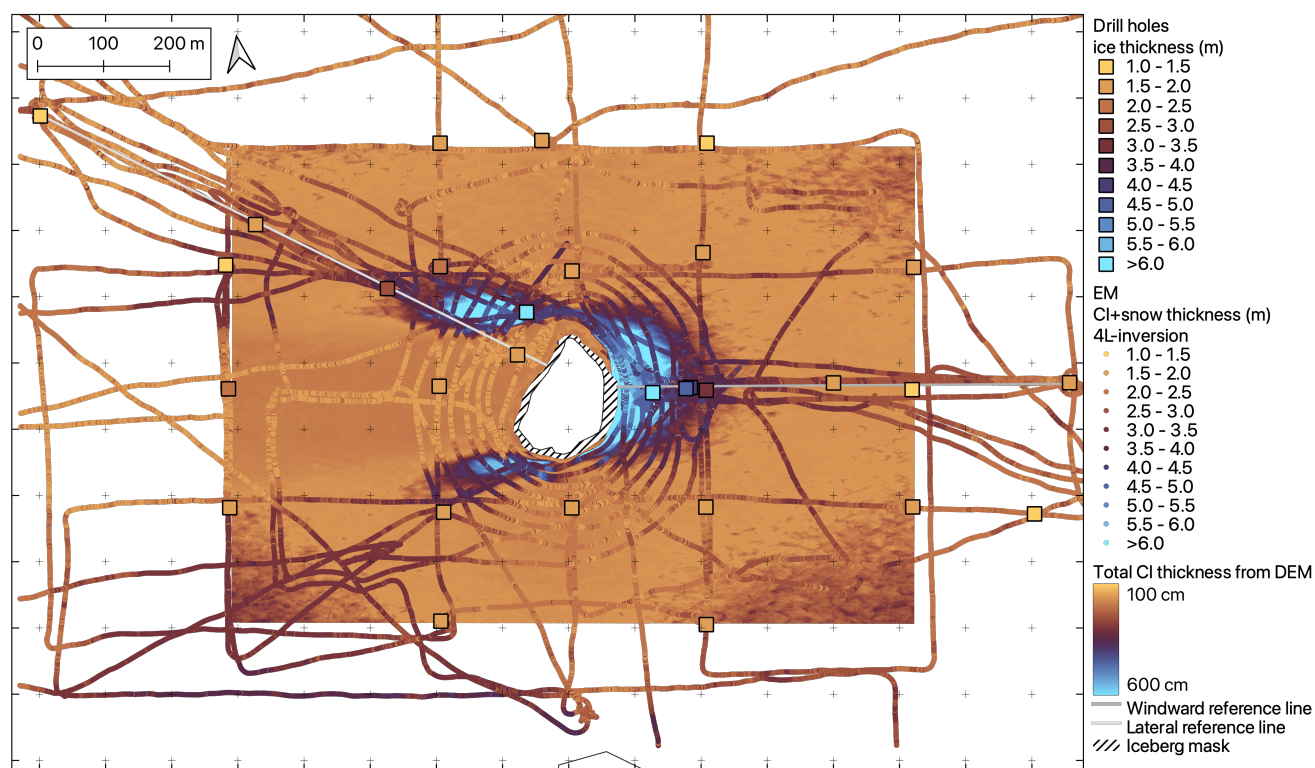


Figure 5. Map of total consolidated ice thickness derived from the DEM using the curves in Fig. 2 B. Ice thicknesses manually measured at drill holes are shown as rectangles. EM total thickness (consolidated ice plus snow thickness) shown as lines.

245 The lower part of the core is dominated by incorporated platelet ice and mixed polygonal ice textures, typical for Atka Bay (Hoppmann et al., 2015).

The salinity profile can be divided into three distinct zones: (1) a relatively saline upper section (0–90 cm; 2.9–3.9 ppt), (2) a low-salinity middle section (90–311 cm; 1.2–1.9 ppt), and (3) a more saline basal section below 311 cm (2.6–3.6 ppt). The changes in salinity broadly coincide with variations in pore structure along the core. Sections characterized by larger and more interconnected pores may have experienced more extensive brine drainage immediately after core extraction, leading to lower measured bulk salinities. Therefore, part of the observed salinity variability may be attributable to post-recovery brine loss rather than differences in the ice formation environment alone.

In contrast, the core collected 6 km East of the iceberg (ATKASNE04) is dominated by platelet ice (PI), demonstrating substantial small-scale spatial variability in basal ice-growth processes. Beneath a thin surface layer of coarse orbicular granular ice (0–11 cm) and a narrow transition zone of mixed platelet ice (11–30 cm), the remainder of the core (30–164 cm) consists of platelet ice. Salinity decreases gradually from 7.6 ppt at the surface to approximately 3.5–4.0 ppt through the middle of the core, before increasing slightly to 4.8 ppt in the basal section. Compared to the core below the windward snowdrift, which



exhibited a distinct low-salinity middle section and abrupt salinity changes associated with variations in pore structure, the ATKASNE04 core shows a gradual vertical salinity gradient.

Table 1. Summary of ice textures and bubble characteristics along the ice core from the windward snow drift retrieved on 29 November. Depths are given in cm (og = orbicular granular, col = columnar, mixed PI = columnar ice and incorporated platelet ice, PI = incorporated platelet ice).

From (cm)	To (cm)	Texture	Description
0	-325	og	Abundant smaller pores (Fig. B1 D)
-325	-332	og	Few larger pores
-332	-364	col	Columnar ice (Fig. B1 E)
-364	-383	mixed PI	Mixed columnar and incorporated platelet ice
-383	-397	PI	Incorporated platelet ice

260 The results of upscaling these point observations to spatially distributed bottom-grown sea ice and snow-ice thicknesses using the surface-elevation-to-ice-thickness relationships derived in Section 2.7, and applying these relationships to the DEM are illustrated in Fig. 6 and Fig. 4. Over the entire DEM area, the resulting mean snow-ice thickness (Fig. 6 C) is estimated at 0.49 m (0.36–0.71 m), while the mean thickness of the bottom-grown consolidated ice (Fig. 6 B) is estimated at 1.56 m (1.47–1.63 m). These values are model-dependent estimates derived from the DEM-based thickness retrieval and are not
 265 direct measurements of snow-ice and bottom-grown ice thickness. Based on these estimates, the snow-ice to bottom-grown ice thickness ratio is 0.31 (0.22–0.48) in the DEM area. The application of ± 0.10 m surface-elevation offsets results in the ranges in brackets, showing the sensitivity of the model-derived thickness estimates to small biases in surface elevation and underscoring the importance of accurate vertical referencing.

The reduced mean bottom-grown ice thickness beneath the windward and lateral snowdrifts (0.85, 1.03, and 1.19 m, respec-
 270 tively) is contrasted by greater bottom-grown ice thickness in the snow-free tail, where the mean thickness reaches 1.90 m. This estimate is supported by drill-hole measurements in the tail, which yield a mean ice thickness of 1.94 m. The bottom-grown ice thickness beneath the snowdrifts is constrained by only a single ice core, so the DEM-derived estimates for these areas cannot be validated directly.

3.5 Flooding

275 Manual observations of flooding of the lowest snow layer were conducted at 35 locations (squares in Fig. 7). Slush thicknesses ranged from 0 to 33 cm (mean 5.3 cm). At 8 locations, a slush sample was taken. The salinities ranged from 13 to 40 ppt, with the lowest salinities found in the main snow pits and the highest at a location where a layer of hard refreezing slush was present on top of soft slush.

At all 29 drill holes, a freeboard measurement was taken. In 12 cases, the freeboard was positive or zero. These measurements
 280 were associated with no slush in 10 of 12 cases, while 1 and 2 cm of slush were observed in the remaining two cases. In 17

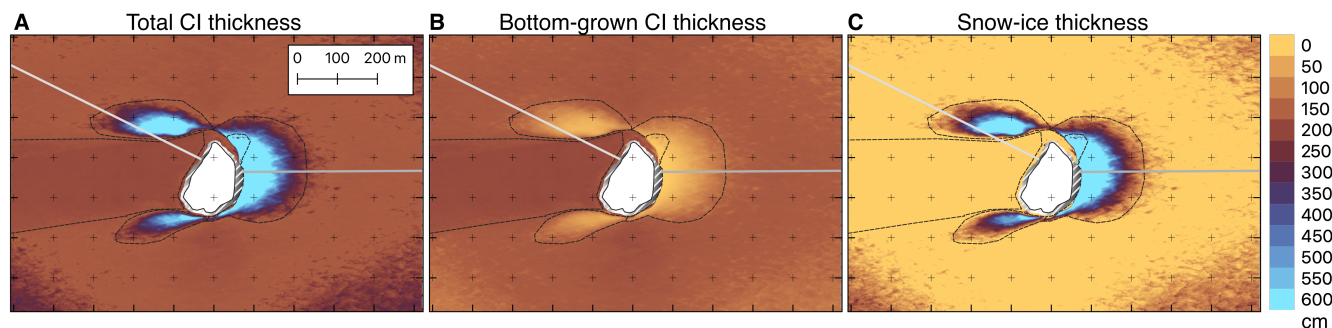


Figure 6. A) Map of total consolidated ice thickness, B) the bottom grown consolidated ice thickness, and C) the snow-ice thickness derived from the drone photogrammetry (DEM) using the surface-elevation-to-ice-thickness relationship. Outlined areas are the tail, and the lateral and windward snowdrifts. The color indicates the respective thickness.

cases, freeboard was negative. Agreement of slush depth and freeboard within ± 2 cm occurred in 8 of these 17 observations, while 9 of 17 observations differed by more than 2 cm. For observations with differences greater than 2 cm, the differences ranged from -9 to 3 cm, with a mean of -6 cm. Here, negative values indicate that the observed slush thickness was less than the negative freeboard would have allowed for, or in simple terms that water infiltrated previously dry snow after drilling. This suggests that, in some location snow layers have not been flooded, although could have been flooded under local hydrostatic equilibrium.

The EM slush thickness agree with manual measurements (Fig. 7 B), but overall the EM method overestimates the slush thickness by 6 cm, which may be attributed to the very high and variable slush salinity and ambiguity in the layering. The mean absolute error of 7.3 cm is larger than the mean absolute error of 4 cm in Neudert et al. (2026). The larger error can have several reasons; the additional uncertainty introduced by a SIPL layer, potentially higher ice conductivity due to a different snow-ice porosity compared to columnar sea ice, and larger instrument roll and tilt in this study compared to the application on level fastice in Neudert et al. (2026).

3.6 SIPL thickness distribution

The SIPL inferred from thickness drilling surrounding the iceberg has a mean thickness of 5.25 m and a standard deviation of 0.57 m. No obvious spatial pattern was identified in the distribution of the sub-ice platelet layer.

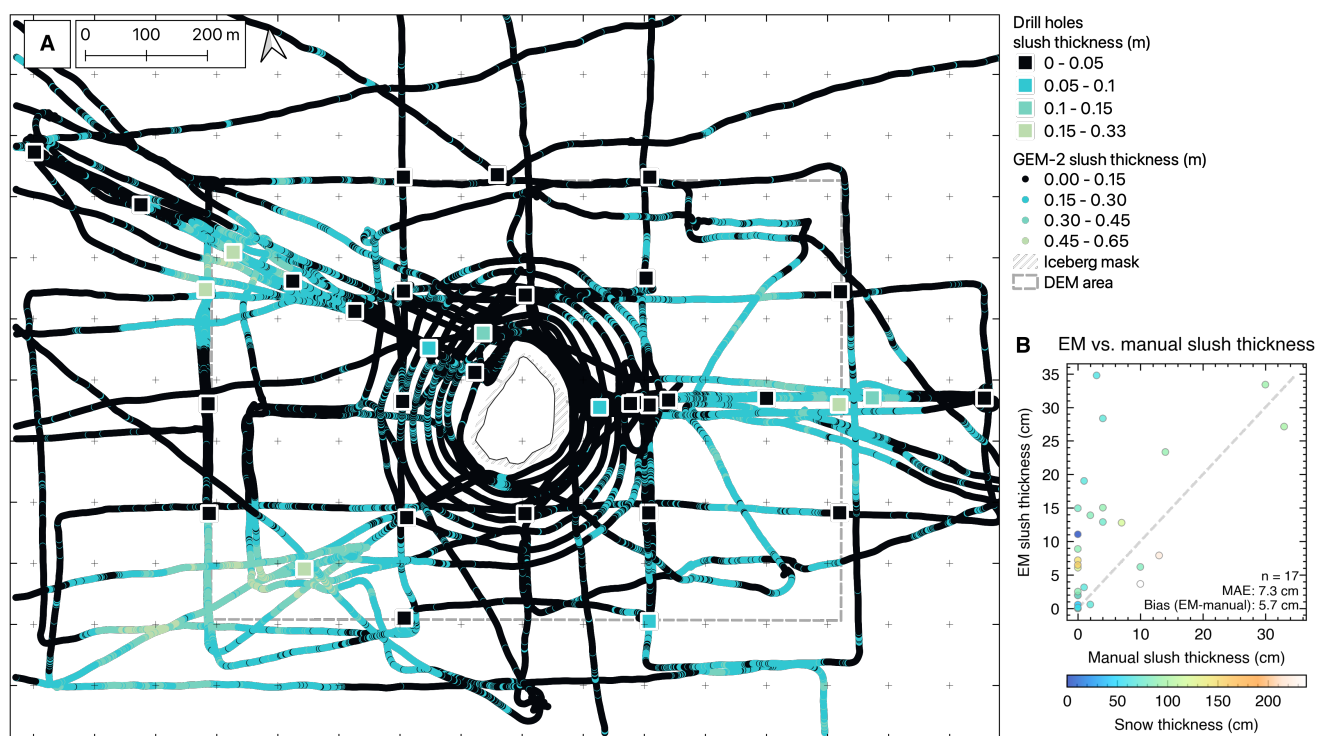


Figure 7. A) Map of slush thickness. Slush thicknesses measured at drill holes and snow pits are shown as rectangles. EM slush thickness shown as lines. B) Scatter plot of EM versus manual slush thickness.



4 Discussion

The spatial snow accumulation pattern from the drone-based digital elevation model closely match the characteristic snowdrift geometries identified from previous airborne laser scanner observations in Atka Bay, including pronounced windward and lateral snowdrifts and snow-free tails in the lee of icebergs incorporated in fastice (Franke et al., 2025). They are also consistent with numerical simulations of iceberg-induced snow redistribution, which reproduce the same characteristic deposition and erosion patterns around icebergs (Hames et al., 2026). Observed snow-thickness anomalies are controlled by persistent wind-driven redistribution of snow around icebergs.

Although our observations represent only a late-season snapshot of the snow and sea ice cover, they allow for a plausible seasonal evolution of the ice and snow system to be inferred. The observed ice composition preserves the cumulative imprint of thermodynamic and snow-related processes throughout the growth season, allowing key stages of sea ice development to be reconstructed:

Our observations show that flooding of the basal snow layer was considerably less deep and less extensive than would be expected from the high snow loads, if the underlying ice thickness would match the thickness observed in undisturbed areas of Atka Bay. Instead, thick ice was found at locations with several meters of snow, and slush thickness in the snow drifts did not exceed 13 cm. Sea ice cores from these heavily snow-covered areas revealed substantial snow-ice formation. This indicates an efficient conversion of flooded snow into snow-ice by refreezing throughout the winter, thereby adding ice mass from above rather than through basal accretion. Consequently, snow loading contributed directly to the sea-ice mass budget through snow-ice formation. In turn, the snow thickness measured towards the end of the sea ice season is lower than the snow accumulation, as a portion of the accumulated snow has been transformed into ice.

The insulation properties of the thicker snow and the diminished temperature gradient by flooding limit conductive heat flux through the ice, slowing bottom ice growth early in the season. However, simultaneous flooding and refreezing transfer part of the snowpack into the ice column, locally compensating this reduction in basal ice production. Similar processes have been described for Antarctic pack ice, where snow-ice formation represents a significant component of the seasonal ice mass balance (Arndt et al., 2021; Tian et al., 2020; Arndt et al., 2017; Jutras et al., 2016; Jeffries et al., 2001).

The mean ice thickness of 1.9 m within the snow-free tail downstream of the iceberg is 30 cm above the bay-wide mean (1.63 m) and is well supported by drill hole measurements. In contrast, the estimated bottom-grown thickness beneath the snow drifts (0.85–1.19 m) is substantially below the bay-wide mean (1.63 ± 0.17 m) and should be regarded as a speculative modeling result. In Atka Bay, the snow-free tails in the lee of icebergs were generally more extensive than the snow-accumulation zones, thus the enhanced basal ice growth in the tail may compensate for, or even exceed, the reduction in bottom-grown ice beneath the drifts. However, this interpretation remains tentative because the full downstream extent of the snow-free tail was not surveyed, and the conversion of DEM surface elevations into bottom-grown and snow-ice thickness is constrained by only a few manual validation points and performed only for a single iceberg. Increased snow accumulation was observed surrounding nearly all icebergs in Atka Bay, supporting that icebergs act as local snow sinks by reducing wind speeds and promoting snow deposition. Consequently, these areas experienced enhanced flooding and subsequent snow-ice formation. This spatial



relationship is consistent with previous studies that documented elevated snow accumulation around icebergs in Atka Bay through airborne laser and radar measurements (Franke et al., 2025) and simulations (Hames et al., 2026). The sea ice mass balance resulting from reduced basal growth beneath snow drifts and enhanced growth in the tail for icebergs of variable sizes and geometries remains an open question.

Hames et al. (2026) derived a log–log relationship between total snowdrift area and iceberg area. Applying this relationship to the iceberg inventory for Atka Bay in 2025 suggests that snowdrift area is equivalent to approximately 1.5% of the total fastice area in the bay. However, this relationship does not account for the elongated bare-ice zones with little or no snow accumulation observed downstream of icebergs. Our field observations suggest that these snow-free tails may be spatially extensive and could represent a substantial additional fraction of the sea ice area in Atka Bay. Enhanced snow accumulation is also observed near the western ice shelf edge, where snow accumulation on the sea ice reached the level of the adjacent ice shelf in some locations. Given the widespread occurrence of icebergs in landfast sea ice along the coast of Dronning Maud Land, these processes are likely not restricted to Atka Bay but may represent an important mechanism influencing sea ice thickness over larger regional scales (Arndt et al., 2024; Jeffries et al., 1997; Wever et al., 2021).

Although GEM-2 surveys and the subsequent inversion of the data using a 4-layer sub-surface model successfully captured the general spatial variability of ice thickness, they underestimated total thickness in regions with several meters of snow-ice formation and overestimated the slush thickness. This bias most likely results from the complex dielectric properties and stratigraphy of snow-ice, and the large total thickness and therefore reduced instrument sensitivity which complicate electromagnetic thickness retrievals under these extreme conditions.

The observations presented here provide field constraints on small-scale iceberg-induced snow redistribution and snow-ice formation, processes that are currently not represented explicitly in most sea-ice models (Vancoppenolle et al., 2025). They indicate that snow loading around icebergs in fastice can efficiently increase total sea ice thickness through snow-ice formation while simultaneously suppressing basal ice growth. The resulting redistribution of ice production highlights the complex interactions between atmospheric forcing, snow accumulation, and ice growth. Beyond its effects on ice growth, snow loading and subsequent flooding can modify the biological habitat within Antarctic sea ice by creating seawater-infiltrated snow and slush layers that support ice-algal communities (Arrigo et al., 2009; Fritsen et al., 1994; Swadling et al., 2023). The absence of snow in the tail further provides an environment with increased light availability beneath the ice compared with snow-covered areas, potentially influencing the growth conditions for sea ice microalgae (Gosselin et al., 1986).



5 Conclusions

Icebergs incorporated in Antarctic landfast sea ice substantially modify snow redistribution, producing persistent windward and lateral snowdrifts together with snow-free tails in the lee. Our combined drone photogrammetry, electromagnetic induction
360 sounding, and drilling survey, together with snow pit and ice core analyses demonstrate that these snowdrifts strongly influence local sea ice mass balance. Despite several meters of snow accumulation, only limited slush was observed beneath these snowdrifts. Instead, extensive snow-ice formation produced consolidated ice thicknesses of up to 7.8 m, indicating that flooded snow efficiently refroze during Antarctic winter and contributed to upward ice growth.

The observations further show that thick snow and flooding must suppress thermodynamic bottom ice growth beneath the
365 snowdrifts strongly, reducing basal grown ice to 0.65 m in one location compared to 1.60 m in the surroundings. This local reduction is contrasted by enhanced basal growth in the snow-free tail, where ice thicknesses were 1.9 m. Together, these findings demonstrate that iceberg-induced snow redistribution alters ice production in fastice. Because icebergs are widespread around the Antarctic coastline, these processes may represent an important local control on the structure, mass balance and ecology of Antarctic landfast sea ice.



Data availability. The data is currently being prepared for submission to PANGEA. DIOs will be included in the manuscript as soon as they are available.



Appendix A: DEM validation

375 Validation of the DEM was performed using three independent sets of manual snow thickness measurements. First, snow thickness was measured at the 16 drone ground-control points (GCPs). Second, snow thickness was measured at drill holes. Third, snow thickness was measured along-track two profiles in lateral and windward direction of the iceberg.

During DEM generation, the snow thickness measured beneath the GCPs was used to reference the GCP elevations to the local snow–ice interface. Ice freeboard was not measured at the GCP because the photogrammetry survey was performed prior
 380 to the main sampling campaign and invasive drilling was avoided to prevent disturbance of the site. Consequently, the resulting DEM elevations are referenced to an inferred ice-surface elevation rather than to a common absolute elevation. This approach assumes that spatial variations in ice freeboard are relatively small and occur over longer spatial scales than variations in snow thickness.

Across the independent manual datasets, the error statistics indicate good agreement, with RMSE values <17 cm and absolute biases <11 cm. The small negative bias, indicating that DEM-derived snow thicknesses are lower than the corresponding
 385 manual measurements on average, may be related to systematically lower ice freeboard at the locations of the manual measurements compared with the GCP locations. The agreement between the DEM-derived snow thickness and the independent measurements from drill holes and snow-thickness profiles nevertheless supports the assumption that spatial variations in freeboard are sufficiently small relative to the observed variations in snow thickness.

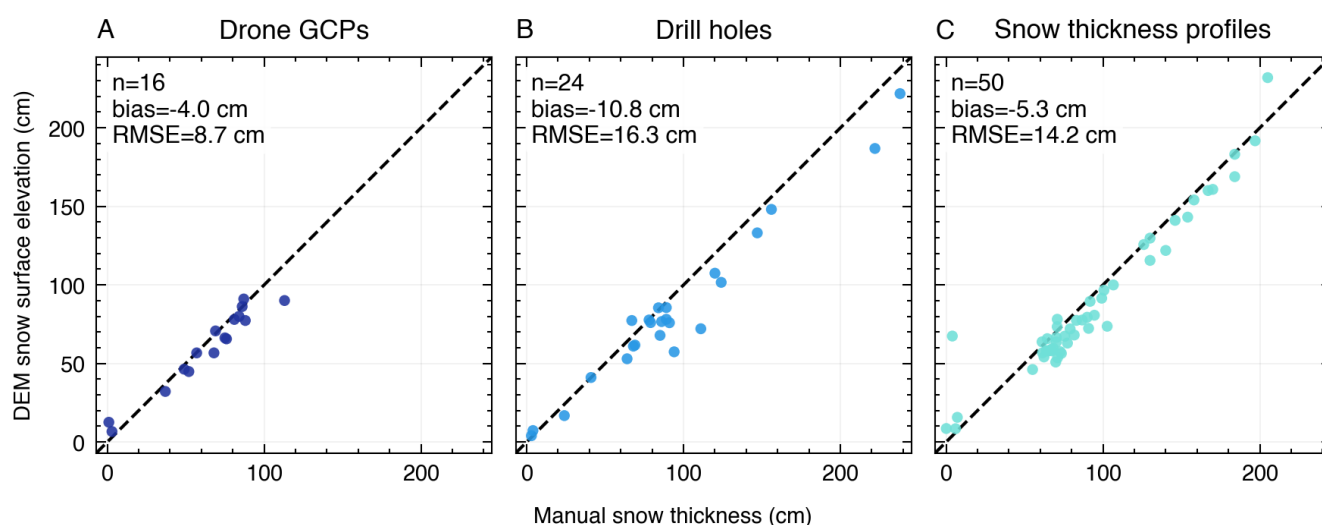


Figure A1. Comparison of DEM snow surface elevation and manually measured snow thickness for (A) drone ground-control points, (B) drill holes, and (C) snow thickness profiles. Points represent individual snow thickness observations, while the dashed line indicates the 1:1 relationship corresponding to perfect agreement between DEM and manual measurements. Statistics shown in each panel include the number of observations (n), mean error (bias, DEM minus manual), and root-mean-square error (RMSE).



390 Another source of uncertainty is horizontal co-location: the drone target locations were surveyed with high-accuracy GNSS, whereas drill hole positions and snow profile coordinates were recorded with a handheld GPS. The resulting geolocation uncertainty can introduce additional mismatch when sampling the DEM at point locations, particularly where surface elevation gradients are strong.



Appendix B: Photographs illustrating the sea ice core texture

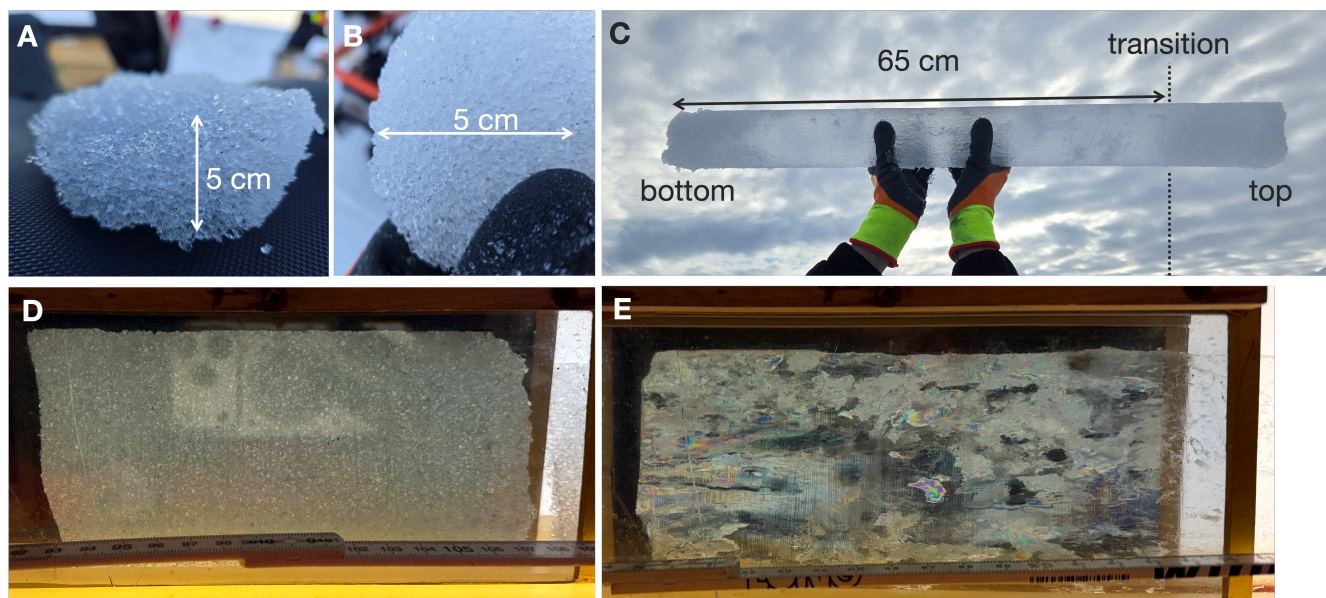


Figure B1. Photographs illustrating the sea ice core and its textural transition. (A and B) Close-up of a center section of the core, showing coarse ice crystals with characteristic rounded shape (orbicular granular texture). (C) Bottom core segment highlighting the transition from the clearer ice at the bottom to more opaque upper ice 65 cm from the bottom. The dashed line marks the visually identified transition between the two ice types. (D and E) Photos of vertical thick-sections (3-mm slices) between crossed polaroids on a light table for (D) a core segment from 100-120 cm (snow-ice, orbicular granular), and (E) a core segment from 344-361 cm (columnar texture).



395 *Author contributions.* SvdV, DP, and MN conceptualized the study and conducted the investigation. MN performed the data analysis. SA and CH contributed to the investigation by ice core processing. MN prepared the original manuscript. All authors contributed to the interpretation of the results and reviewed and edited the manuscript. SA and CH supervised the study and acquired funding.

Competing interests. CH is an editor of The Cryosphere. All other authors declare that they have no competing interests.

Disclaimer. TEXT

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