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Fractures and Floods: Small-scale spatial and temporal processes driving change in a relic ice-shelf basal channel

Journal:	<i>Journal of Glaciology</i>
Manuscript ID	JOG-2026-0145
Manuscript Type:	Article
Date Submitted by the Author:	27-Sep-2026
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Keywords:	Ice shelves, Ice/ocean interactions, Ground-penetrating radar
Abstract:	Basal channels, which form beneath ice shelves due to localized warm or fast-moving water, may weaken ice shelves by localizing deformation that promotes crevasse formation. Here, we test this hypothesis and investigate the role of fracturing and melt generation in the evolution of an ice shelf basal channel. We use ground-based ice-penetrating radar imaging of change over 3 years for the prominent ``relic'' basal channel

	in the vulnerable Thwaites Eastern Ice Shelf. We find an interplay of unsteady melt dynamics, ice deformation, and fracturing in a channel that is responding to the loss of connection to its subglacial discharge source. The channel exhibits an asymmetric 'M' shape that we infer results from local melt plumes on sidewalls under weak along-channel currents. Compression in the ice and wind-driven snow redistribution may be responsible for an offset between the channel and its associated surface depression. From observed basal fracture features, we deduce that the stress regime promoted fracturing from approximately 2010 through 2015. These crevasses likely act as sources (floods) or sinks of meltwater, generating locally variable channel geometry. The channel stress regime, however, inhibits crevasses from crossing the channel. These processes, key to ice-shelf stability, are largely muted in satellite analyses.



Fractures and Floods: Small-scale spatial and temporal processes driving change in a relic ice-shelf basal channel

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ABSTRACT. Basal channels, which form beneath ice shelves due to localized warm or fast-moving water, may weaken ice shelves by localizing deformation that promotes crevasse formation. Here, we test this hypothesis and investigate the role of fracturing and melt generation in the evolution of an ice shelf basal channel. We use ground-based ice-penetrating radar imaging of change over 3 years for the prominent “relic” basal channel in the vulnerable Thwaites Eastern Ice Shelf. We find an interplay of unsteady melt dynamics, ice deformation, and fracturing in a channel that is responding to the loss of connection to its subglacial discharge source. The channel exhibits an asymmetric ‘M’ shape that we infer results from local melt plumes on sidewalls under weak along-channel currents. Compression in the ice and wind-driven snow redistribution may be responsible for an offset between the channel and its associated

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surface depression. From observed basal fracture features, we deduce that the stress regime promoted fracturing from approximately 2010 through 2015. These crevasses likely act as sources (floods) or sinks of meltwater, generating locally variable channel geometry. The channel stress regime, however, inhibits crevasses from crossing the channel. These processes, key to ice-shelf stability, are largely muted in satellite analyses.

INTRODUCTION

Ice shelves, which form where grounded glaciers or ice sheets flow into the ocean and begin floating, provide a vital buttressing effect that slows the rate at which ice flows into the ocean (Thomas and MacAyeal, 1982; Dupont and Alley, 2005). When these shelves collapse, the glacier terminus can become unstable and rapidly retreat, especially if the bed slopes downward in the upstream direction (a retrograde slope, Mercer, 1978; Gudmundsson, 2013). We can improve our ability to predict ice sheet instability and sea level rise in the coming decades and centuries by better constraining the factors that contribute to the rate and magnitude of ice-shelf breakup.

The Thwaites Eastern Ice Shelf (TEIS, Figure 1a) is the floating extension of the Florida-sized Thwaites Glacier. In the next century, Thwaites Glacier is likely to be the largest single-glacier contributor to sea-level rise (Scambos and others, 2017). TEIS has experienced relatively high thinning rates compared to other Antarctic ice shelves in the recent past (Adusumilli and others, 2020; Milillo and others, 2019); although current rates are lower (Wild and others, 2022). Additionally, TEIS is vulnerable to changes in atmospheric and oceanic circulation due in part to its exposure to relatively warm modified Circumpolar Deep Water (e.g. Reese and others, 2018; Dotto and others, 2022). TEIS is presently weakening and on the verge of breakup (Sharp and others, in review; Benn and others, 2022; Alley and others, in review a). The magnitude of upstream acceleration due to its weakening and ultimate breakup is uncertain (Gudmundsson and others, 2023), but recent studies have shown measurable changes upstream of the grounding zone (Wild and others, in prep). The rapid changes in TEIS, along with recent field efforts through the International Thwaites Glacier Collaboration, offer an opportunity to study ice-shelf stability in more detail. Here, we focus on the most prominent basal channel on TEIS (Figure 1a).

Ice shelf basal channels, or simply basal channels, can form wherever relatively warm or fast-moving

water locally impacts the ice shelf base, which either melts a channel in the base of the ice shelf or activates channelized flow in suture zones or shear zones, or some combination of these factors (Alley and others, 2016). The source is typically a subglacial discharge plume that incises the base of an ice shelf and funnels the water by melting an upside-down U channel into the ice (e.g. Drews, 2015). If that source water is diverted or reduced, the channel will respond by closing due to viscous processes, which Wild and others (in review) suggests happened to the prominent basal channel on TEIS. We designate this type of feature as a “relic” basal channel to differentiate it from basal channels still connected to their initial source.

When a basal channel incises into an ice shelf, the shelf responds both viscously (e.g. Drews, 2015; Wearing and others, 2021), and elastically (e.g. Humbert and others, 2022; Buck and Lai, 2021) to hydrostatic disequilibrium. The local ice thinning due to melt generates local stresses that promote crevasse formation both underneath and at the surface of ice shelves (Vaughan and others, 2012; Dow and others, 2018). As a source of locally high melt rates, fracturing, and thinner (and presumably weaker) ice, basal channels have been seen as a source of instability for ice shelves and grounding lines, including on TEIS (Chartrand and others, 2024). However, in some cases they are linked to stabilizing processes via regulating basal melting (Millgate and others, 2013), showing a need for a better understanding of basal channel dynamics.

The local ice thinning due to melt also produces a surface depression that is detectable from satellites and provides insights into basal channel behavior. Examples include channels responding to large-scale sub-ice-shelf circulation (Gourmelen and others, 2017), channels causing thinning and weakening of shear margins (Alley and others, 2019), and channels migrating due to Coriolis-driven asymmetry of melt inside channels (Alley and others, 2024). Flow, flexure, and fracture can all generate similar surface expressions (albeit on different timescales), meaning that the geometry of the surface depression conflates multiple key basal processes. Local snow deposition and redistribution processes further obscure direct links between surface topography and basal processes (Drews and others, 2020). This is especially important to consider when examining radar stratigraphy, as surface accumulation has a nine times stronger effect on stratigraphy than basal melting and refreezing (Drews and others, 2020).

Ground-based ice-penetrating radar (both monopulse and phase-sensitive radar systems) helps link surface topography with basal and englacial features to reveal important processes invisible to satellites. This is especially true when integrated with modeling. For example, Dow and others (2024) used satellite products combined with ice-penetrating radar to identify basal fractures and link them to surface features on the Nansen Ice Shelf. Drews (2015) used radar measurements and modeling results to investigate bridging

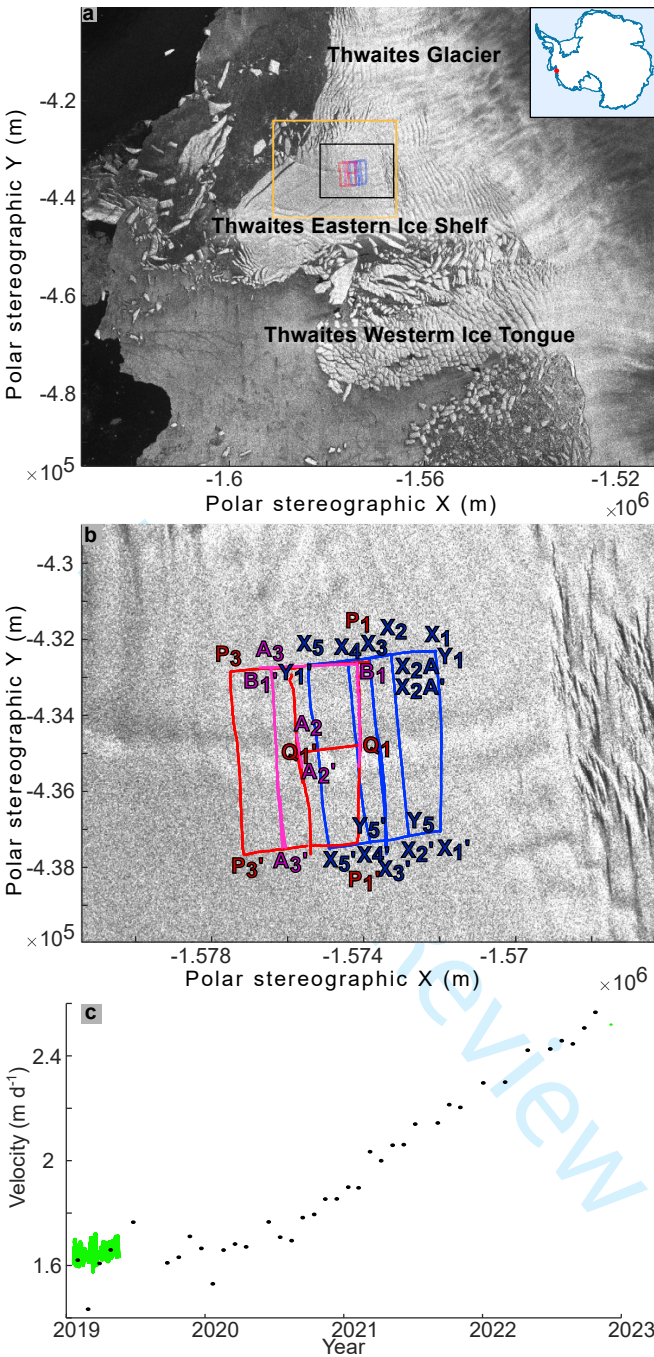


Fig. 1. Panel **a** shows a map view of Coastal Thwaites Glacier (Image is Sentinel-1A, 1/12/2020). The black box is the field site shown in **b**, and the larger yellow box is the bounding box of Figure 2. In **b**, blue lines correspond to 2019/20 100 MHz profiles, red lines to 2022/23 100 MHz profiles, and pink lines to 2022/23 20 MHz profiles. Only profiles used in figures are labeled. Profiles labeled X and Y are 2019/20 100 MHz, profiles labeled P and Q are from 2022/23 100 MHz, and profiles labeled A and B are from 2022/23 20 MHz. Panel **c** shows the velocity of the center of the field site from Sentinel-1 data (black) and AMIGOS data (green). Sentinel velocity is derived from Banerjee and others (2025).

effects over a channel on the Roi Baudouin Ice Shelf.

Here we use radar-derived basal topography and englacial-layer patterns to investigate processes affecting a ~ 4 km reach of the TEIS “relic” channel. We present evidence that 1) the channel is closing from viscous deformation; 2) ambient melt plumes are active on channel walls; 3) crevasses have formed as recently as 2015, but not since then; 4) the surface depression created while the channel was active has since been modified by transverse compression and asymmetric infilling of snow; and 5) the channel inhibits propagation of basal crevasses. Together, these intertwined processes shape the channel geometry and likely play a role in the stability of the ice shelf.

METHODS

Study area

We use field data collected over a ~ 4 km reach of the relic basal channel on TEIS free of visible surface crevassing (Figure 1). Across two field seasons, we collected data in a Lagrangian framework: returning to the same ~ 4 km $\times$ 5 km area of ice, accounting for ice flow using satellite-derived velocity fields over the three-year gap between field seasons in 2019/20 and 2022/23 (Figure 1). While we placed flags at numerous points along the radar paths to re-occupy the same profile, COVID-19 delayed our field season, and many flags were buried, necessitating post-field analysis discussed later in the Methods section.

Figure 2 shows the annual flow velocity and along-flow surface strain rate composites for 2020, gridded to 120 m, created using data from Banerjee and others (2025). Banerjee and others (2025) derived the velocity components by stacking all available ITS_LIVE image-pair global ice velocities and taking the median grid value for every year. They also derived the longitudinal, transverse, and shear strain rates from the velocity records. Using these, we calculate the principal minimum/maximum compression composites assuming plane strain. The plane strain assumption is useful but imperfect because of small but non-zero vertical strain rates.

Radar data collection

We used two monopulse radar systems: GSSI SIR4000 with 100 MHz shielded antenna (both seasons) and Blue Systems IceRadar with 20 MHz unshielded antenna and a Kentech transmitter (second season). The 100 MHz system, which has the smaller wavelength, allows for more detailed imaging of the near-surface internal layers as well as higher spatial resolution of basal topography. However, due to system electronic

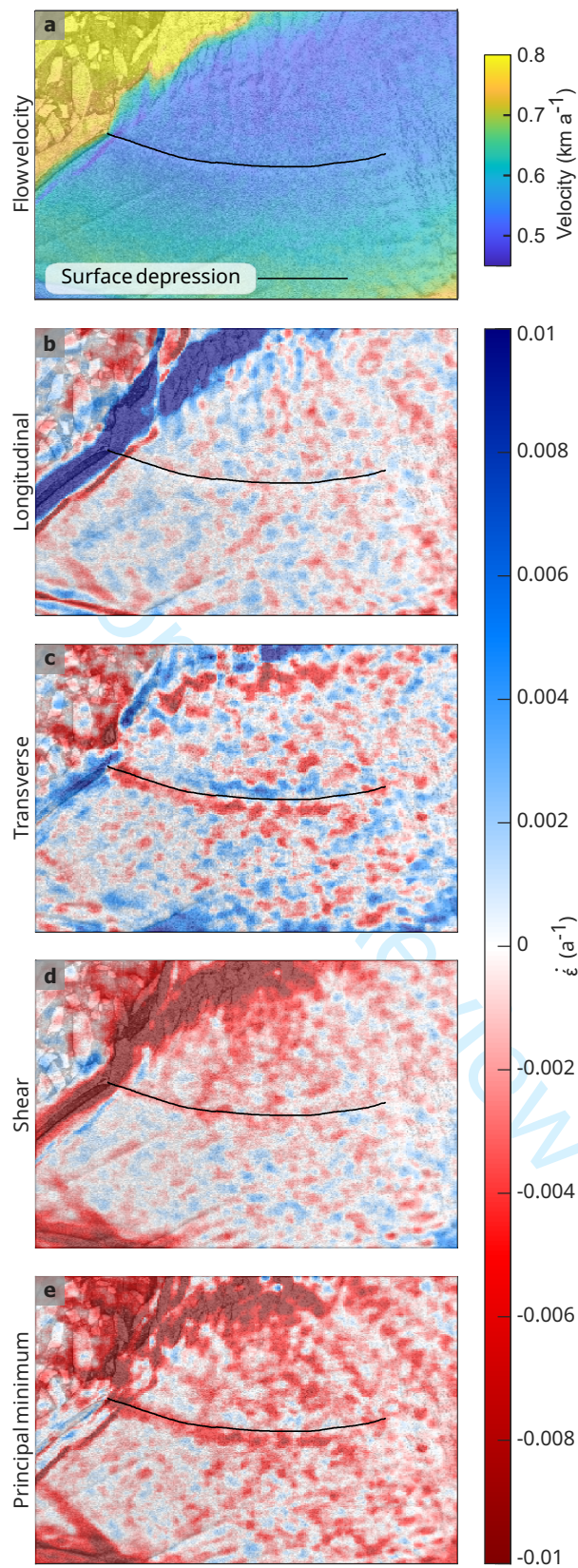


Fig. 2. Annual flow velocity and strain rate composites, gridded to 120 m, over the field site from 2020, created using data from Banerjee and others (2025). The bounding box can be seen in yellow in Figure 1a. Panel **a** displays velocity composites, **b** shows along-flow longitudinal strain rates, **c** shows along-flow transverse strain rates, and **d** shows along-flow shear strain rates. Panel **e** shows principal minimum (maximum compression) strain rates. The black lines in each panel are surface depression picks.

limitations (e.g., slower analog-to-digital conversion) and higher attenuation in the ice, the 100 MHz system is less capable of resolving deep englacial features such as basal crevasses. The 20 MHz system has a lower spatial resolution due to longer wavelengths, but also a lower attenuation and a faster digitizer to better resolve englacial structures.

For accurate surface topography to interpret the radar data, we use precision Trimble R7 GNSS receivers processed as 1-second kinematic data using the Canadian Spatial Reference Systems Precise Point Positioning (CSRS-PPP) service, using the ITRF2000 reference frame. Given sled mounting and variable surface conditions, the horizontal and vertical uncertainties of these data are ± 0.05 m and ± 0.10 m, respectively. The relative uncertainty between nearby points along a profile is lower. Three identical GNSS receivers collected data on different snow machines and radar sleds to have redundancy in case of power loss or other issues. In post-processing, we realized that some GNSS data had high errors that we linked to a damaged antenna; these data have been removed or used with caution, depending on available redundant data. Vertical GNSS positions were corrected from ellipsoid to local sea level using geoid heights from EGM96. We corrected for antenna heights, tides, and inverse barometric effects using ERA5 reanalysis data with the same process as outlined in Wild and others (2022) that uses the Circum-Antarctic Tidal Solution developed by Padman and others (2008).

To further ensure the quality of our surface elevation measurements, we compared our values in each season to those taken from NASA's ICESat-2 satellite (Smith and others, 2025), acquired via OpenAltimetry (Khalsa and others, 2020). We compared our elevation values to ICESat-2 points within 6 meters of our elevation points and within two months before or after our field season. Results show that our elevation values are, on average, $0.18 \pm .70$ m higher compared to ICESat-2 elevations in the 2019/20 field season, and on average $0.15 \pm .84$ m lower in the 2022/23 field season. These results show no significant bias in our data compared to ICESat-2, and differences in elevation could be due to variations in surface conditions over short periods of time and space.

Finally, we compare the final GNSS-derived surface topography in 2019/20 to TanDEM-X surface topography from February 8, 2020, which was vertically adjusted to nearby ICESat-2 elevations (Figure S1). While the stated vertical accuracy of TanDEM-X DEMs in this area is lower than that of our precision GNSS, the more complete spatial coverage provides a check for our sparse surface elevation data and a method to fill small gaps in our data. For the 2022/23 season, available TanDEM-X imagery is more limited. Given the rapid changes in the ice shelf, we did not use TanDEM-X to fill gaps in field data in 2022/23.

143 Instead, when necessary, we use a taut-spline or linear interpolation over small gaps.

144 **Radar processing**

145 We apply standard radar processing steps (similar to that described in Lilien and others, 2020), using a
146 customized version of matGPR (Tzanis, 2010), which includes stacking, filtering, and applying gain. When
147 necessary for visual interpretation, we perform additional processing to remove ringing artifacts present in
148 the near-surface that obfuscate internal ice layers.

149 *Travel time to depth conversion*

150 The ice thickness varies significantly over short distances within each profile; therefore, we use an iterative
151 approach to convert the radar returns from travel time to depth. We start with an estimate of a mean
152 density and associated radar wave speed for each point along the radar profile. We assume, based on hot
153 water drilling (Scambos and others, 2025), that the pore close-off is approximately 45 m deep, which is a
154 significant percentage of the ice thickness (250–400 m). We assume, based on firn core density measurements
155 (MacLennan and others, 2023), that the near-surface density is approximately 450 kg/m³. We use the
156 exponential approximation for steady-state firn densification recommended by Cuffey and Paterson (2010)
157 to calculate mean density as a function of ice thickness. We convert to mean wave speed according to Robin
158 and others (1969) using a dielectric permittivity of 3.15 for pure ice. Given uncertainties in our surface
159 density and pore close-off depth, our wave speeds are approximately 172.4 ± 0.4 m/ms (for 300 m thick ice).
160 For each point along the radar profiles, we iterate between initial ice thickness estimate, mean density, wave
161 speed, and updated ice thickness to converge on an ice thickness that best matches the density profile and
162 two-way travel time.

163 For englacial layers, a wave speed based on depth-averaged wave speed is biased: near-surface layers
164 appear shallower than the true depths of these reflectors. Figure S2 shows the offset for reflections as a
165 function of depth due to the use of depth-averaged wave speed. We estimate that the maximum offset in
166 depth due to this effect occurs at approximately 55 m depth; radar reflectors that appear 55 m deep have a
167 true depth of ~59 m. Above and below this depth, the offset due to the real depth profile of wave speed is
168 less. We take this into account in estimating depths of shallow features.

169 We use the conventional $\lambda/4$ estimate of the uncertainty of the depth of a reflector, where λ is the
170 central wavelength (Yilmaz, 2001), which depends on the radar system (approximately ±0.43 m for 100

MHz data and ± 2.14 m for 20 MHz). Uncertainty in the mean wave speed translates to ± 0.7 m depth at 300 m depth. Combining reflector depth and mean wave speed uncertainty in quadrature along with surface elevation uncertainty (± 0.1 m), the uncertainty at the ice-ocean boundary is ± 0.83 m for the 100 MHz data and ± 2.25 m for the 20 MHz data.

Migration

We interpolate the data along the profile to create equally spaced traces for migration. We use an F-K migration algorithm as implemented in matGPR (Tzanis, 2010). We use both migrated and unmigrated profiles for interpretation due to complex topography and off-nadir englacial features. All figures in this paper show unmigrated profiles. All figures in the supplement show unmigrated profiles with the exception of several 2022/23 100 MHz profiles, which are appropriately labeled.

Radar interpretation techniques

We present results integrating analyses of 17 profiles from the 2019/20 field season and 19 profiles from the 2022/23 field season (13 100 MHz and 6 20 MHz). The full data set is archived at Pettit and others (2026a) for 2019/20 data and Pettit and others (2026b) for 2022/23 data. For both field seasons, the transects cover a grid running just under five kilometers across the channel and just over three kilometers along the channel (Figure 1b). The subset of profiles that were repeated between field seasons is identified in Table S1.

Identifying ice-ocean interface and englacial reflectors

We use a semi-automated picking method to select points along the ice-ocean interface and the brightest englacial layers. This method works by identifying the highest energy reflector within a human-specified depth range. A human then removes outliers.

The most important englacial layer for our analysis is a particularly bright layer about 30 m deep (near the bottom of the firn) seen in all profiles in both field seasons. We use this “bright layer” as a reference layer to compare profiles from different years to minimize uncertainty due to firn compaction. Given the depth and spatial contiguity of this layer, we infer that this layer was created due to katabatic-wind-related warming and surface melt at the change in surface slope at the grounding zone; thus, it is a local isochrone, forming when each region of ice travels through the grounding zone. It allows us to separate new snow and firn deposited after the ice crossed the grounding zone (where the channel forms) from that deposited

farther upstream; thus, this layer provides a method to constrain down-warping of layers associated with basal melt. Its role as a local reference, however, does not depend on its formation mechanism.

Refining alignment of radar between field seasons

In order to compare radar profiles between field seasons in a Lagrangian sense, we refine the pre-field estimates for ice motion with post-field analysis. The ice shelf flowed approximately 2.5 km at an accelerating rate (Figure 1c) between field seasons. To quantitatively compare radar profiles between years, we estimate the spatial offset between profiles and, thus, define the sources of uncertainty in profile comparison. To do this, we first recalculated the ice flow using a more comprehensive time series of satellite and ground-based velocity fields. The ground-based velocity was derived from the precision GNSS mounted on a tower in the center of our study area (“AMIGOS” Station, Scambos and others, 2025, and Figure 1c); position data was uploaded daily via Iridium until the tower electronics failed. The tower was re-occupied during the 2022/23 season with a Trimble R7 for a 6-hour period and processed as kinematic data using the CSRS-PPP service. This 6-hour occupation was sufficient to detect the direction and speed of the ice. We use the direction, rotation, and velocity of ice shelf flow to determine spatial offsets between radar profiles. Because this region has low horizontal strain rates (Figure 2), we assume this $\sim 4\text{ km} \times 5\text{ km}$ reach moves and rotates approximately as a block.

Second, while we assume the ice in this region is moving as a block, we know it is deforming internally. The error in our block assumption increases with distance from the GNSS Station. We refine our alignment of specific radar profiles by assuming that basal crevasses within the ice are markers fixed to the ice and provide horizontal positioning between years. We limit the use of this final alignment method to within 500 m on either side of each crevasse. For profiles that were neither close to the AMIGOS station nor had crevasses to use as reference, we use the average displacement seen in nearby profiles that did have reference features to estimate their displacement.

Ultimately, we find that perpendicular offsets between any two along-flow radar profiles in different years are typically less than $\pm 5\text{ m}$. The parallel offsets are generally less than $\pm 100\text{ m}$, and closer to $\pm 10\text{ m}$ near crevasses. The across-flow profiles have potentially larger offsets between profiles, but still less than a few 10’s of meters. In interpretation, we recognize that even small offsets can result in large differences for steeply dipping layers.

Dating and analyzing englacial layers

In addition to the “bright layer”, we observe features in the near-surface internal layering. Regardless of the formation mechanism, distinct englacial features that are overlain by smoother and more continuous firn isochrones can be dated by using an approximate depth-age scale for the firn. We assume that the first smooth isochrone above a feature signals a change in the process affecting the firn layering. For example, the internal layering below the bright layer is often more chaotic than the layering above the bright layer. The age of the bright layer, therefore, signals a change in dynamics from a mechanism (such as surface crevassing) that causes chaotic surface topography to a mechanism that allows snow to drape smoothly across the surface.

To date an internal layer, we use the same firn age-depth model as Drews and others (2020), and assume a surface density of 450 kg/m^3 , a pore close-off depth of 45 m, and an accumulation rate of 1 m a^{-1} ice equivalent based on Copernicus ERA5 values. We recognize its simplicity (steady state, constant accumulation); however, it agrees well with observed firn-core values from MacLennan and others (2023). As accumulation rates may vary in time and space (especially within and outside the channel), our results should be viewed as approximate ages. A persistent 20% higher accumulation rate would lead to a 10 m deep firn layer being a full year younger (just 5 years old compared to over 6 years old). More complicated methods of firn modeling could resolve the age more accurately, but are outside the scope of this study.

Tracking the surface expression of the channel through time

To examine channel evolution on longer timescales, we digitized the surface depression centerline on Sentinel-1 images every three months from November 2014 through November 2023. Two people independently traced the channel over every image to reduce the impacts of human error. To isolate changes in channel behavior from advection due to ice shelf flow, we used ITS_LIVE yearly velocity mosaics to advect earlier surface depression locations to their estimated position in 2023.

RESULTS

The surface of the ice shelf in our study area is smooth with gently undulating topography, typically 500 m across and less than a few meters deep. In the field, we found no evidence of surface crevassing, either visually or by surveying near-surface layers with high-frequency radar for safety. The satellite-derived strain

rates show no significant horizontal deformation (Figure 2), despite this section of the ice shelf accelerating at rates as high as 150 m a^{-2} . The speckled pattern in the strain rate fields is a sign that we are near the strain-rate resolution of the method. We do not smooth these data further, however, because within this minimally deforming region, Figure 2c,e show small but measurable transverse compressive strain along the relic channel. In sum, such a minimally active area from the surface belies complex 3D basal topography and internal structures (Figure 3). Figure 3 shows selected cross sections of the channel from the 1920/20 field season, with similar features appearing in the 2022/23 field season. The columns represent upstream, middle, and downstream locations along the reach. Figures 4 shows examples of specific basal and englacial features used in our analysis. Figure 5 shows details of structure within the near-surface layers along cross sections of the channel relative to the basal channel geometry. All profiles used in this study, including a map of profile locations, are provided in Figures S3 to S37.

Basal channel geometry

Basal channel cross-sectional geometry

The most upstream (southernmost) channel cross-section is X1 (Figure 3d,g). It shows a narrow, deeply incised channel that is 80 m tall (nearly one third the ice thickness) and less than 500 m wide at the base. X1 and the corresponding 2022/23 profile P1 (Figures S24-25) display a smooth upside-down U shape.

The most downstream (northernmost) profile (X5), on the other hand, exhibits an asymmetric ‘M’ shape cross section (Figure 3f,i), with the western peak taller than the eastern peak. Of the five transects showing the M-shaped channel (Figures S9-S11 and S23-S28), the difference in elevation between the western (taller) and eastern (lower) local maxima forming the M ranges from 9 to 15 meters.

In between these two locations that are just over 3 km apart, the channel cross sections show an apparent transition between the U and M shapes, with significantly less pronounced M shapes in cross sections X2 and X3 (Figures S5-S8 and S9) compared to the middle cross section X4 (Figure 3e,h), and X4 having a less pronounced M than downstream cross section X5 (Figure 3f,i). Similarly, the corresponding 2022/23 upstream cross-sections display no apparent M shape (P1, Figures S21-S22), while the downstream (P3, Figures S26-28) shows a strong M shape.

Figure 3d, e, and f show the near-surface radar layers corresponding to 2019/20 channel cross sections X1, X4, and X5, respectively. The surface depressions associated with this basal channel are approximately 3.7 m deep and 1500 m wide, and have uplifted “shoulders”. The surface depression in all profiles is offset

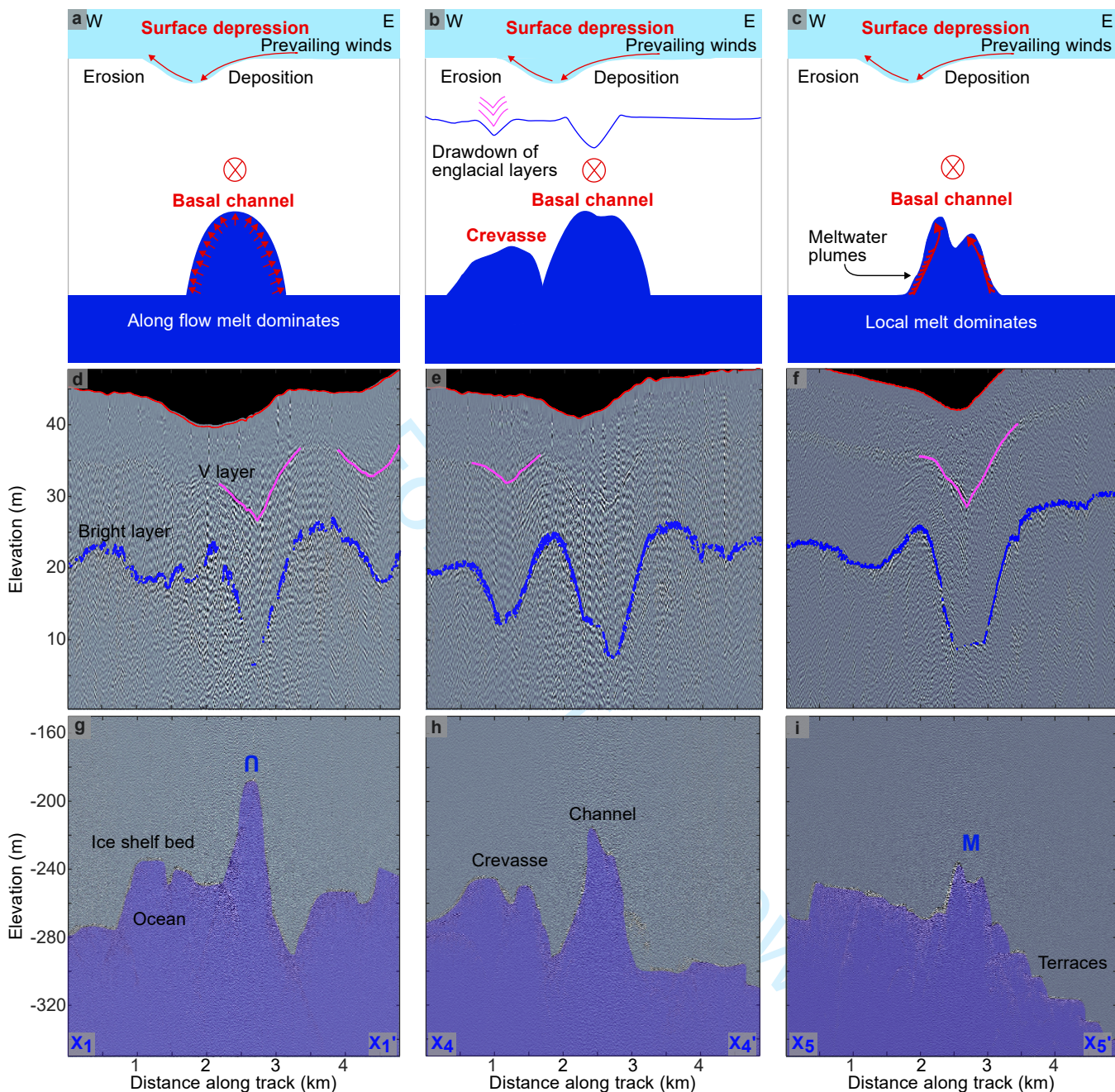


Fig. 3. Unmigrated 2019/20 radar profiles showing characteristic basal channel geometries, englacial layer patterns, and surface topography. The top row (a-c) shows cartoons of hypothetical processes shaping observed features. The central row (d-f) shows near-surface englacial layers and surface topography (red). The V layer features discussed in the text are in pink, and the bright layer found across all profiles is traced in blue. The bottom row (g-i) displays bed topography including the basal channel. The ice is shaded in gray, and the water below the ice shelf is shaded in blue. From left to right in each row, panels show progressively more down-ice flow geometries. The view for all panels is downstream.

to the west by 200 to 400 m from the basal channel apex (approximately one ice thickness). In addition to being offset from the channel, the surface depression does not reflect the M shape of the basal channel, despite shallow englacial firn layers, including the “bright layer” (in blue), exhibiting a mirror image of the basal M shape in most cases.

Basal channel 3D geometry

Beyond the transition in shape of the channel from U to M, the geometry shows spatial variation along the reach. The channel apex is higher in elevation at the upstream end and is progressively lower toward the downstream end (Figure 3g,h,i and S39). Buoyant meltwater flowing along the channel from upstream to downstream along this increasing hydraulic potential means the water is flowing “uphill” (against the hydraulic potential). The mean slope of the channel apex is ~1.4% (Figure S39b,c). In both years, the gradient in the hydraulic potential is slightly shallower in the first half of this reach and slightly steeper in the second half. The channel also shows progressive widening in the downstream direction (by approximately 150%), as seen in Figures 3g,h,i and S39a.

Other related features in radar data

While the basal channel is the most prominent feature in this area of the ice shelf, several other features make up the ice shelf’s basal topography and englacial structure that are relevant for our analysis. Figure 4 summarizes these features. Our identification of features relies on the following assumptions:

1. Internal layer reflectivity is due to spatial variations in real physical properties of the ice (density, crystal fabric, impurities), and thus the pattern of reflectivity will be similar from year to year. These variations in ice properties that are associated with snowfall are isochrones.
2. Water-filled basal crevasses appear in radar as strong reflectors with a geometry that depends on the angle of the radar profile with respect to the strike of the crevasse. The reflector geometry can vary from stacked hyperbolas (across-strike direction) to a horizontal line mimicking the bed (along-strike).
3. Crevasses that are partially refrozen may appear as weak reflectors or diffuse scattered regions due to the absorption and scattering properties of slushy brine (Badgeley and others, 2017).
4. For radar profiles capturing crevasses in the along-strike direction, the crevasse may be indistinguishable from a normal bed reflector without additional information. The additional information we use includes

309 nearby radar profiles imaging the crevasse from a different direction or with a different radar system
 310 that can detect weaker englacial features.

311 *Crevasses approaching channel from the side*

312 Satellite images (such as Figure 1a,b) show linear surface depressions perpendicular to the channel's surface
 313 depression. They are more prominent on the true east side of the channel. The radar profiles B1, Y1, and
 314 Y2 across these features (Figures 4a,b,S12,S13,and S36) show large undulations in the bed reflector, a notch
 315 or missing reflector at the apex, and stepped-terrace topography on the flanks of each undulation. B1 and
 316 Y1 (Figures 4a,b and S12) show large undulations in the bed reflector, a notch or missing reflector at the
 317 apex, and stepped-terrace topography on the flanks of each undulation. Along with surface topographic
 318 depressions, down-warped internal layers, and (in 20 MHz radar) stacked hyperbolas, form the signature of
 319 basal crevasses modified by localized melt (Wanzer and others, in prep).

320 *Stepped terraces on flanks of channel*

321 Dutrieux and others (2014) found stepped-terrace features (basal terraces separated by steep ice slopes)
 322 along the flanks of a basal channel in Pine Island Glacier (Antarctica). Rignot and others (2025) found
 323 similar features along the flanks of a basal channel in Petermann Glacier (Greenland). We observe similar
 324 stepped-terrace features along the eastern flank of this relic basal channel, primarily where the flank extends
 325 deep in the water column at the downstream end of our reach (Figure 4g,h). These stepped-terrace features
 326 consist of flat or gently curving terraces that can be fifty to several hundred meters long. The vertical steps
 327 between them range from several meters to several tens of meters. Most of these terraces slope slightly down
 328 toward the east, away from the channel at an angle that ranges from zero degrees to ten degrees. There are
 329 a small number of terraces that slope toward the channel instead, with a maximum angle of five degrees.

330 *V-shaped internal layers associated with Basal crevasses*

331 Figure 4a-f show three pairs of 2022/23 20 MHz near-surface and basal radargrams. These show examples of
 332 basal crevasses (identified with black arrows in **b** and **d**) that are vertically aligned with “broken” V-shaped
 333 reflectors in near-surface internal layers (here, termed “V layer” and highlighted in pink). We find these
 334 V-layers throughout our study area, almost always aligned with a basal crevasse or the channel (Figure 6).
 335 Luckman and others (2012) noted similar down-warping of near-surface layers above basal crevasses on

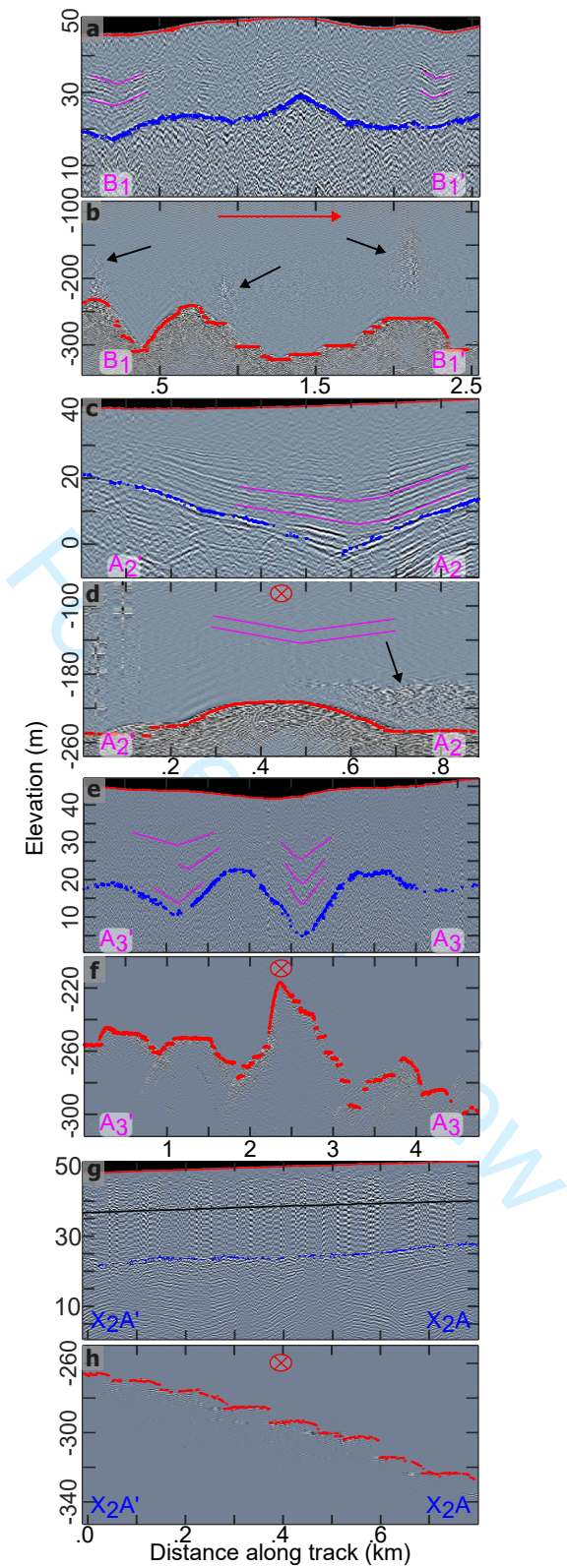


Fig. 4. Various basal and englacial features found within the field area. Note the different scales on each unmigrated radargram, selected to highlight features of interest. **a,b** and **c,d** from profiles B1 and A2: two sets of bed and near-surface radargrams showing basal crevasses (stacked hyperbolas identified with black arrows) collocated with buried V layers (pink). **e,f** from profile A3: particularly complex radargrams (bed and near-surface) across the channel (tallest feature), with evidence of stepped terraces along both flanks as well as multiple V-layer features. **g,h** from profile X2A: stepped-terrace features along the eastern channel flank, similar to those described by (Dutrieux and others, 2014).

the Larsen C Ice Shelf. We do not see the same pattern of surface troughs in our radar, possibly due to differences in crevasse size and height and/or snow infilling.

Figure 4a,b is a profile running parallel to the channel. V layers appear in the firn above two stacks of hyperbolas that likely represent basal crevasses. Panels **c** and **d** in this same figure cross the basal channel and show a feature, identified with a black arrow, that is likely a crevasse being imaged from the side (the radar profile follows the strike of the crevasse). This crevasse terminates above and slightly to the right of the basal channel apex. There are several V layers above that we associate with this crevasse; however, given the three-dimensionality of the ice structures in the region of the channel and the angle at which we have imaged the crevasse and the V layer, it is difficult to confirm the relation between these features.

Panels **e** and **f** show a V layer spatially aligned with the sharp apex of the channel; while there are no specific signs of a crevasse, this suggests that a crevasse may extend from the apex. In all of our radar profiles, V layers were uniformly associated with signs of basal crevassing or locations where we expect basal crevasses. Not all basal crevasses, however, show V layers above them. To our knowledge, the association of V layers with basal crevasses has not been described in the literature before.

To understand the role of crevasses in the evolution of the basal channel, however, we use the ages of formation of numerous V layers across both field seasons as a proxy for fracturing activity (Figure 6). V layers tend to appear in a vertical stack; therefore, we date the uppermost distinguishable V layer as the date the crevasse formed. Snow falling after the crevasse formed will fill topographic lows and lead to a transition from a sharp V isochrone to a smooth isochrone. For the shallowest V layers (youngest ages), the radar does not always resolve a smooth isochrone “capping” the stack of V layers due to near-surface blanking; thus, our date is a maximum age. The crevasses may be younger, but still old enough and covered in recent snowfall to be undetected by our safety assessment for the area. Across both field seasons, we find a median V layer deposition year of roughly 2012. The oldest crevassing event we identify in the radar is from 2002; additional older events may have occurred, but post-crevassing ice deformation may make the V layer shape difficult to identify. The most recent crevasses we were able to date are from 2015; however, as discussed above, these dates may be biased and appear older than they are. Our data are too sparse to have confidence in spatial patterns within the field site.

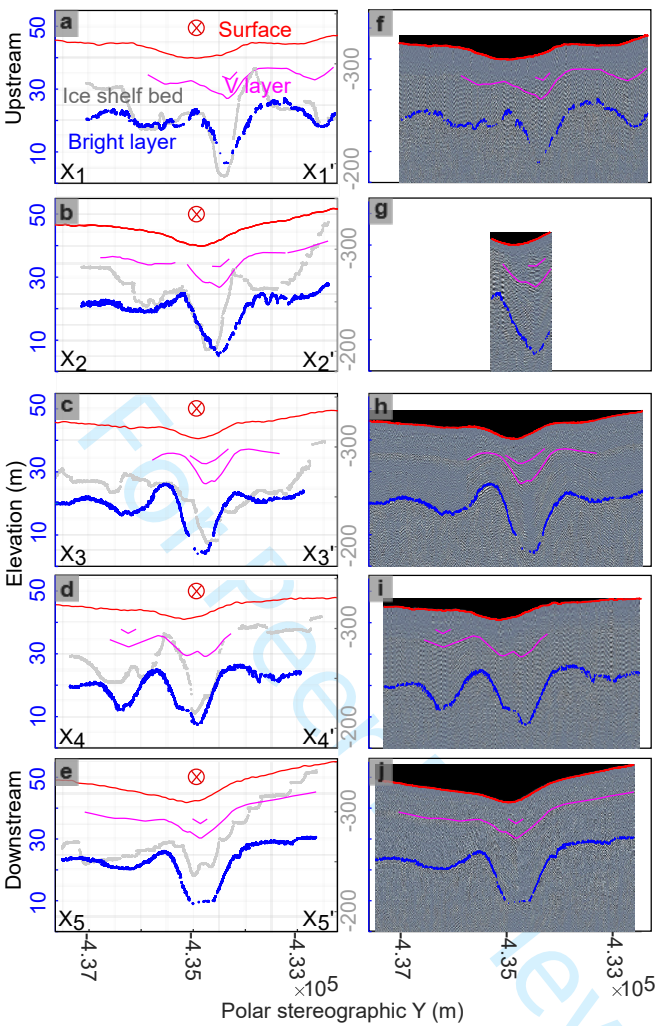


Fig. 5. Sections from the five unmigrated across-channel 2019/20 profiles (X1 to X5), from upstream (top) to downstream (bottom), with each profile oriented to look downstream. Both columns display the same profiles, with the left column displaying solely the surface topography (red), V layers (pink), and bright reference layer (dark blue), as well as the basal topography flipped upside down (light gray) to compare with surface and englacial geometry. Note the scale on the right axis (light gray data) spans three times more depth range than the scale on the left axis. The right column displays the same information, but without the basal topography and with the radargram included. Each profile displays how the surface depression is offset to the west of the basal channel and bright reference layer. The bright reference layer very closely mirrors the basal channel geometry.

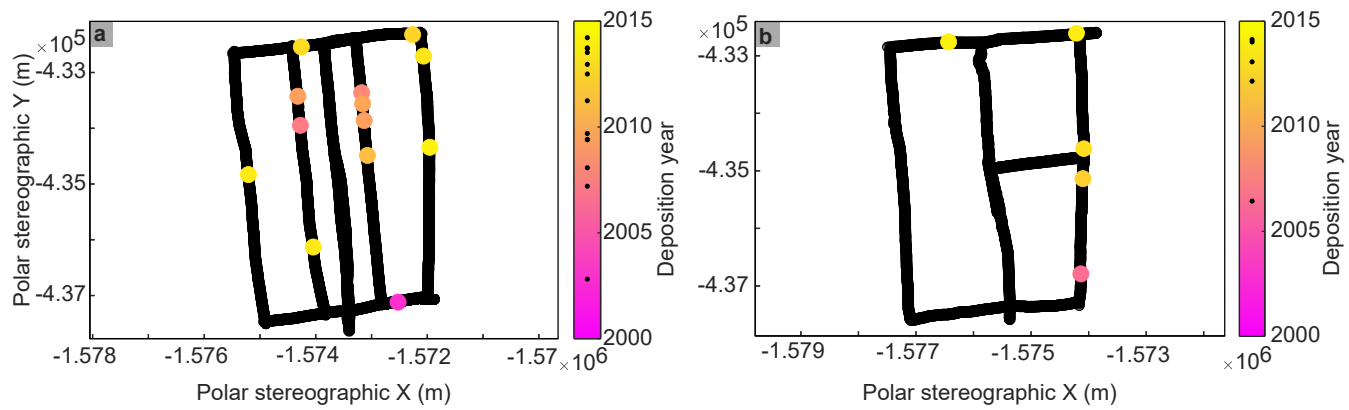


Fig. 6. Map views of the estimated ages of the V layers observed in 2019/20 (a) and 2022/23 (b). Each colored dot represents the age of the nearest-surface V layer observed in that location. The locations of all radar profiles are shown in black. The ages of the crevasses are also shown as black dots on the color bar to better show their distribution.

The relation between surface topography, englacial layers, and bed topography

Stepping back out to the largest scale, for all basal channel cross sections (such as shown in Figure 3d,e,f, and particularly Figure 5c,h), the surface depression is offset from the basal channel by several hundred meters. The surface depression and the basal channel are situated in a more gently curving topography that is thicker on either side of the basal channel (Figure 3g-i). In an even broader context, the ice is thicker downstream of our study area (Figure S1), likely inherited from upstream. The subglacial discharge carved into this thicker ice as it crossed the grounding zone. We note this because the movement of water in complex topography depends on hydraulic gradients that may be influenced by the topography on this broader scale.

Whether englacial layers follow the surface topography or the basal topography provides clues to dominant processes. In our study area, all layers visible in the radar deeper than approximately 10 m align more with the basal topography than they do with the surface depression. Near-surface layers shallower than ~ 5 m (where they are visible) sometimes, but not always, follow the surface topography and surface processes. Figure 5 shows the near-surface layering compared to the basal reflector for five profiles, each crossing the basal channel at different locations. The left column shows selected radar layers: surface (red), the bright layer (blue), near-surface V layers (pink), and the basal channel topographic profile (gray, flipped upside down for easier comparison with the near-surface layers). In this sequence of figures, from upstream to downstream, we make the following observations: 1) in Figure 5a,b,c (the upstream cross sections) the

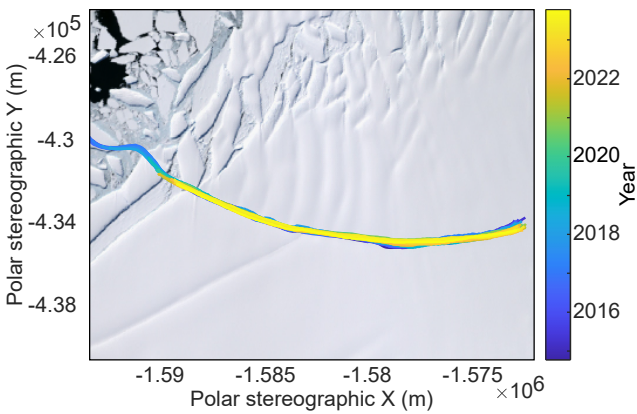


Fig. 7. Surface depression picks taken every three months from November 2014 to November 2023. In this figure, ice flows from right to left. All picks have been migrated to November 2023 to visualize changes in surface depression geometry independent of overall ice shelf flow.

bright layer mirrors the basal channel shape with the bright layer down-warped by approximately 1/3 the height of the basal channel; 2) in Figure 5d the heights of the two peaks of the M shape are reversed in the bright layer; 3) in Figure 5e the bright layer down-warping is much bigger in relation to the size of the basal channel height compared to previous profiles; despite the strong M shape of the basal channel, the bright layer has a muted M shape; 3) Figure 5a,b,e all have a sharp V layer at ~ 10 m depth, aligned vertically with the basal channel and offset from the surface depression; 4) Figure 5c shows a more rounded near surface layer that is along a trend between the bright layer and the surface depression, similar to what we expect for gradual shift by snowfall over time (Drews and others, 2020); 5) Figure 5d shows a sharp V layer within 10 m of the surface, but offset to the west by over a kilometer from the basal channel.

To further quantify the process responsible for the offset between the englacial layers and the surface depression, we estimated the location of the surface depression through time in Sentinel-1 imagery. A surface feature that is actively migrating with wind-driven snow erosion and deposition will show change distinct from advection by ice flow. Figure 7 shows the surface depression from each year, advected to November 2023. Downstream of our channel reach, the channel has slightly moved toward the west (due to human error in picking, this movement may not be significant). Near and upstream of our channel reach, we see more distinct variation in surface depression geometry between years, with possibly a slight clockwise rotation of the channel surface expression.

DISCUSSION

This reach of the TEIS channel has complex topography and numerous englacial fractures and folds, more so than nearby areas outside of the channel. For comparison, to the west of our channel reach, at the same thickness and the same distance from the grounding zone, there is no evidence of recent fractures or folding (Figure S31 and S32 and similar features presented by Wanzer and others (in prep)). Figure S41 shows thicker ice, more uniform englacial layering, and smoother basal topography. We argue that the complexity in this relic channel reach is an outcome of a locally compressive ice environment combined with a lack of strong along-channel currents. Without currents in the channel, melt processes are more variable in time and space compared to an active channel. Our relic channel observations are in contrast to the “steady state” channel studied by Wearing and others (2021) in which melt drives persistent viscous secondary flow.

Beyond being a relic channel, this channel reach is located in a thin (<300 m) ice shelf that has a core of cold ($\sim -20^{\circ}\text{C}$) ice inherited from upstream, sandwiched between ~ 80 m of warm firn and ice damaged by past surface crevassing, and ~ 30 m of ocean-warmed ice at the base (Sharp and others, in review). Figure 4a, for example, shows disturbed layers under—but not above—the grounding-zone “bright layer”. It is likely that the ice experienced surface crevassing just before the formation of the bright layer. The response of layered ice like this will depend critically on the style (normal or shear) and magnitude of applied stress, which in this case is changing through time, as we discuss more below. Here we discuss how local processes drive complexity in basal topography and englacial structure in this relic channel, how this complexity relates to the surface depression that is visible from satellites, and suggest implications for ice-shelf stability.

Channel closure resulting from “relic” channel status.

Because an unconfined ice shelf will tend to spread viscously (e.g. Weertman, 1957), topographic relief will tend to smooth out. Lower topographic relief (which generates very small deviatoric stress gradients) will smooth out more slowly than larger topographic relief due to the nonlinearity of the ice flow law (e.g. Weertman, 1983). We thus expect a relic channel to initially viscously close more rapidly when active basal melt shuts off, then more slowly.

The TEIS channel is viscously closing. First shown by Wild and others (2024, in review), who show thickening ice over the channel using phase-sensitive radar. We expand on this with additional observations. First, the downstream channel cross sections are wider with a lower-elevation apex compared to upstream

(Figure 3g,h,i and S39a). Given vertical strain rates measured by Wild and others (2024), the ice column above the channel could thicken by as much as 40 m in the 5 to 6 years it would take for ice to travel from the upstream end of our reach to the downstream end, in agreement with Figures 3g,h,i and S39a. This lowering and widening of the channel apex is unlikely to result in an active basal channel current because buoyant meltwater would need to flow against the hydraulic gradient through this channel reach. In a river system, flowing against the local hydraulic gradient can occur when a lake forms to fill the low topography. While it may be possible for a meltwater “lake” to form on the underside of an ice shelf, it is unlikely to form in an active basal-channel because along-channel currents would tend to melt the ice, and incise the channel so as to maintain active flow.

The strain rates in Figure 2e show significantly more negative principal strain rate along the channel than elsewhere. Given that this strain rate is concentrated in the transverse direction (Figure 2c) rather than longitudinal (Figure 2b), continuity indicates that the transverse compression must be accommodated by vertical extension (thickening) within the channel. Alley and others (in review b) shows that some reaches of the channel have likely experienced shear in addition to compression. This thickening over the channel implied by the compressive strain rate also appears in our repeat radar profiles. Figure S38c shows thickening ice within the channel over the three-year period as compared to elsewhere. Due to steeply dipping crevasse walls that obscure the local bed topography in radar, artifacts appear in these comparisons that need careful analysis. Thickening or thinning is a combination of deformation and basal melt or freezing; local melt rates from Wild and others (2024) and Wanzer and others (in prep) are typically less than 2 m a^{-1} and there is no evidence of freeze-on. Therefore, we infer that thickening within the channel is most likely due to the channel viscously closing faster than it is being melted.

Interestingly, if the channel started viscously closing 15 to 18 years before our data collection, according to the timeline proposed by Wild and others (in review), and this were the only process acting on the system, then we would expect only a small remnant basal channel and a small associated surface depression filling in with snow during that time. This is not what we see; therefore, we look to additional local processes to sustain the complex basal topography with an incised channel.

Elevated melt on steep channel walls: M shape

We expect weak along-channel currents in a relic channel. These conditions allow local melt processes to dominate, and we argue that the distinctive and persistent M-shaped channel profile we see in radar cross

sections (Figure 3c,i) is a signature of these local melt processes.

The M shape forms from elevated melt on the channel sidewalls, with peak melt at the top of each steep slope instead of in the center of the channel. Elevated melt on steep ice faces is due to convection-dominated melting driven by buoyant plumes of meltwater that rise along the ice-ocean interface (Figure 3c). According to Villhauer and others (in prep), who modeled the M-shape channel based on our observations, melt rates scale with convective plume-driven velocities under weak along-channel currents. They show that the magnitude of along-channel currents as well as the geometry of the channel modulate the buoyancy-driven convective regime in a few ways. First, taller channels provide an opportunity for greater vertical development of plumes, while steeper wall slopes increase the buoyancy component, both of which lead to higher melt rates. Second, moderate along-channel velocities enhance plume entrainment and, therefore, melt rates, over either extremes of high velocities (“active” basal channel behavior) or near-zero velocity. Stronger along-channel currents tend to confine plumes and reduce mixing efficiency, thus reducing sidewall melt.

The notion that steeper ice interfaces have elevated melt rates compared to horizontal interfaces has been studied in laboratory (e.g. Mondal and others, 2019) and observed in geophysical settings (Hester and others, 2021; Wanzer and others, in prep), yet it is still rare to see the effect reflected in basal topography. As Wanzer and others (in prep) shows, the stepped-terrace features we see (Figure 4h) are sustained by elevated melt on steep ice faces (in the Wanzer and others study, they show the topography is persistent for at least 13 years in an area just west of our channel). Once an ice interface is steep enough, locally derived melt plumes can sustain the steep interface.

The shape of a basal channel is also affected by the Coriolis Effect because it acts to deflect plumes to the left (westward) in the downstream direction (e.g. Gladish and others, 2012). We attribute the asymmetry we observe in channel profiles, with the western side being taller than the east side (also shown in modeling, Villhauer and others, in prep). Washam and others (2023) describe an asymmetric “trapezoidal” shape to basal crevasses on the Ross Ice Shelf, which may derive from a similar melt process. The asymmetry of the channel—due to asymmetry in melt rate—implies that external flow (under the ice shelf) is in a direction that allows the Coriolis Effect to add momentum to the plumes on the western side of the channel (Villhauer and others, in prep).

Our most upstream cross sections (Figure 3g) show narrow U-shaped geometry instead of the M shape. The U shape indicates that melt from stronger along-channel currents dominates over that from sidewall

convective plumes. Importantly, we see the U shape in profiles where the channel is narrow and more incised into the base of the shelf, implying recent or current local basal melt rates are higher. The M shape appears in profiles where the channel is less incised and wider, suggesting lower basal melt rates. Such a narrow, deeply incised U-shape within a few kilometers of an M-shape channel suggests localized high-velocity currents—“floods”—that may stem from a crevasse opening at the apex, breaching the channel from the side, or redirecting water from a more distant fracturing event.

The spatially and temporally variable melt patterns suggested by the M-shaped topography within this one reach of the channel support the notion that melt plumes are active. Further, the U-shaped profiles in our data set point to locally sourced along-channel currents. The variable nature of the flow is in contrast to either consistent subglacial discharge-driven along-channel currents or more quiescent “lake”-type ponding of meltwater filling features in the topography.

Recent active fracturing influences basal topography

The presence of basal fractures in the radar is evidenced by hyperbola stacks, scattered internal reflections, missing basal reflectors, or V layers (in the middle to upper ice column). That many of the crevasses we see are associated with dateable V layers points to these fractures being recent, rather than grounding-zone-generated basal crevassing (as seen elsewhere in the ice shelf as shown by Wanzer and others, in prep).

A complete understanding of the underlying processes that create the V layers is beyond the scope of this study, but we hypothesize that the layers are due to either elastic (flexural) response to crevasse formation (Buck and Lai, 2021) or viscous response (e.g. Wearing and others, 2021) to very localized, extremely high melt within a newly opened crevasse or, most likely, a combination of these two. Regardless of the exact formation mechanism, we are confident these V layers are associated with crevasse formation and thus can be used to constrain the age of crevasses in our radar.

We associate this evidence of active basal crevassing with the local transverse compressive stress regime shown in Figure 2, which Alley and others (in review b) shows has evidence of some shear as well. Furthermore, the firn layers show evidence of syncline/anticline folding as well (Figure 3 and 5). Given the colder, stiffer ice in the middle of the ice shelf and the warmer, softer ice in the upper 80 m, a combination of folding and fracturing may be expected, especially when coupled with evidence of shear in the past (Alley and others, in review b).

This combination of folding and fracturing is supported by an unusual vertical strain profile in the center of our study area measured by Wild and others (in review) using phase-sensitive radar. This profile shows the top and bottom ~ 100 meters of ice as thickening, while a central layer of cold ice at roughly ~ 100 to 150 m depths is thinning. With a single profile of vertical strain rate, it is difficult to identify the mechanisms; this type of vertical profile is more likely to occur within a combined stress regime in a three-dimensional setting in a layered medium, as we see here. While modeling the details of three-dimensional folding and fracturing is beyond the scope of this paper, we suggest that a combination of ductile and brittle failure mechanisms creates the complex englacial radar layers seen in the firn and the apparent thinning amidst an overall ice thickening signal.

While the vertical strain profile described by Wild and others (in review) persisted for the 9-month deployment in 2022 at the same time the satellite-derived transverse compression was present, the V layer dating suggests the most active period of crevassing was before this, between 2010 and 2015. We infer from this that a stress regime change occurred around 2015 or soon after that reduced the local stresses below the fracture toughness of the ice. This timing is the same as the initial growth of the “Dagger” fractures described by Wild and others (2022) and further analyzed by Sharp and others (in review). Specifically, Sharp and others (in review) show that a transition in stress occurred when the connection between TEIS and the Thwaites Glacier Tongue weakened in 2016. The ice shelf flow rotated to flow more directly into the pinning point, which reduced the shear component of stress along the channel, reducing the likelihood of fracturing. Further, the formation of the Dagger fractures possibly also accommodated deformation that reduced stresses within the channel.

Therefore, the active period of fracturing in the channel was between ~ 2010 and 2015 (Figure 6): after the channel disconnected from its original source (becoming relic, ~ 2006) and just before the opening of the Daggers in 2016. Each of these crevassing events may be sources or sinks of meltwater and influence the pattern of melt and topography in and around the channel through two primary mechanisms. First, crevasses present steep ice interfaces that will drive convective plumes that generate meltwater. Second, they act as new pathways (sources or sinks) for meltwater in and out of the channel from elsewhere. Both of these processes may create local elevated melt and help sustain complex topography through time even though the channel does not have a persistent source of along-channel currents.

Stepped-terrace features: two hypotheses

The terrace features seen on the sides of the channel (Figure 4g,h) are similar to those described by Dutrieux and others (2014) and Rignot and others (2025). They are couplets of steep ice slopes separated by horizontal surfaces. These steep ice slopes will tend to generate convective plumes (Villhauer and others, in prep), which may allow them to be self-sustaining. They need, however, a steep ice slope as a “seed” to form. From our study, we envision two possible formation mechanisms. First, the crevasses that formed will expose steep surfaces to the ocean, which can then initiate melt. Second, “flood” events from the opening of a crevasse may generate local along-ice currents that rapidly incise into the basal ice through shear-driven melt (Rosevear and others, 2025), leaving behind steep walls that are then susceptible to convective-plume melt. This second hypothesis is similar to river terrace formation.

Channel limits basal crevasse propagation

Both within our field site and further downstream along the channel, we see multiple relatively large basal crevasses that intersect the channel, but do not cross the channel. We believe this is a direct result of the stress regime within the channel. As discussed previously, the ice above the channel is experiencing vertical extension and transverse compression, despite overall ice shelf spreading. We argue that the spatial transition from a weak to negligible extensional regime outside the channel to a stronger compressive regime over the channel decreases the effective viscosity (increasing the fracture toughness) and alters the crack tip opening stresses. Changes in both fracture toughness and crack tip opening stress can act to arrest propagating crevasses (similar to that described by Sharp and others, in review). In this sense, the channel may in part act as a stabilizing feature for the ice shelf, preventing these large basal crevasses from extending further across the shelf.

Transverse compression contributes to offset of surface depression from basal channel

The surface depression is offset from the basal channel by approximately one ice thickness (Figure 3 and 5); however, only the shallowest englacial layers (within ~ 5 m) follow this surface topography; all deeper englacial layers more closely follow the basal topography instead. Furthermore, the surface depression as mapped by satellite imagery (Figure 7) does not show significant lateral migration relative to flow direction, ruling out gradual migration driven by asymmetry in accumulation or basal melting Drews and others (2020). The rapid transition between englacial layers following the basal topography and those following

the surface topography points to a “sudden” (within a few years) process that disconnected the surface depression and basal channel. Theory suggests that active basal channels draw down the surface directly above them (Wearing and others, 2021; Humbert and others, 2022). Elastic response of the ice shelf to a basal channel would form a shallow, broad depression (Buck and Lai, 2021; Turcotte and Schubert, 2002; Rosier and others, 2017), while viscous response creates a depression the same width as the basal topography with a depth that depends on the ice column reaching hydrostatic equilibrium (Humbert and others, 2015, e.g.).

Drews and others (2020) show that either wind-driven asymmetry in snow deposition or asymmetry in basal melt can cause migration of the surface expression and basal channel relative to ice flow direction. They further show that tilted syncline/anticline patterns can capture this ongoing change. In our case, the one ice-thickness offset of the surface depression from the basal channel and the lack of tilt in the englacial layer pattern suggest a more sudden (within just a few years) shift of the surface depression. We surmise that the overall compression acting transverse to the basal channel caused fracturing and folding that created a change in the surface topography, such that the “shoulders” (anticline) on the upwind side of the channel temporarily increased the rate of wind-driven snow redistribution in recent years. On an otherwise flat ice shelf, even a few meters of change in local relief can significantly alter the snow redistribution in a short period of time.

CONCLUSIONS AND IMPLICATIONS

In this study, we present a detailed look at a relic basal channel on the Thwaites Eastern Ice Shelf that reveals high temporal and spatial variability in channel topography and englacial layering. We find that more recent fracturing, folding, and localized melt have overprinted the original channel geometry and triggered a recent wind-driven infilling and migration of the surface depression.

We find three dominant processes acting to evolve the shape of this channel. First, weak along-channel currents enhance melt from convective melt plumes on the channel sidewalls, which leads to an M-shaped channel. Similar melt plumes likely help sustain other steep interfaces such as the stepped-terrace features that span the flanks of the channel. Second, fracturing and folding of the ice due to the concentration of compressive stress along the channel change the geometry and stimulate melt and changes in snow distribution. And third, potential flooding events from the opening of crevasses further carve the channel in local areas, creating narrow U-shaped channels (upside-down canyons) and steep slopes susceptible to

convection-driven melt. All of these processes and their changes through time lead to sustained channel-like topography, despite losing its primary source of subglacial meltwater over 15 years before our first field season.

The implications of our findings for TEIS and for other ice shelves include:

- Relic channels can persist for decades in thin ice shelves through localized melt, fracturing, and folding, despite ongoing viscous closure.
- Local convective melt plumes along steep ice interfaces are important to the evolution of basal topography; steep ice slopes may persist for extended periods of time.
- Despite being a source for fractures due to localized transverse compressive (or shear) stresses, these newer fractures do not seem to be propagating outward across the ice shelf.
- The channel stress regime appears to alter the fracture toughness of the ice and/or reduce the crack tip opening stress to a degree that limits propagation of fractures that intersect the channel.
- The surface depression associated with the channel as viewed from satellite may hide more complex processes, especially in the case of a relic channel.

In sum, over the time period we studied and the decade prior to that, we show that this relic channel has been acting as a neutral or stabilizing feature for TEIS, in contrast to the perception that basal channels are typically destabilizing features. Given the geometry and damage we observe in the ice, however, if the stress regime switches again to stronger along-channel shear (as discussed in Alley and others, in review b), then the fractured and folded ice may be more susceptible to propagation of along-channel fractures. This relic channel would then become a destabilizing feature of the ice shelf.

SUPPLEMENTARY MATERIAL

The supplementary material for this article can be found at [link pending].

ACKNOWLEDGMENTS

This work could not have been done without the TARSAN Team. We would like to thank Cece Mortenson, Blair Fyffe, and Chris Simmons for keeping the TARSAN team safe for all field seasons; WAIS Divide staff, Twin Otter Pilots, McMurdo staff, the US Antarctic Program, and the US Air National Guard. We thank

Suzanne Bevan for providing the TanDEM-X DEM used in this study. We also thank current and former OSU Glaciology Team members, especially Celia Trunz and Kirill Ivanov, for contributing to the radar processing workflow.

This work is from the TARSAN project, a component of the International Thwaites Glacier Collaboration (ITGC). Support from National Science Foundation (NSF: Grant 1929991) and Natural Environment Research Council (NERC: Grant NE/S006419/1). Logistics provided by NSF-U.S. Antarctic Program and NERC-British Antarctic Survey. ITGC Contribution No. ITGC-169. CTW was partially supported by the Deutsche Forschungsgemeinschaft (DFG DR 822/8-1).

DATA AVAILABILITY

All radar profiles, both raw and processed, and kinematic GPS files are published on Zenodo at the following links: 2019/20 100 MHz (<https://zenodo.org/records/22981674>) and 2022/23 100 and 20 MHz (<https://zenodo.org/records/20495485>). The TanDEM-X DEM can also be found at the first link. All radar data and accompanying files will be publicly available starting June 1, 2027. Sentinel imagery is available from the Copernicus598 Open Access Hub (<https://scihub.copernicus.eu/>). Strain rates and ice velocities from Banerjee and others (2025) are publicly available on the US Antarctic Program Data Center: <https://www.usap-dc.org/view/dataset/601904>. ICESat-2 data is publicly available and was accessed using NASA's OpenAltimetry program: <https://openaltimetry.earthdatacloud.nasa.gov/data/icesat2/>. NASA ITS_LIVE data can be freely accessed at the following link: <https://nsidc.org/apps/itslive/>. matGPR is freely available at: <http://users.uoa.gr/~atzanis/matgpr/matgpr.html>.

All associated TARSAN project data, including AMIGOS data, is available on the US Antarctic Program Data Center: <https://www.usap-dc.org/view/project/p0010162>

AI DECLARATION

Artificial Intelligence (AI), specifically Microsoft Copilot, was utilized during this research to debug code and assist in figure formatting. Grammarly was used in writing for basic grammar and spell-checking. AI was not used to process or analyze data, generate ideas or hypotheses, or generate original text and images for this paper.

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For Peer Review

Fractures and Floods: Small-scale spatial and temporal processes driving change in a relic ice-shelf basal channel

SUPPLEMENTARY MATERIAL

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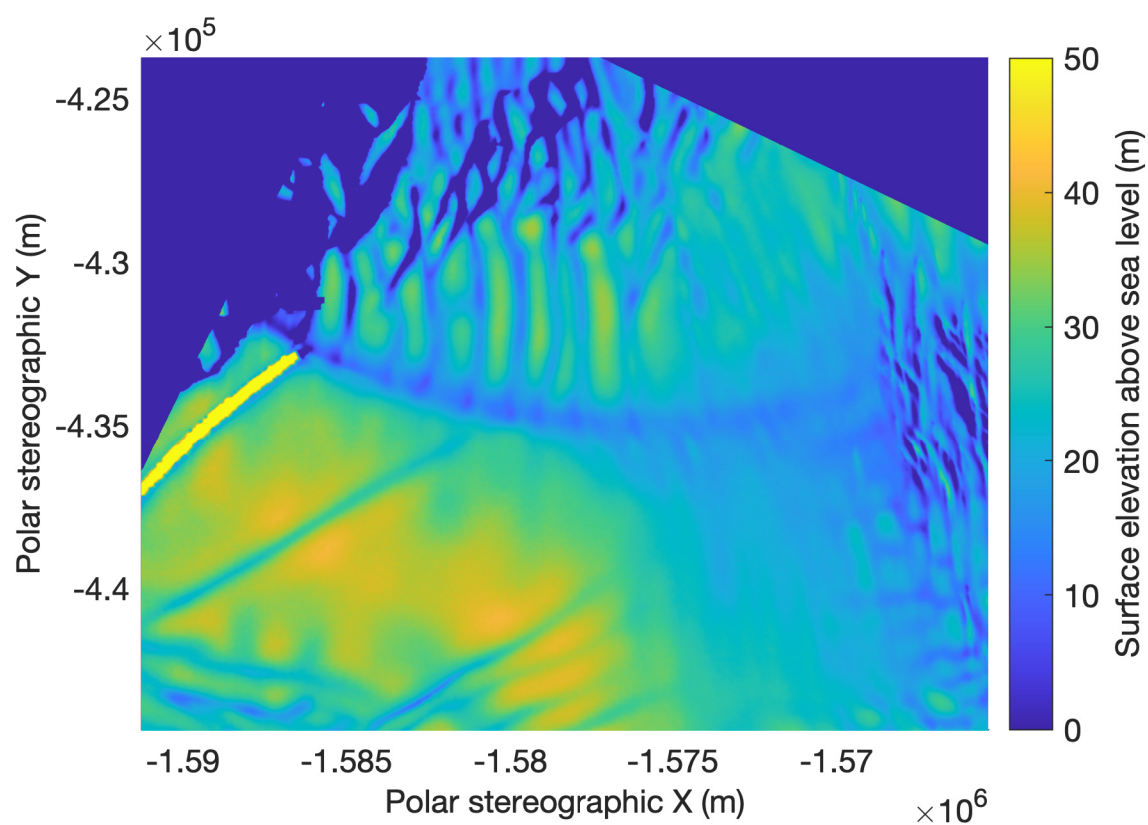


Fig. S1. The TanDEM-X surface topography from February 8, 2020, courtesy of Suzanne Bevan, vertically adjusted to nearby ICESat-2 elevations. This DEM was used to fill in gaps in 2019/20 surface elevation data, as well as to quality-check elevation data gathered from external GPS units in the field.

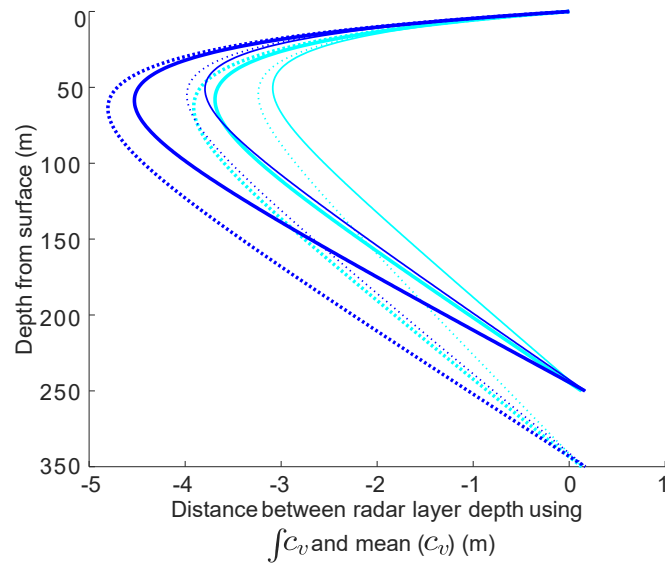


Fig. S2. Estimate of the offset between the true depth of a reflector and the apparent depth as presented here due to wave speed in the firn column. Blue/Cyan is for surface density of 350/450 kg/m³. Thin/thick lines are 30/40 m depth of pore close-off. Solid/dotted lines are for 250/300 m thick ice.

Fig. S3. The locations and names of every profile making up the channel grid from both field seasons. Panel **a** shows 2019/20 100 MHz profiles, **b** displays 2022/23 100 MHz profiles, and **c** shows 2022/23 20 MHz profiles. Certain across-channel transects, X2 from 2019/2020 and P1-P3 from 2022/23, are not single profiles but rather made up of multiple consecutive profiles.

19/20 : 100 MHz	22/23 : 100 MHz	22/23 : 20 MHz
X1	P1	A3
X3	P2	
X4		
X5	P3	

Table S1. Profiles on the same row represent Lagrangian pairs between field seasons, meaning that the 2022/23 profile is meant to capture the same section of the ice shelf as the 2019/20 profile when accounting for ice shelf flow.

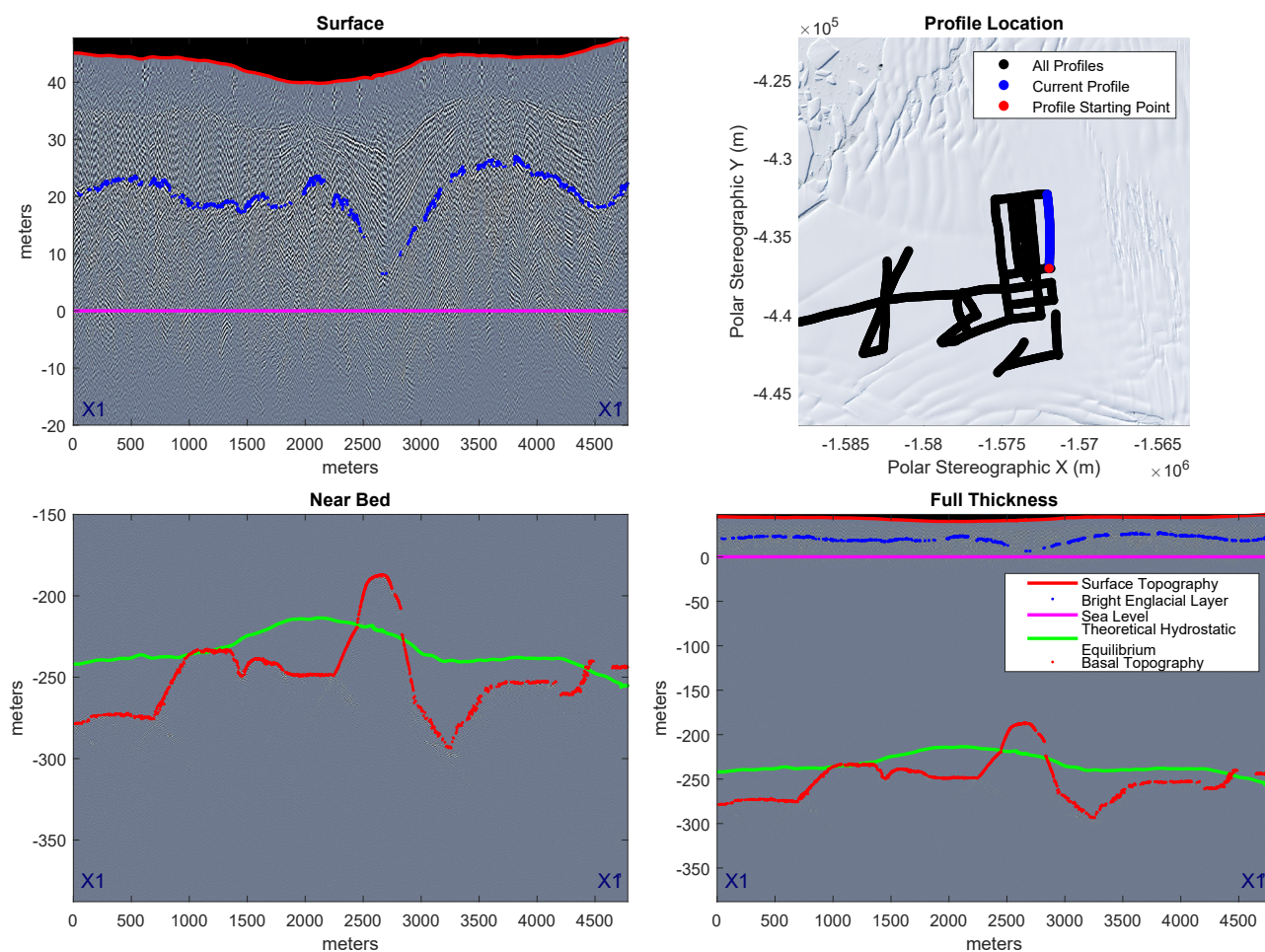


Fig. S4. Profile X1. A 100 MHz profile from the 2019/20 field season.

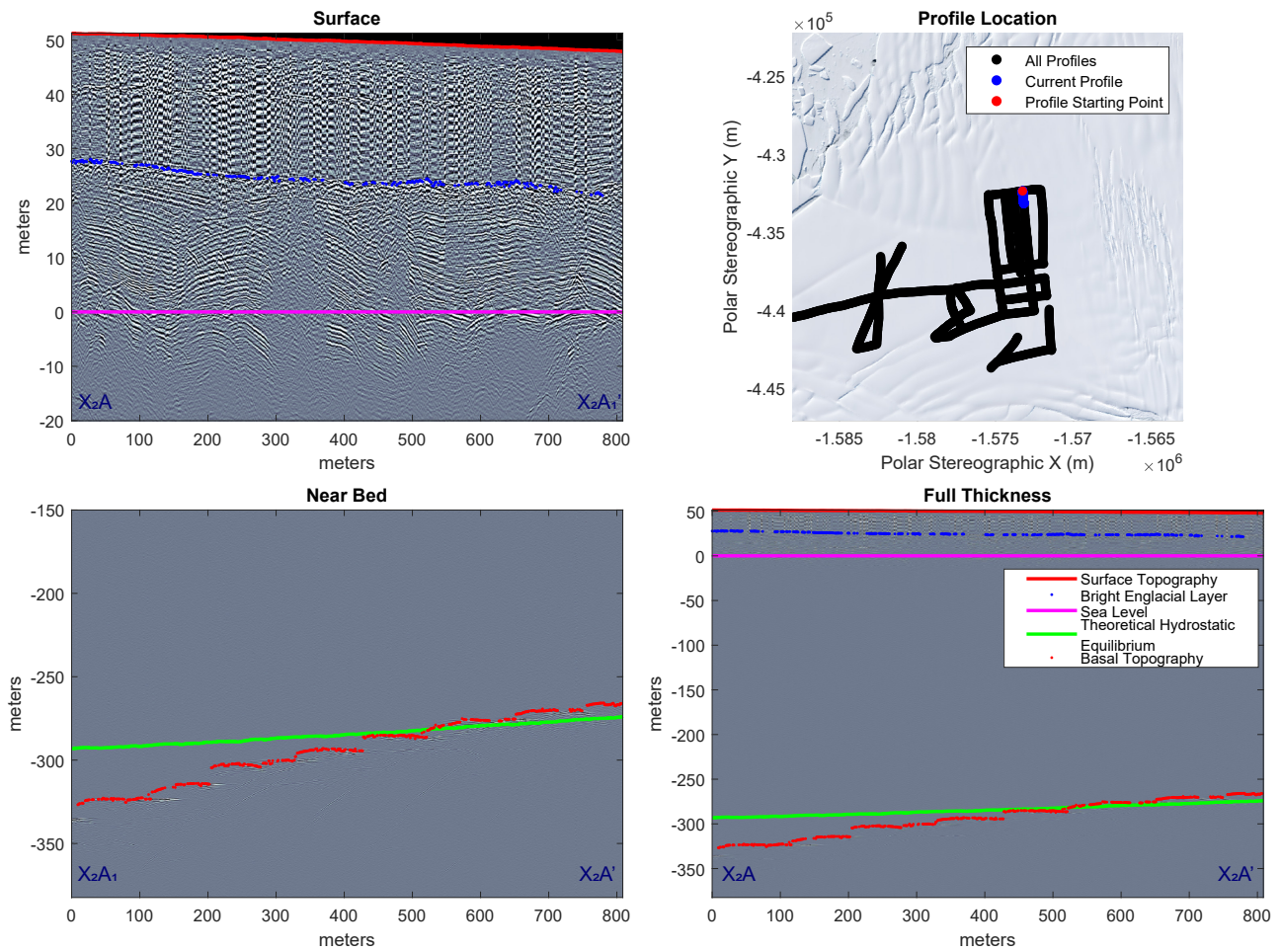


Fig. S5. Profile X2A. A 100 MHz profile from the 2019/20 field season.

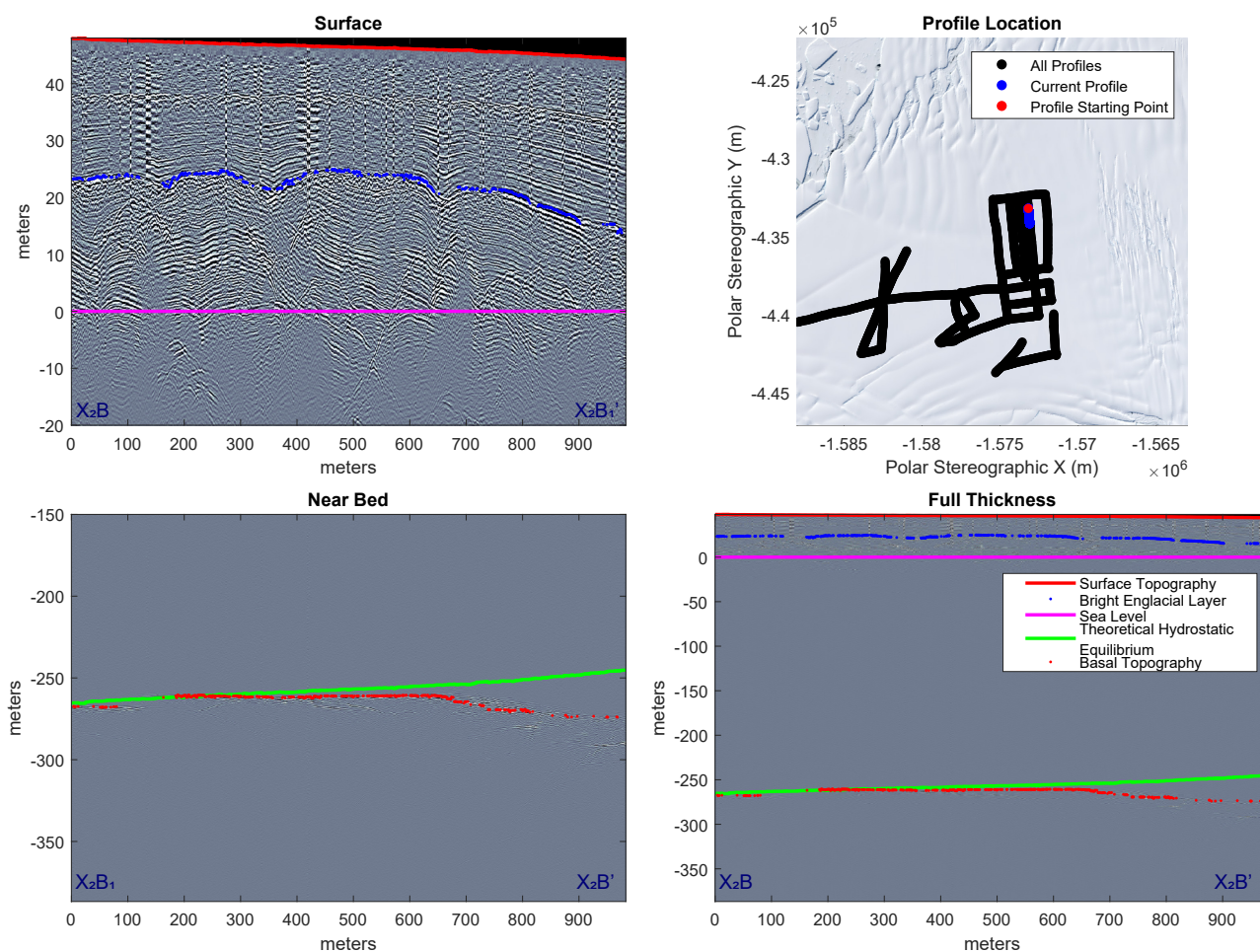


Fig. S6. Profile X2B. A 100 MHz profile from the 2019/20 field season.

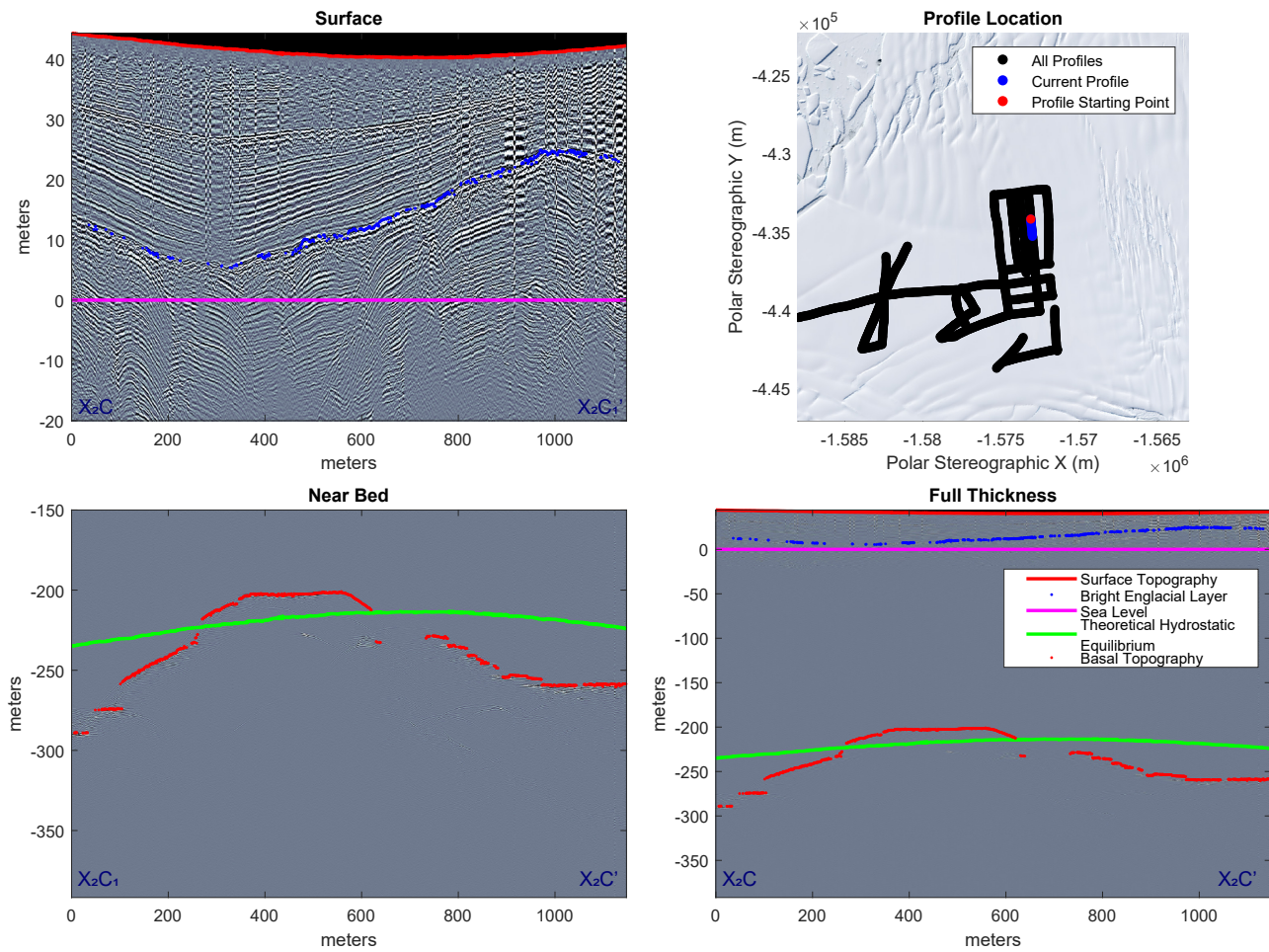


Fig. S7. Profile X2C. A 100 MHz profile from the 2019/20 field season.

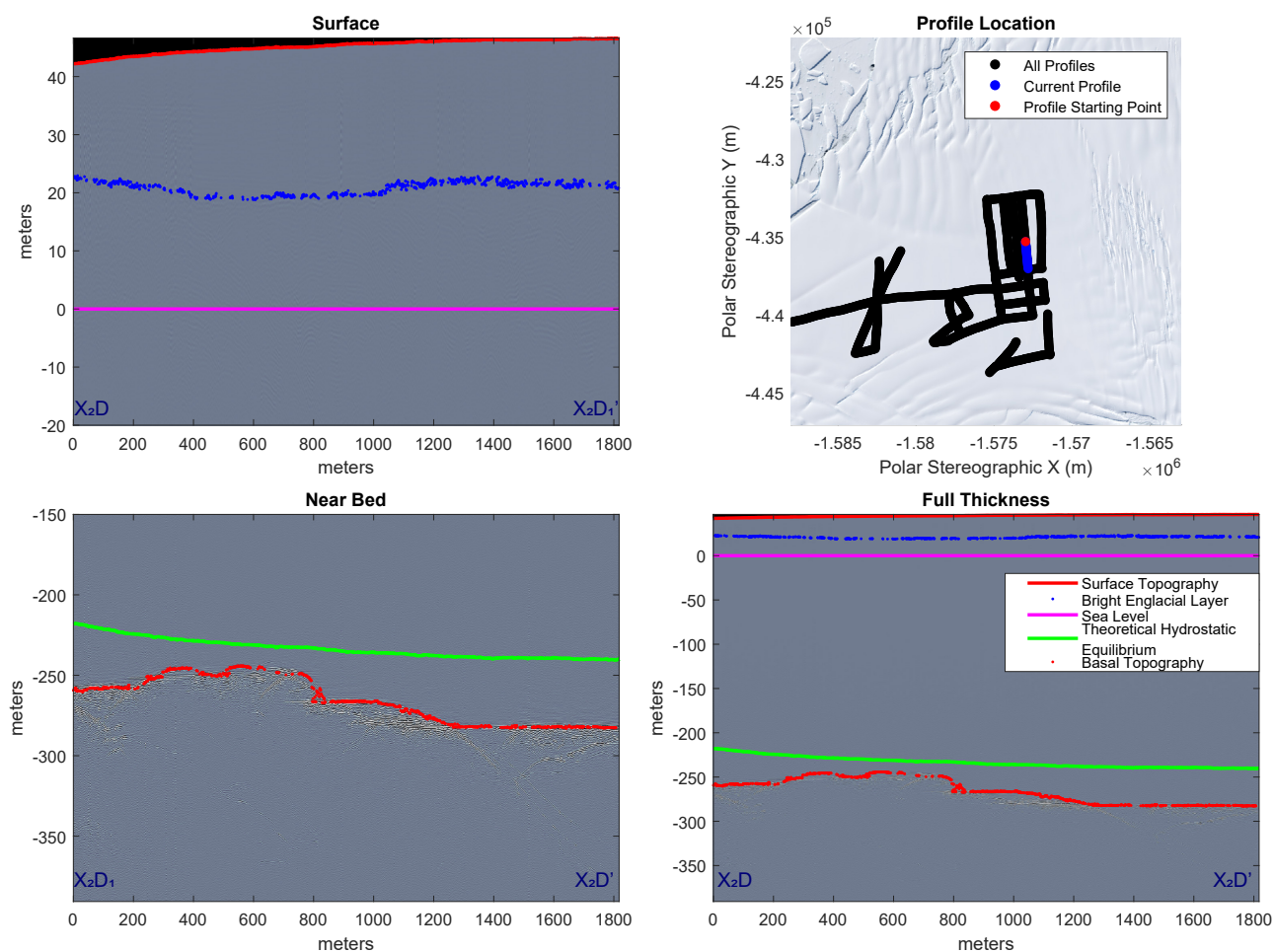


Fig. S8. Profile X2D. A 100 MHz profile from the 2019/20 field season.

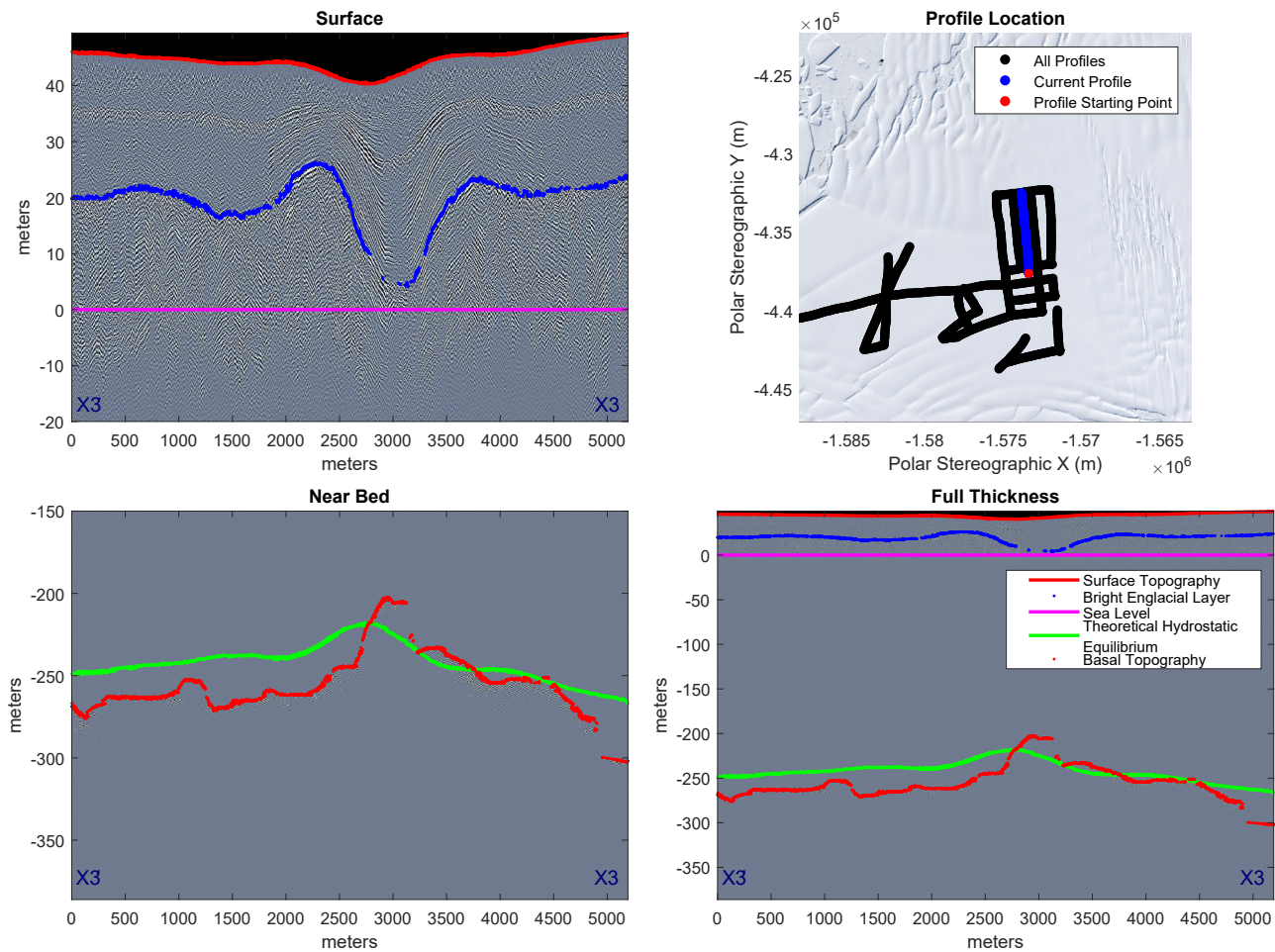


Fig. S9. Profile X3. A 100 MHz profile from the 2019/20 field season.

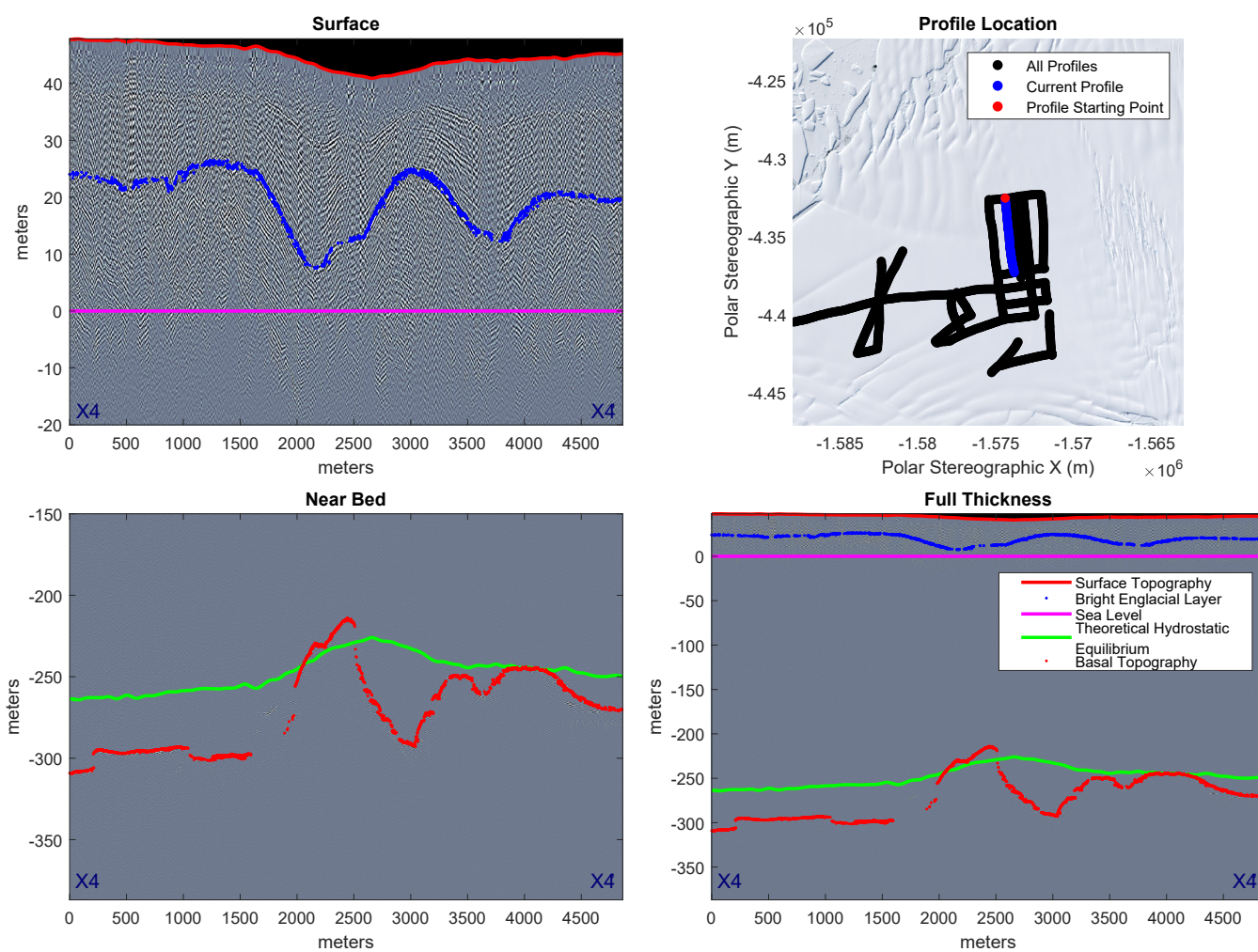


Fig. S10. Profile X4. A 100 MHz profile from the 2019/20 field season.

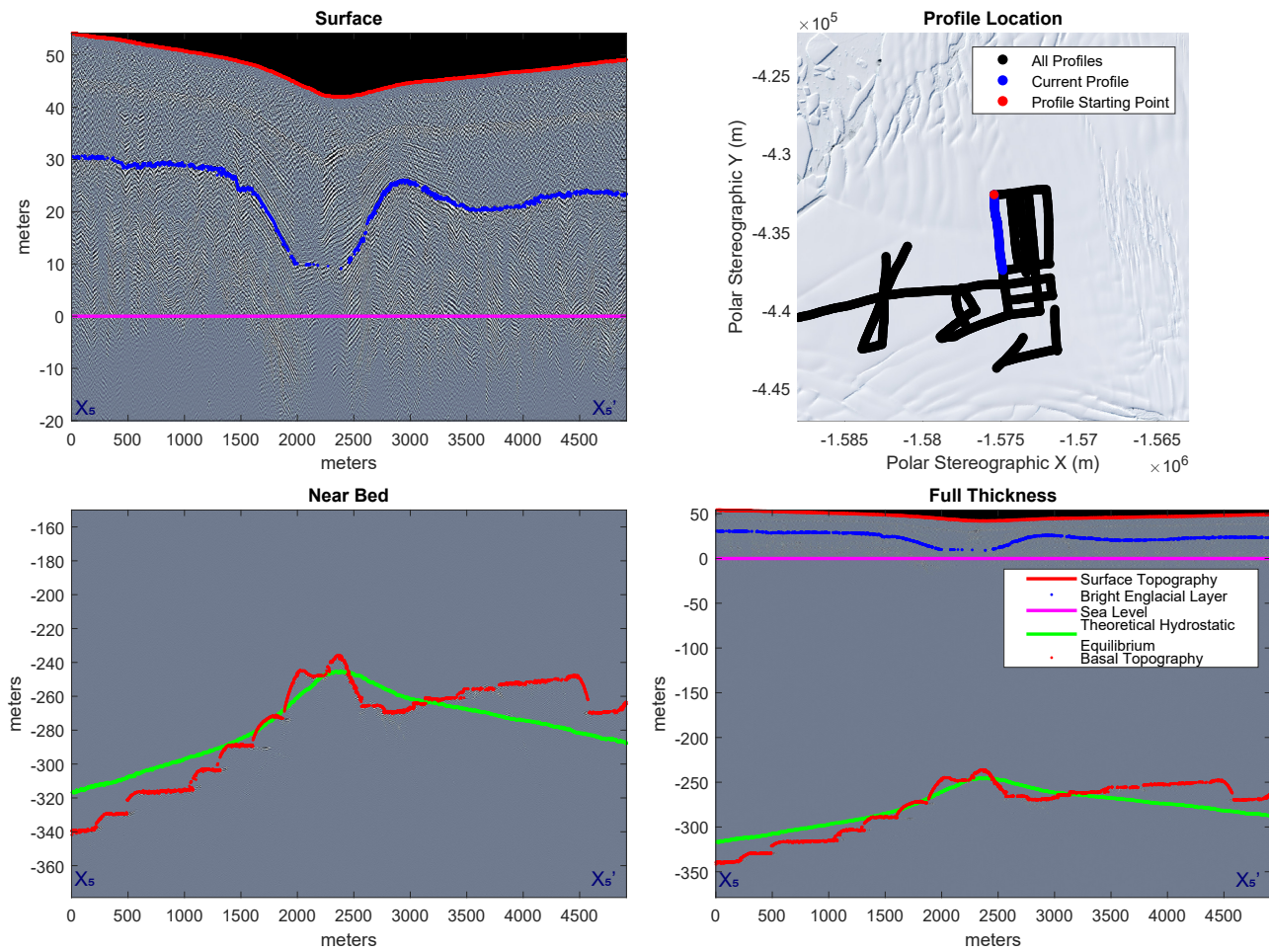


Fig. S11. Profile X5. A 100 MHz profile from the 2019/20 field season.

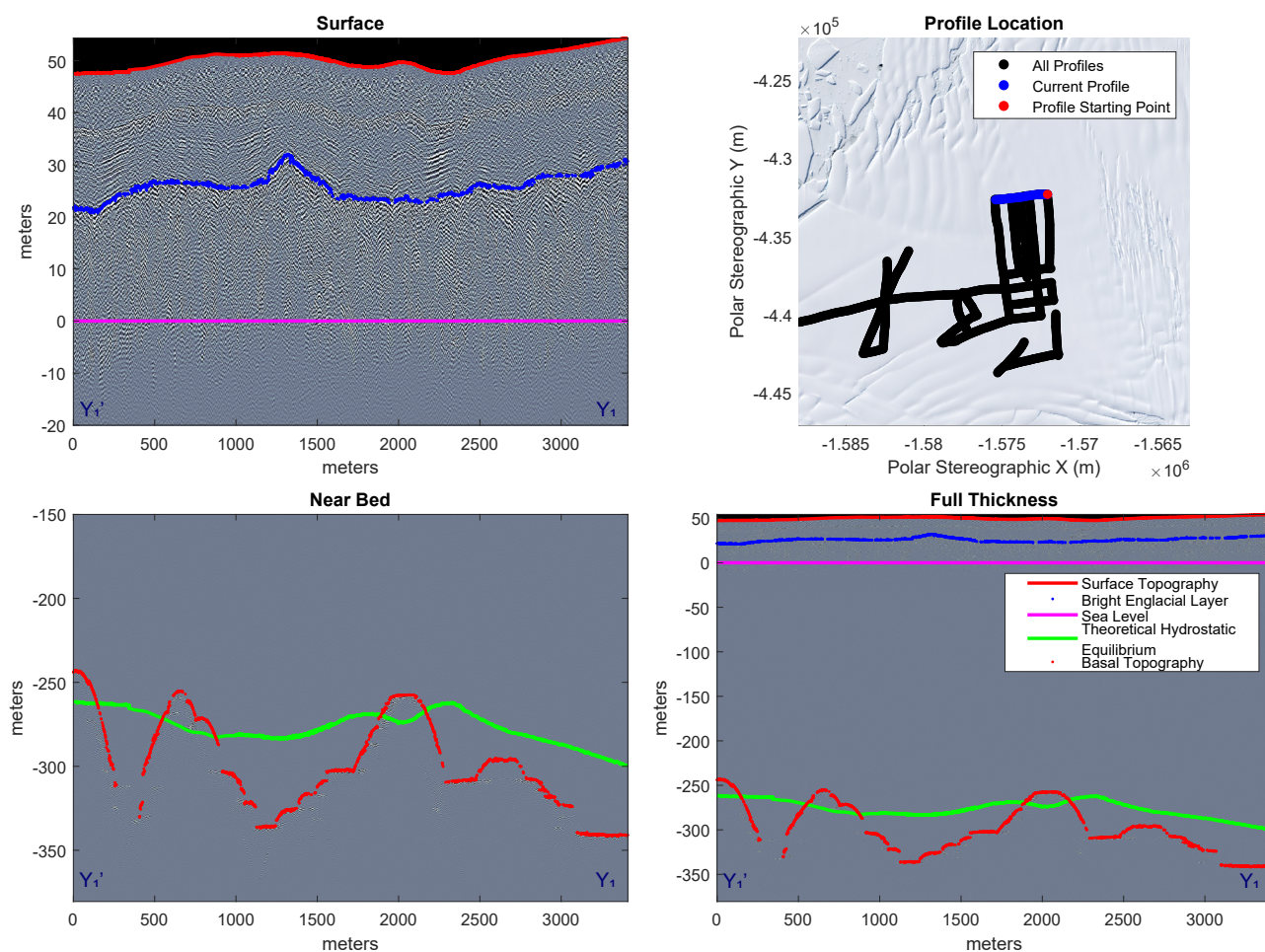


Fig. S12. Profile Y1. A 100 MHz profile from the 2019/20 field season.

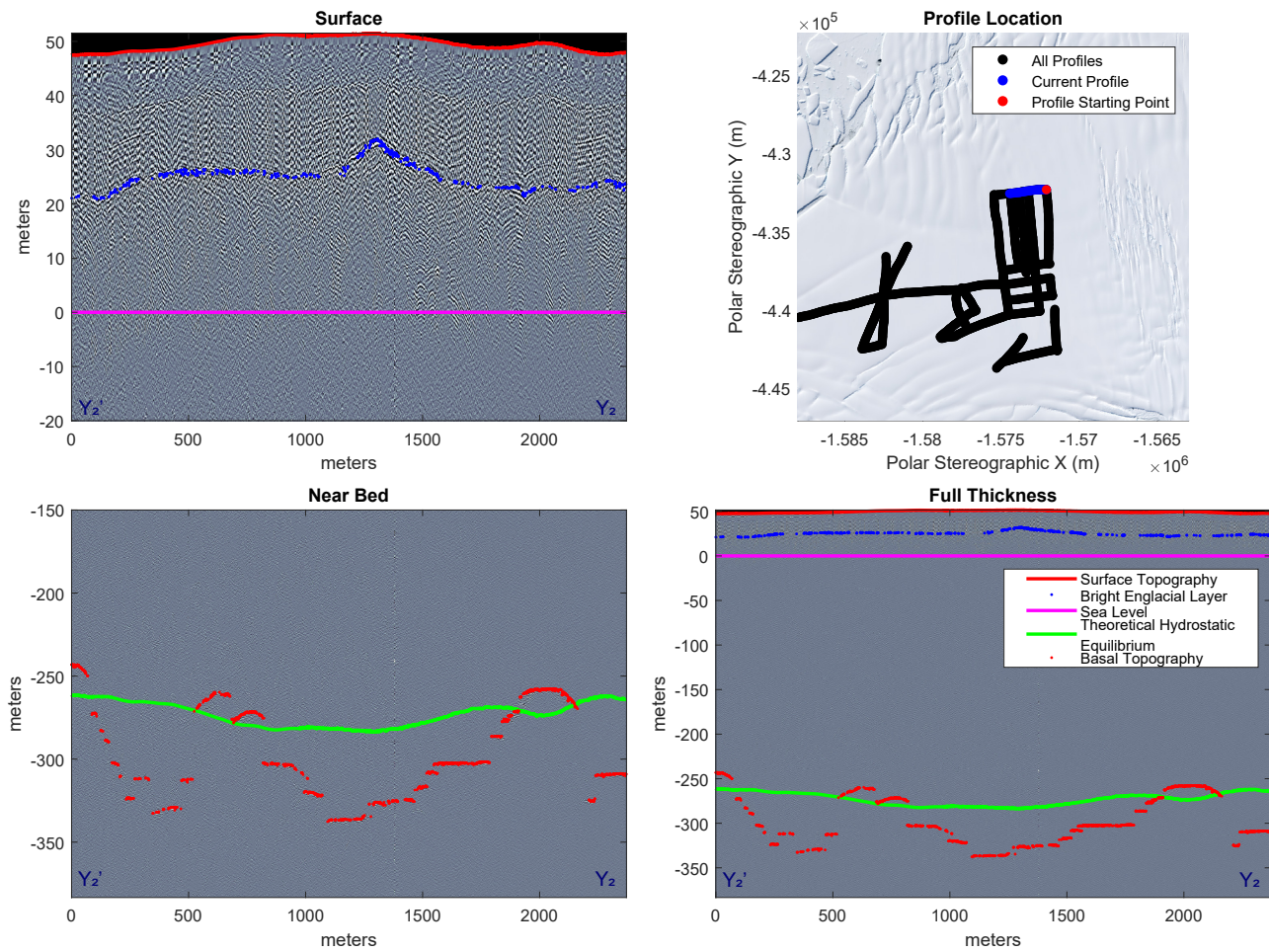


Fig. S13. Profile Y2. A 100 MHz profile from the 2019/20 field season.

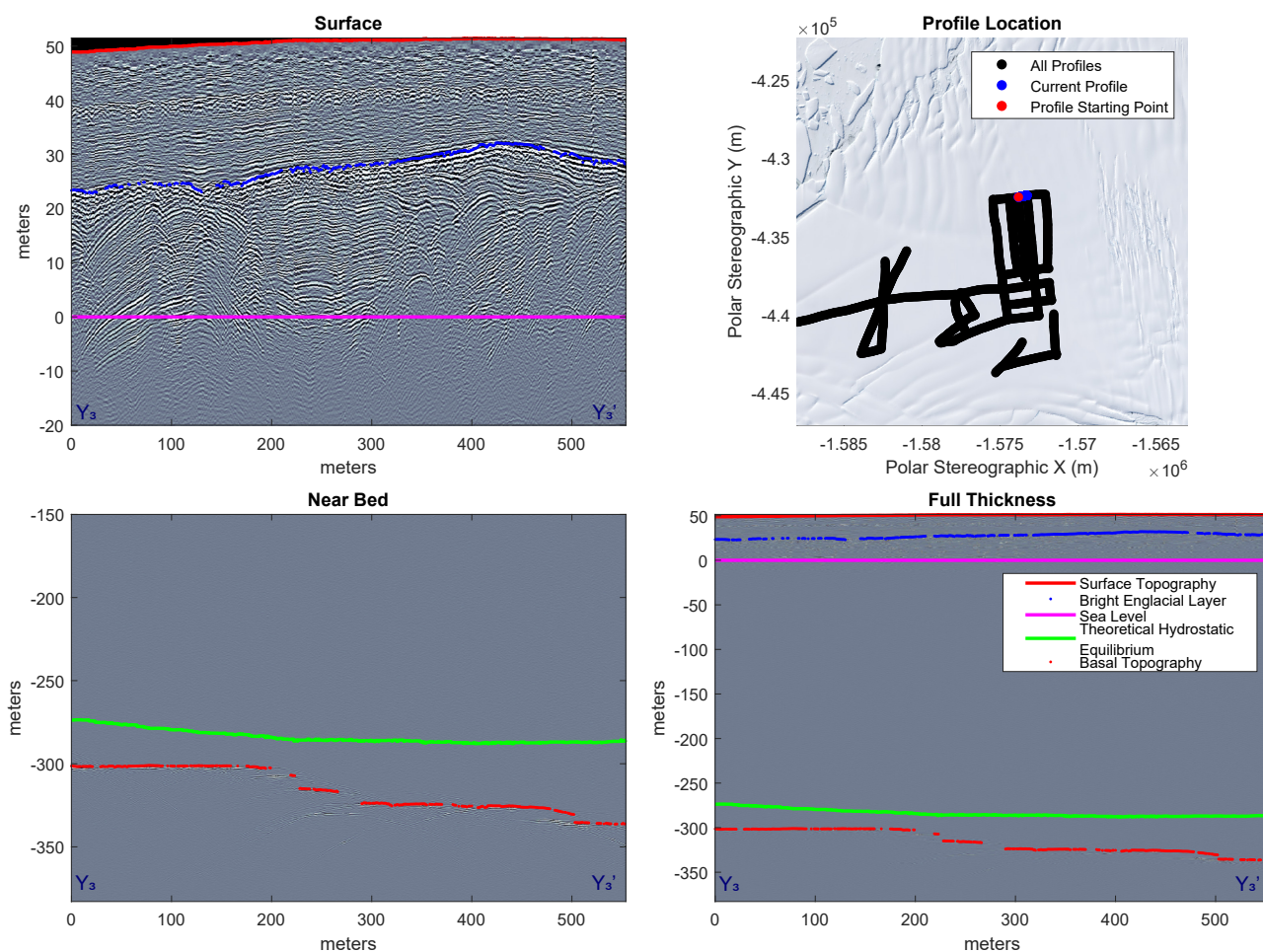


Fig. S14. Profile Y3. A 100 MHz profile from the 2019/20 field season.

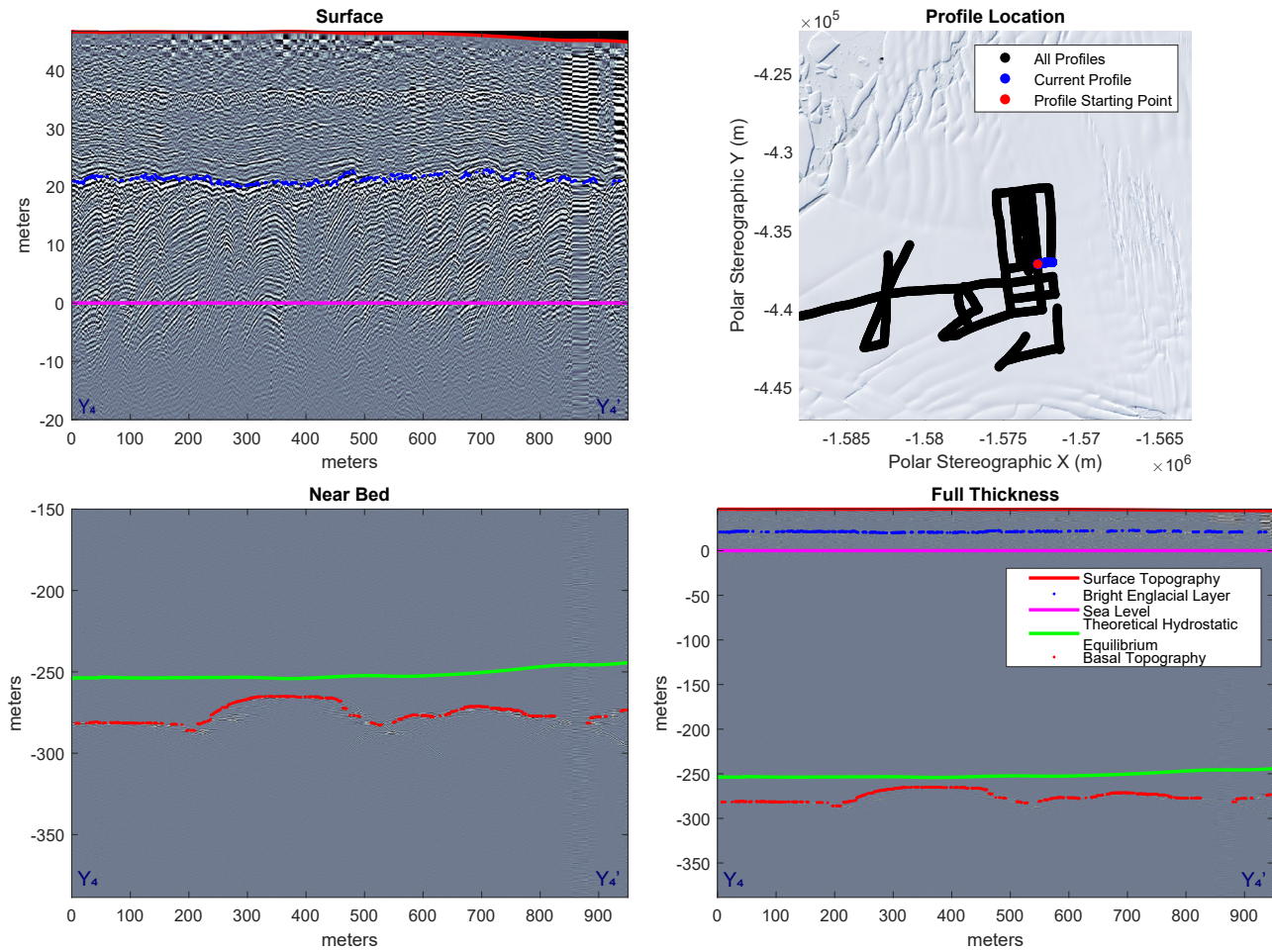


Fig. S15. Profile Y4. A 100 MHz profile from the 2019/20 field season.

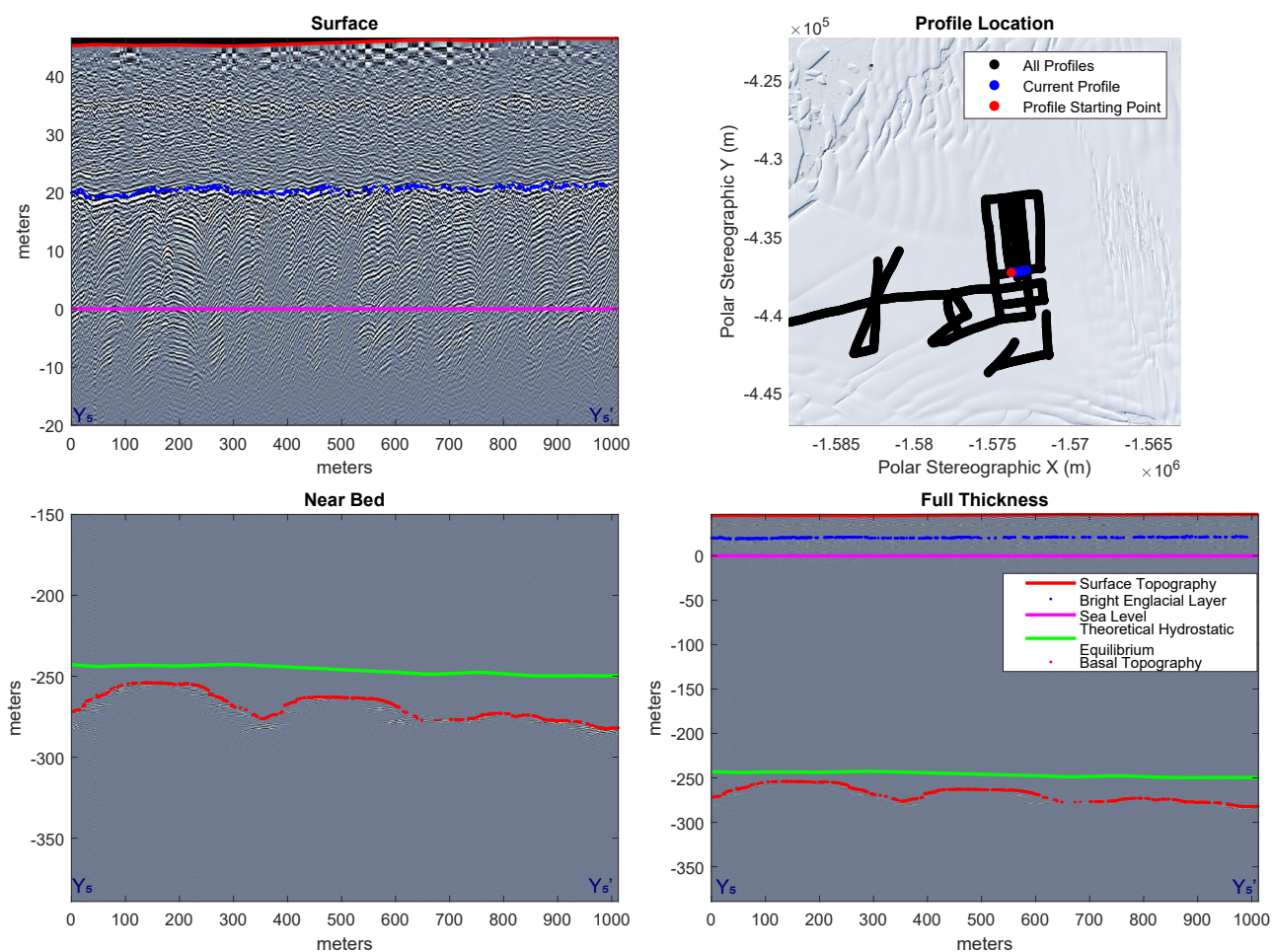


Fig. S16. Profile Y5. A 100 MHz profile from the 2019/20 field season.

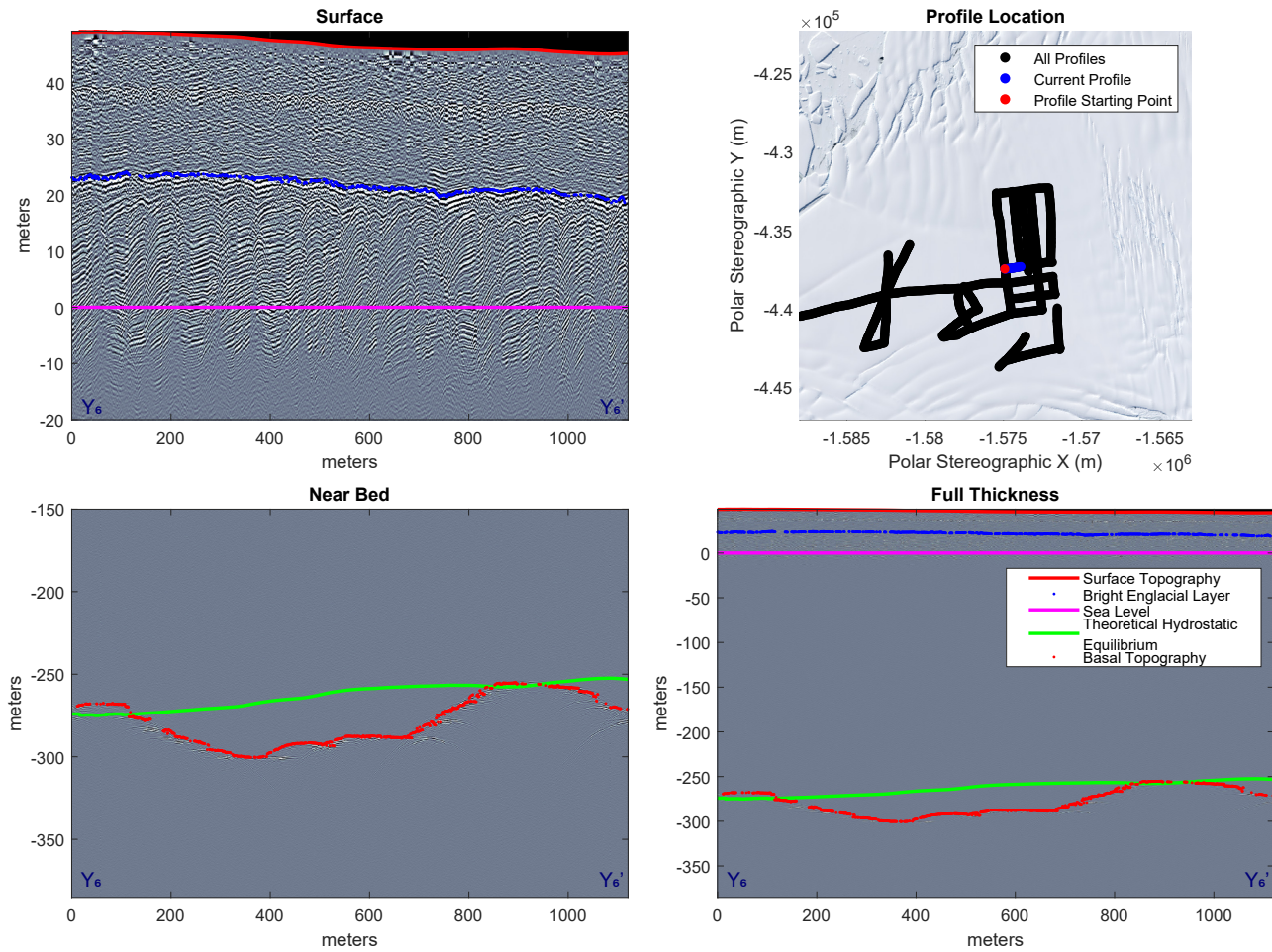


Fig. S17. Profile Y6. A 100 MHz profile from the 2019/20 field season.

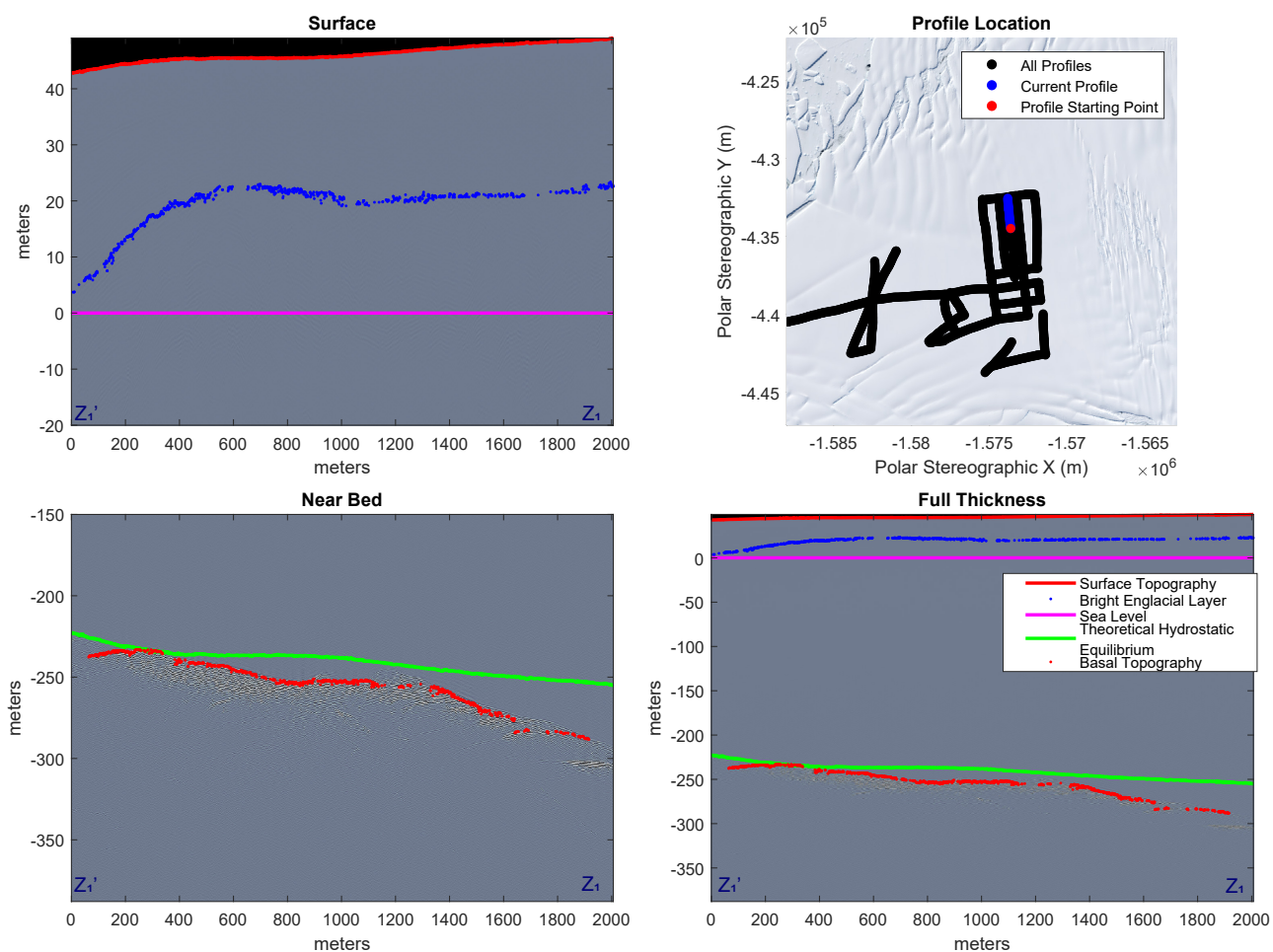


Fig. S18. Profile Z1. A 100 MHz profile from the 2019/20 field season.

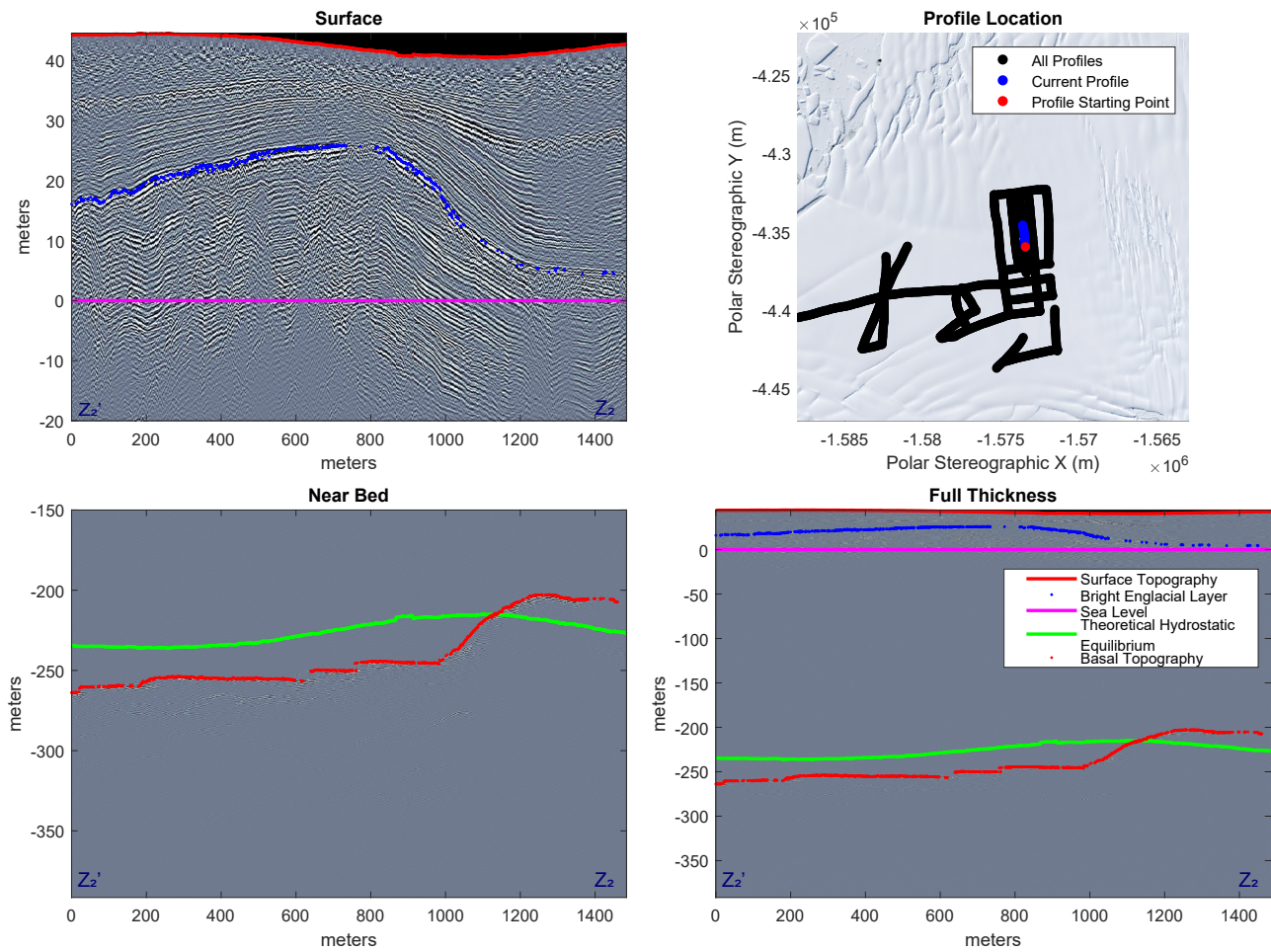


Fig. S19. Profile Z2. A 100 MHz profile from the 2019/20 field season.

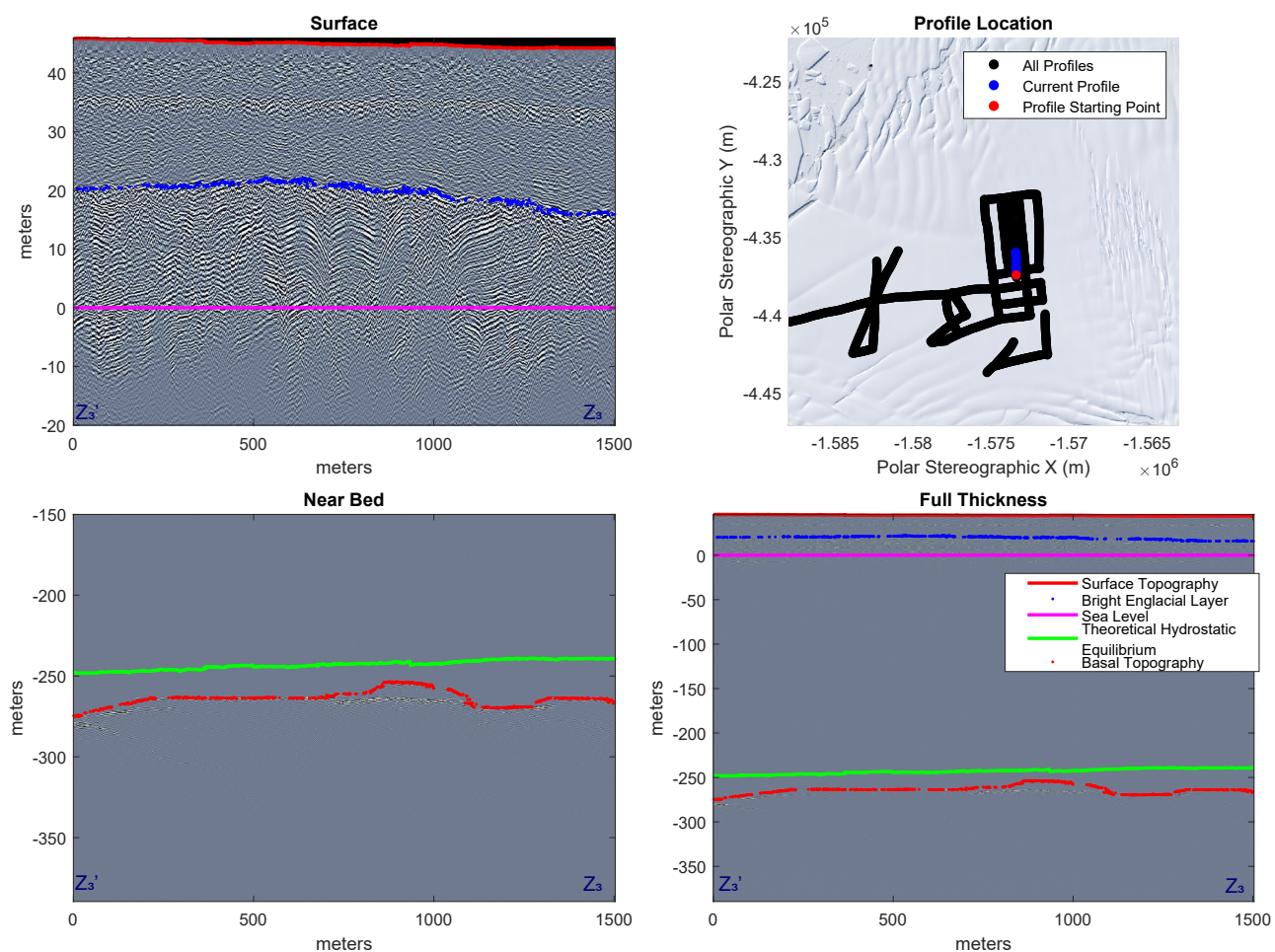


Fig. S20. Profile Z3. A 100 MHz profile from the 2019/20 field season.

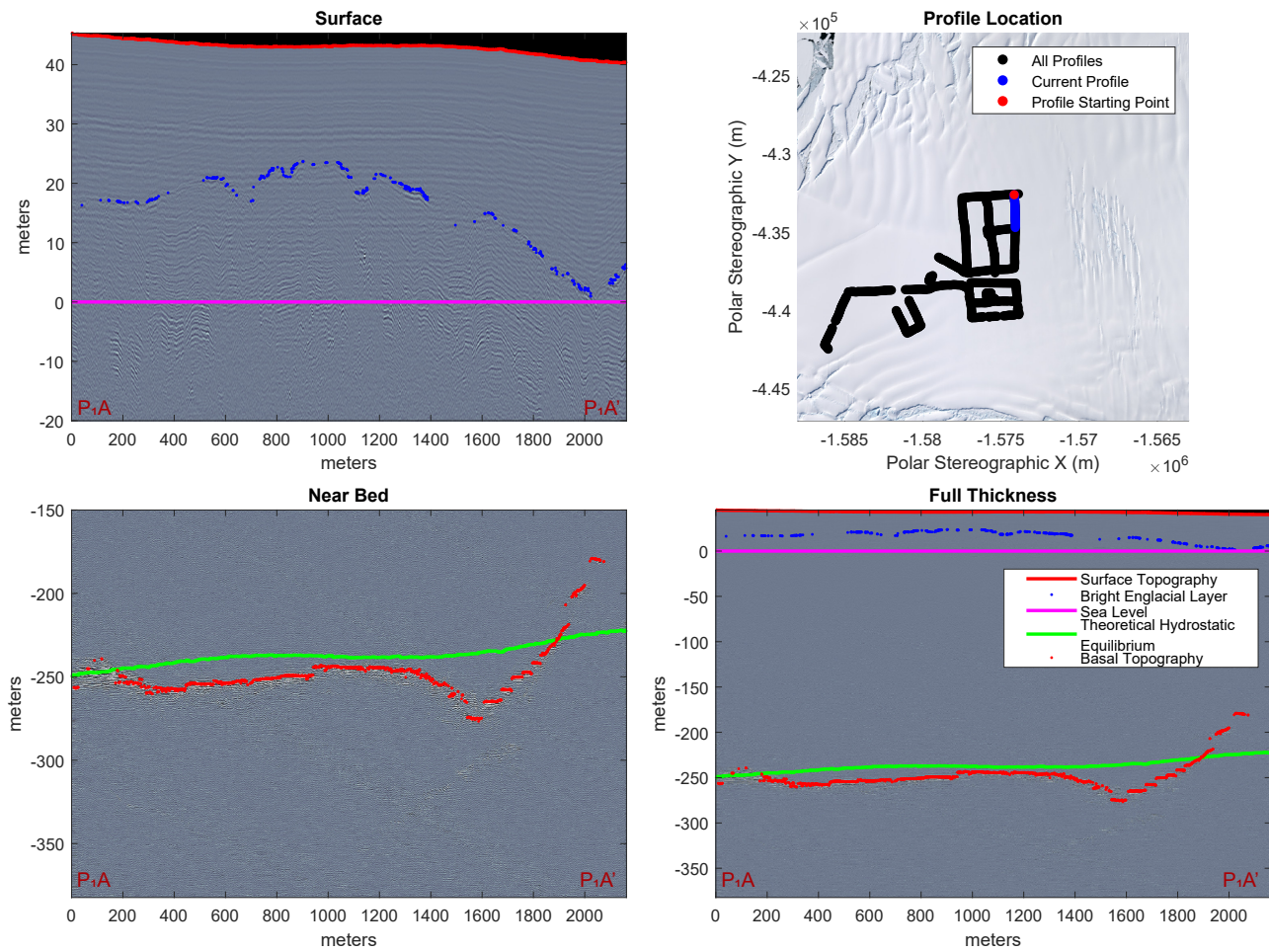


Fig. S21. Profile P1A. A 100 MHz profile from the 2022/23 field season.

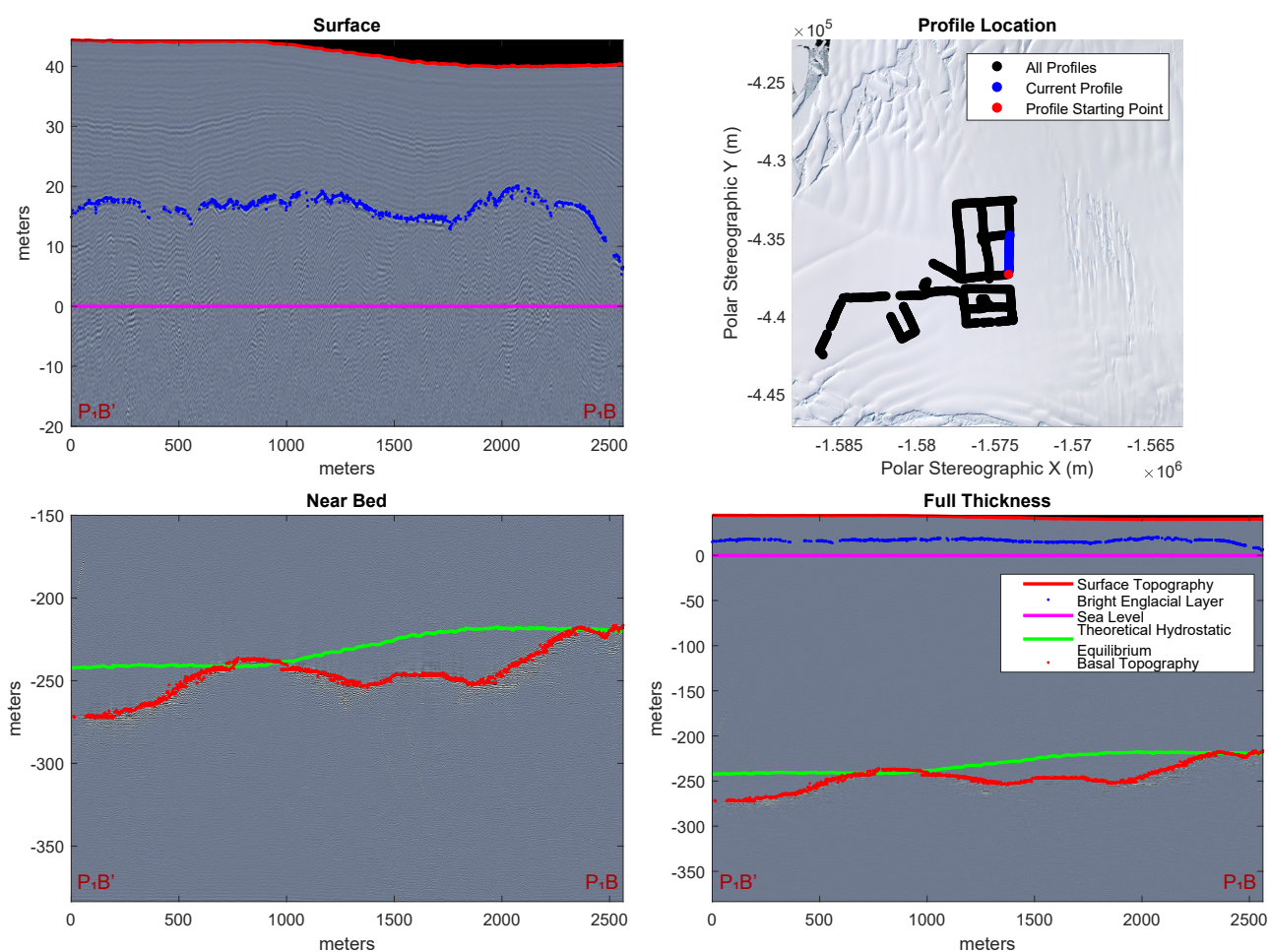


Fig. S22. Profile P1B. A 100 MHz profile from the 2022/23 field season.

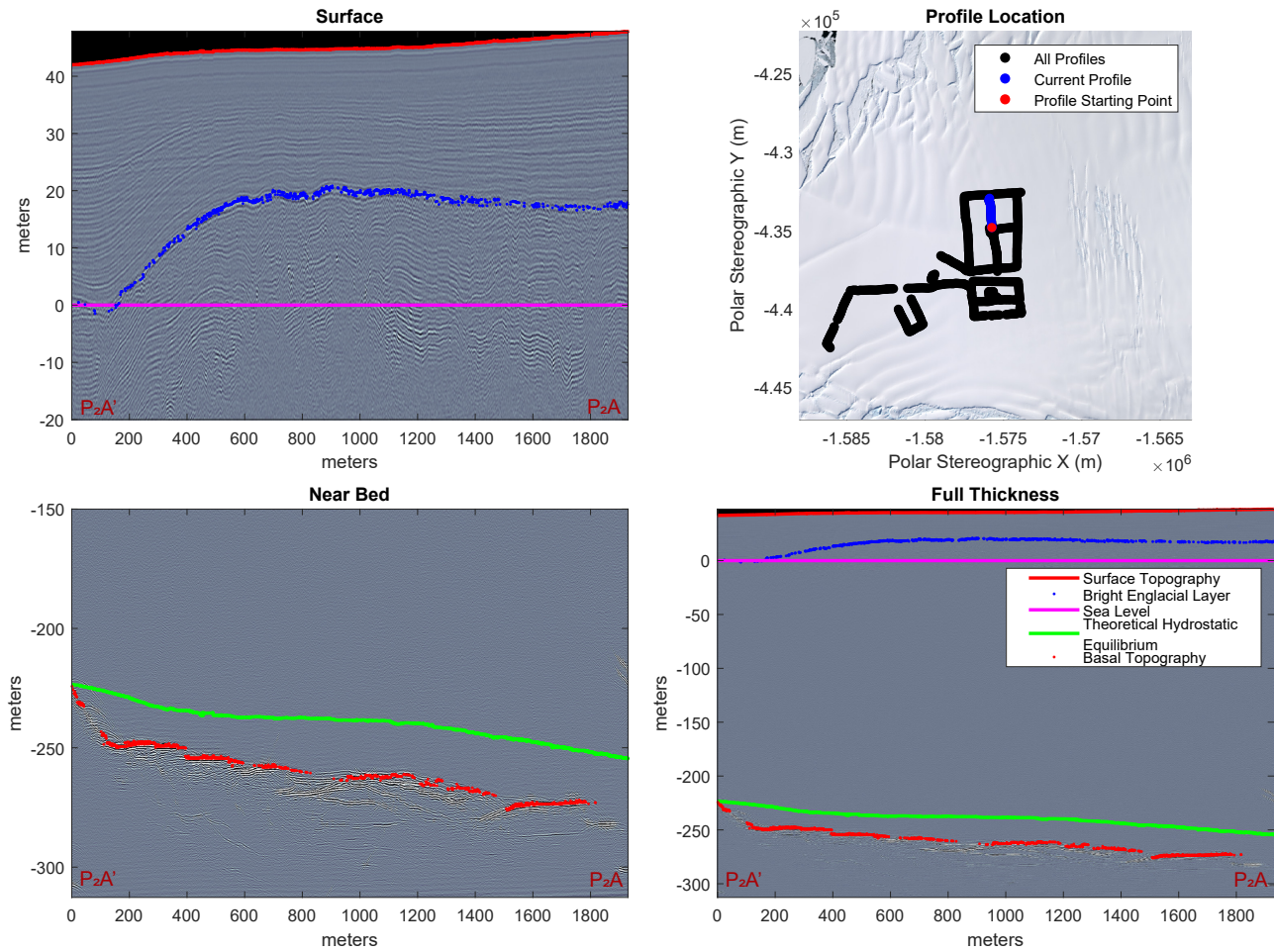


Fig. S23. Profile P2A. A migrated 100 MHz profile from the 2022/23 field season.

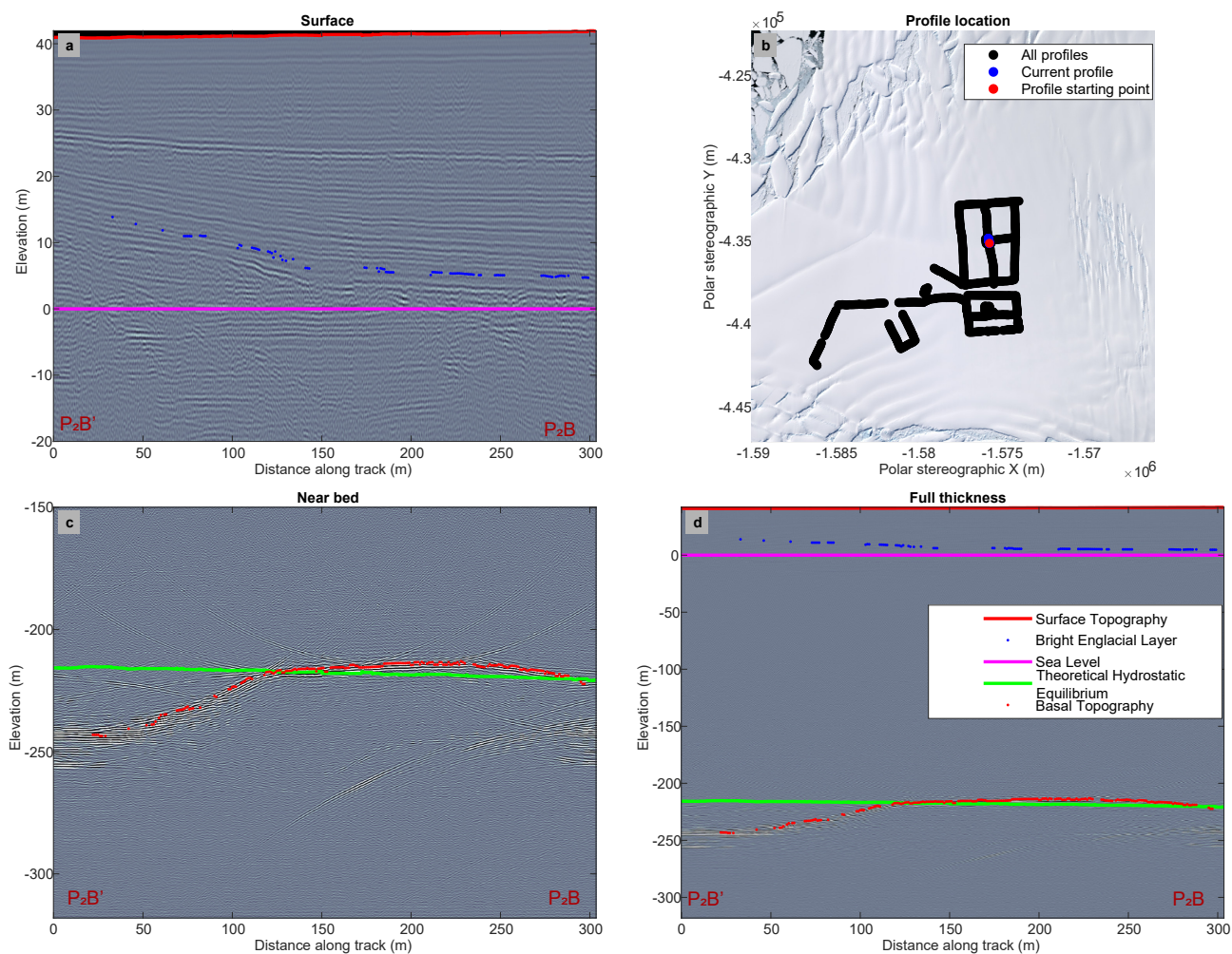


Fig. S24. Profile P2B. A 100 migrated MHz profile from the 2022/23 field season.

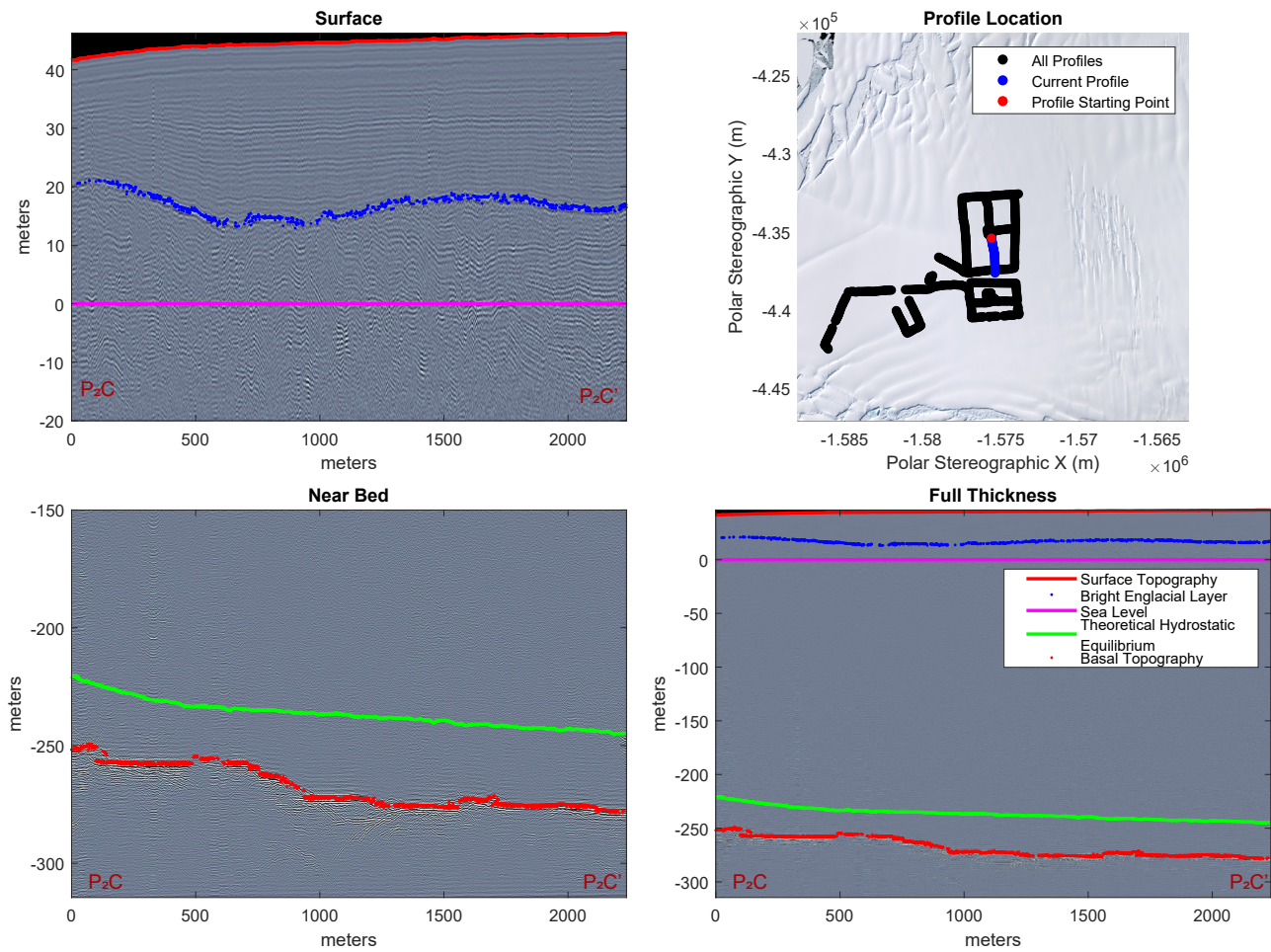


Fig. S25. Profile migrated P2C. A 100 MHz profile from the 2022/23 field season.

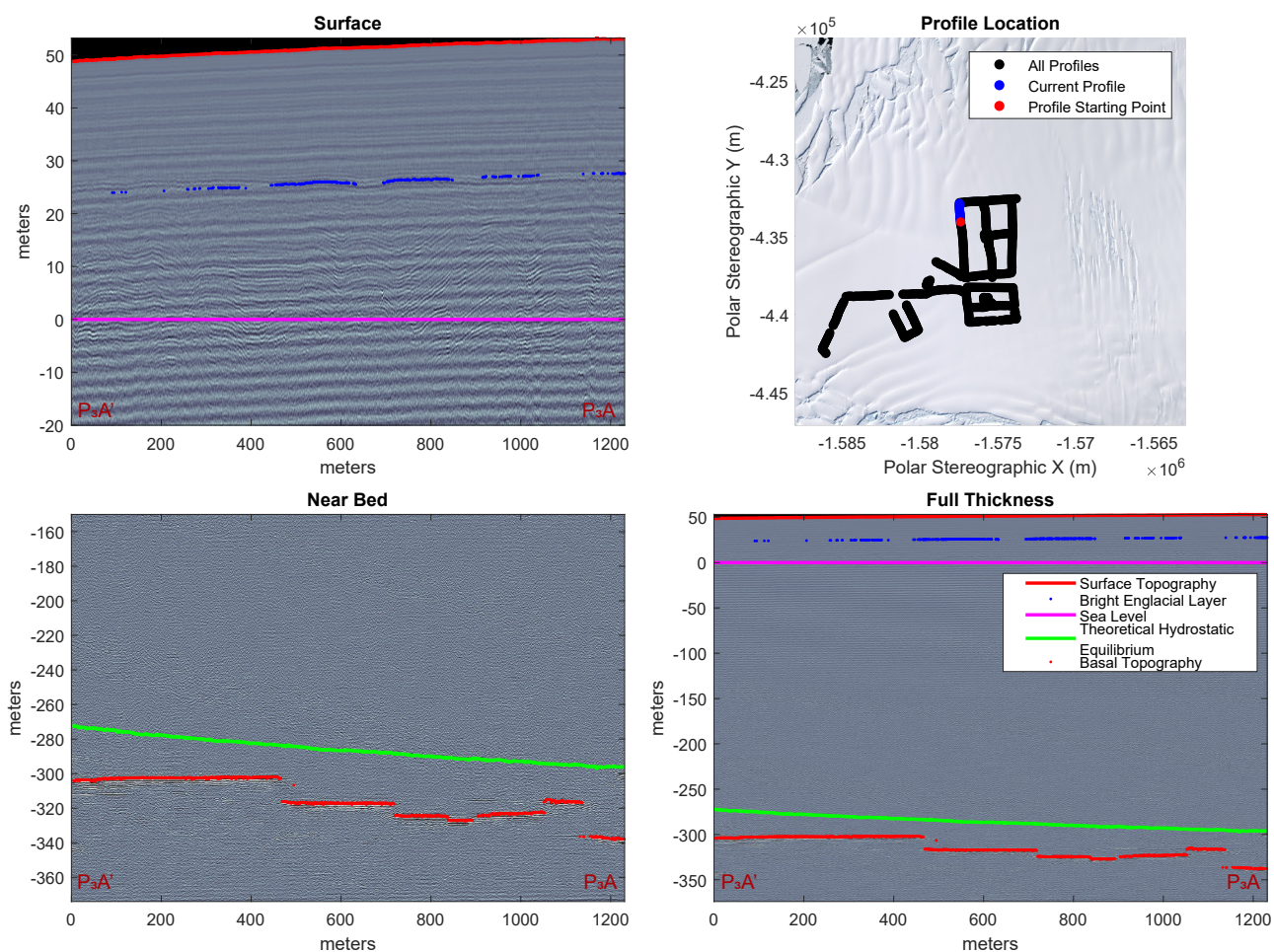


Fig. S26. Profile P3A. A 100 MHz profile from the 2022/23 field season.

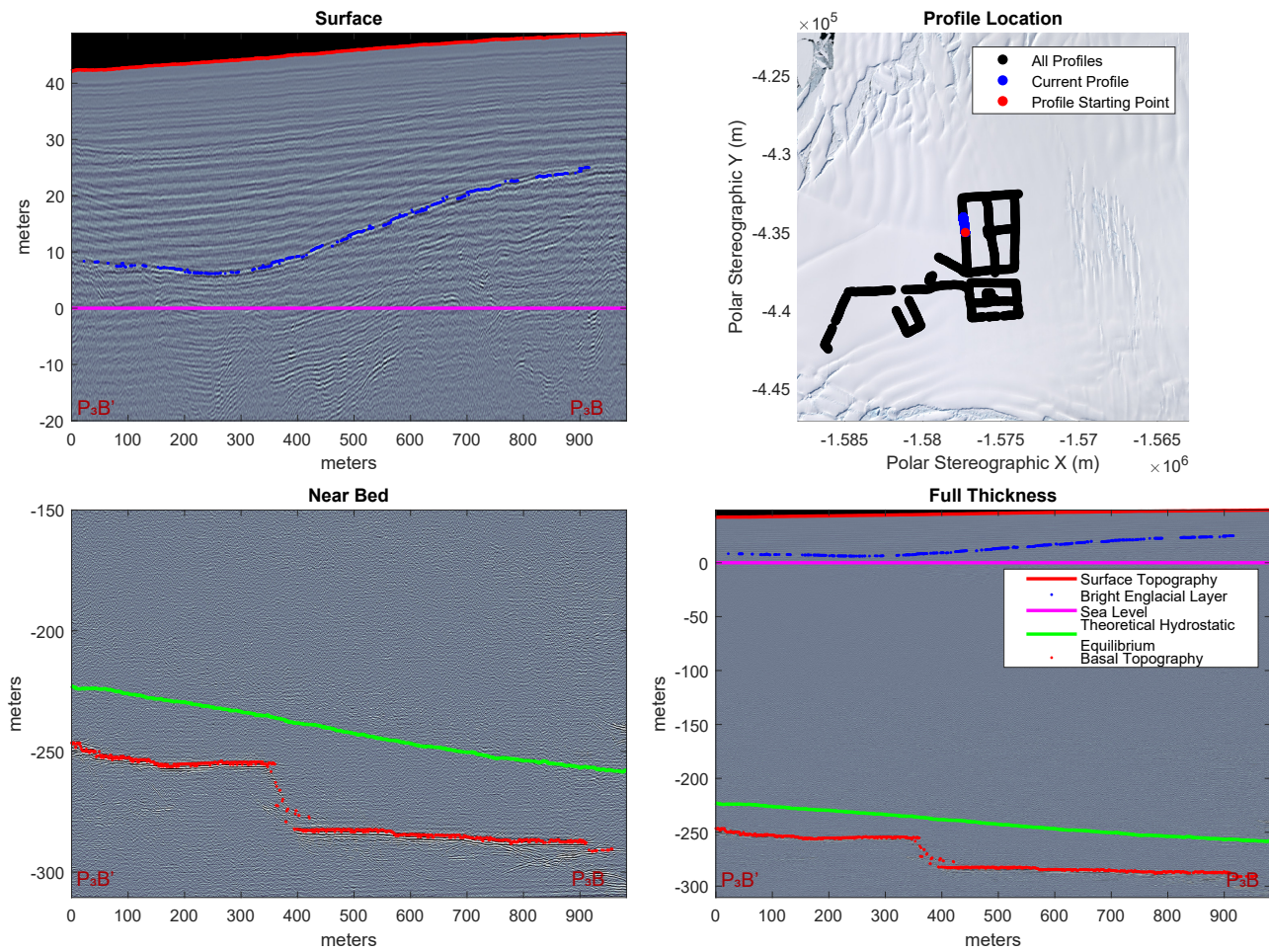


Fig. S27. Profile P3B. A migrated 100 MHz profile from the 2022/23 field season.

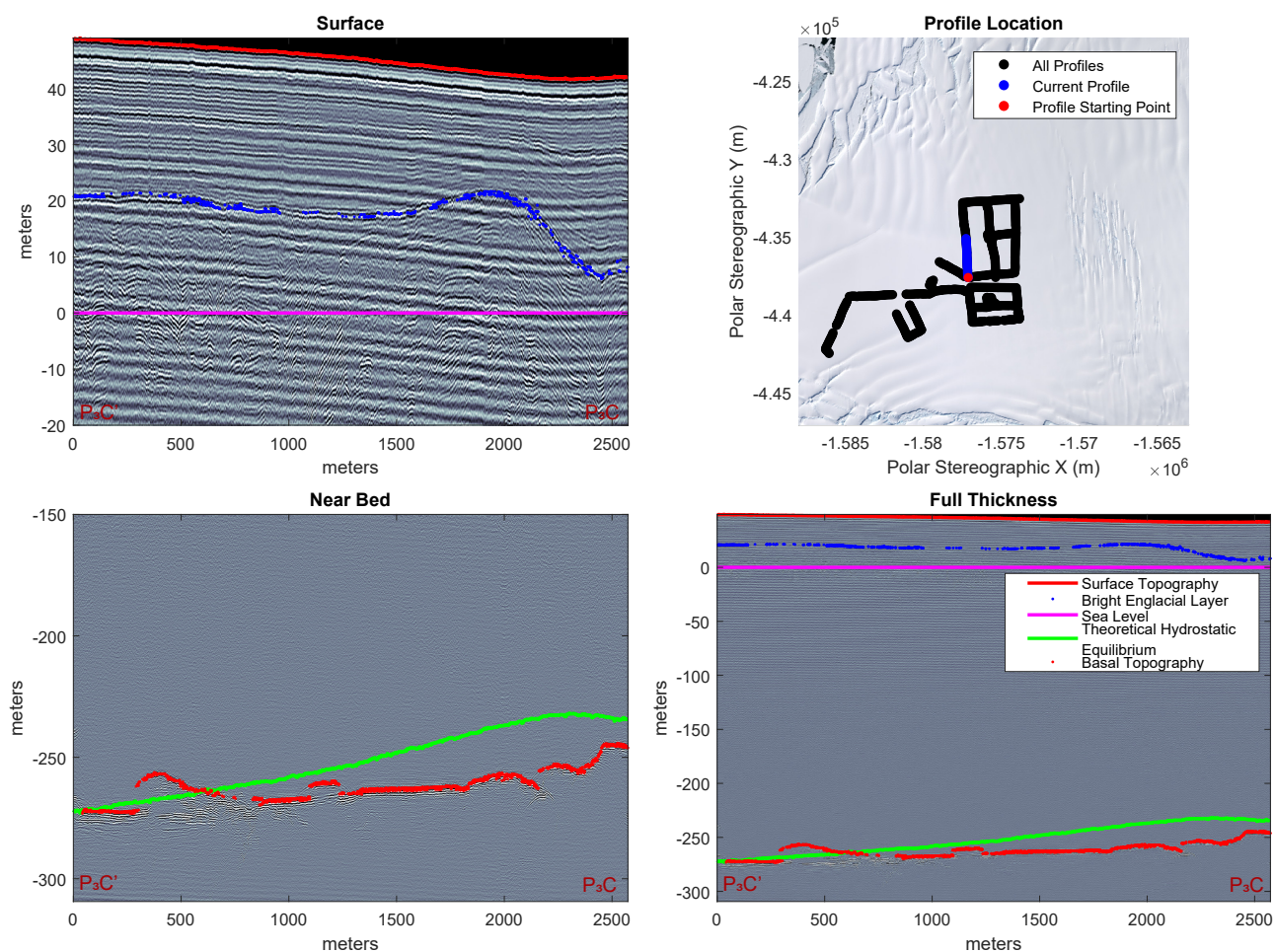


Fig. S28. Profile P3C. A 100 MHz profile from the 2022/23 field season.

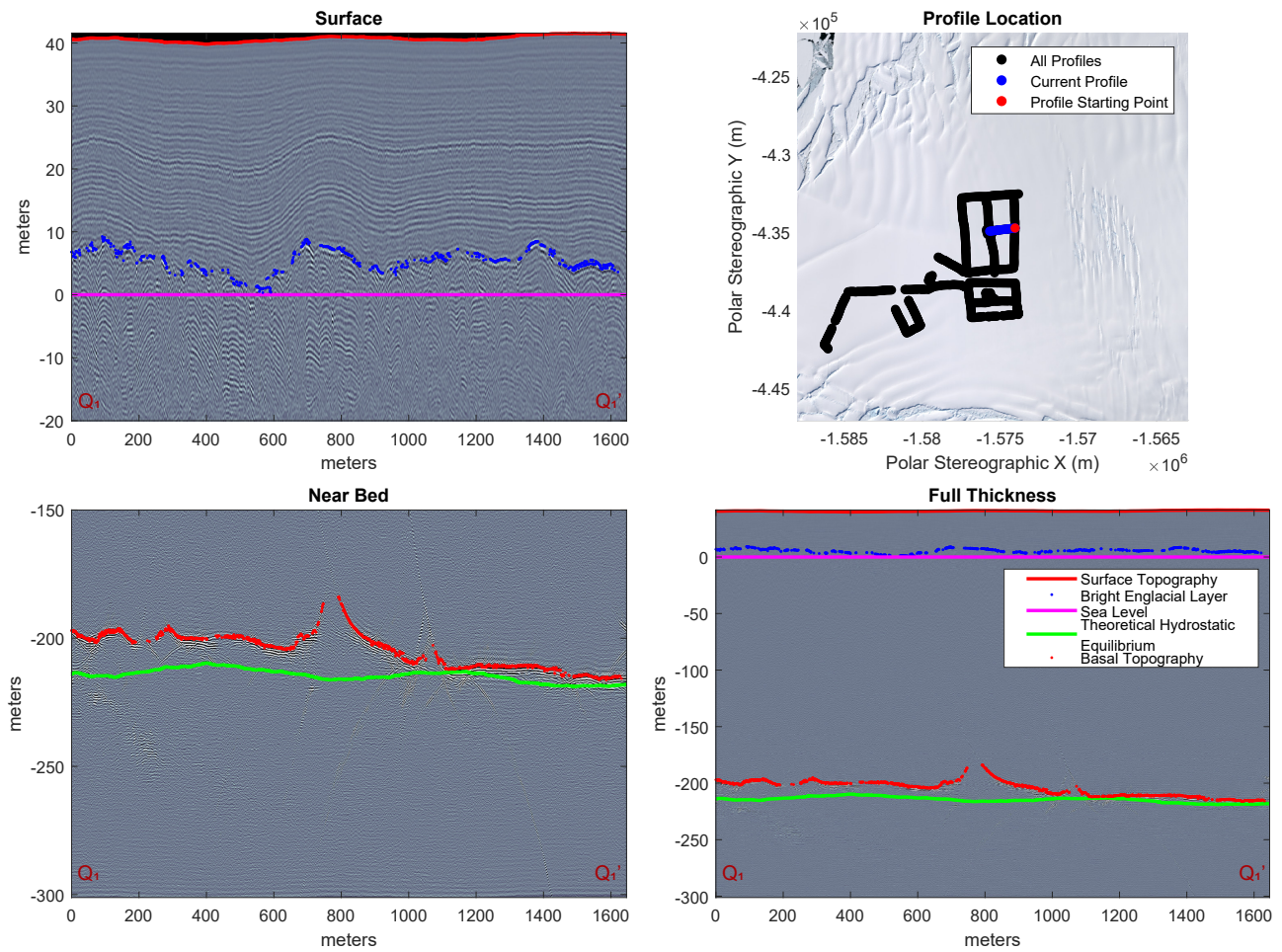


Fig. S29. Profile Q1. A migrated 100 MHz profile from the 2022/23 field season.

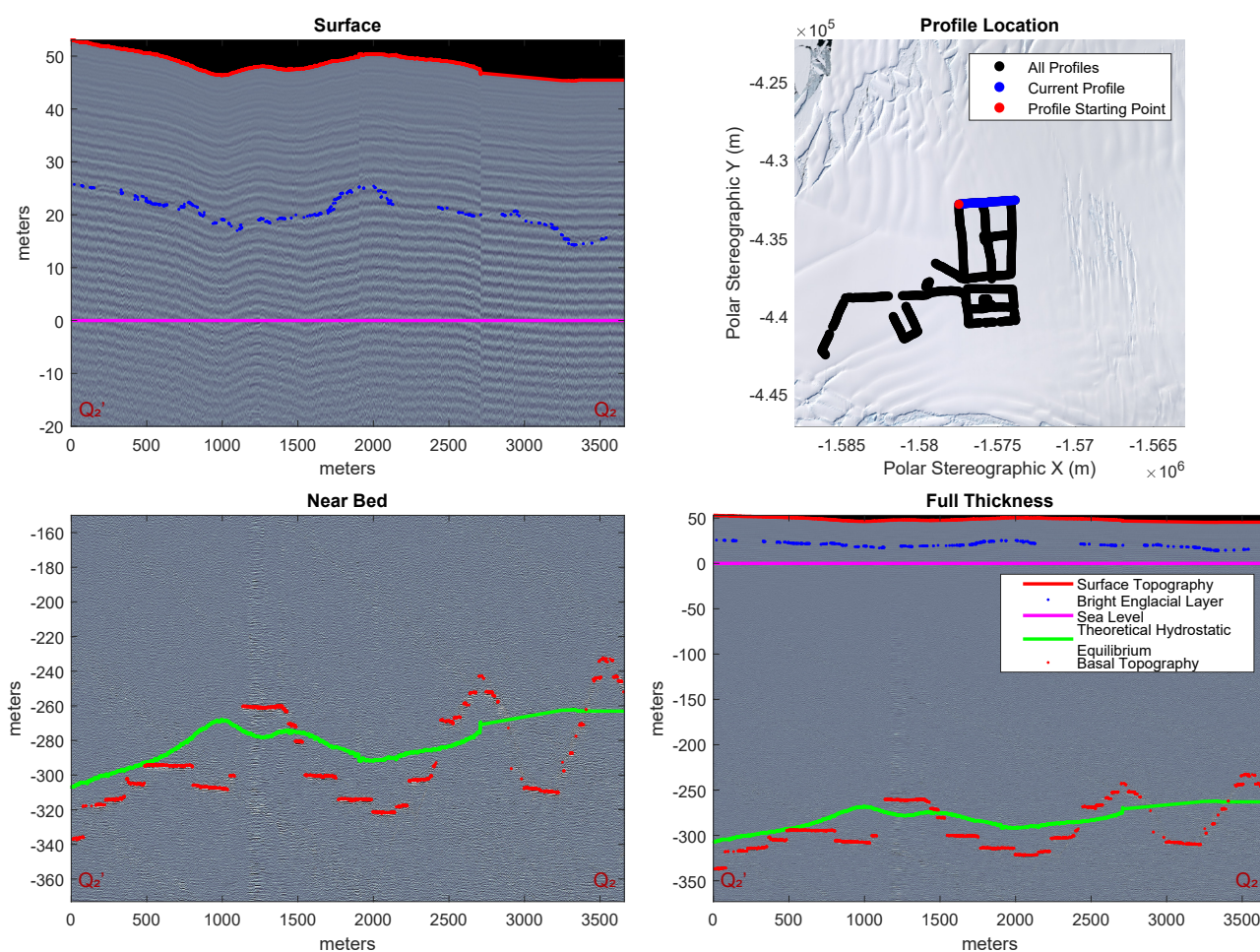


Fig. S30. Profile Q2. A 100 MHz profile from the 2022/23 field season.

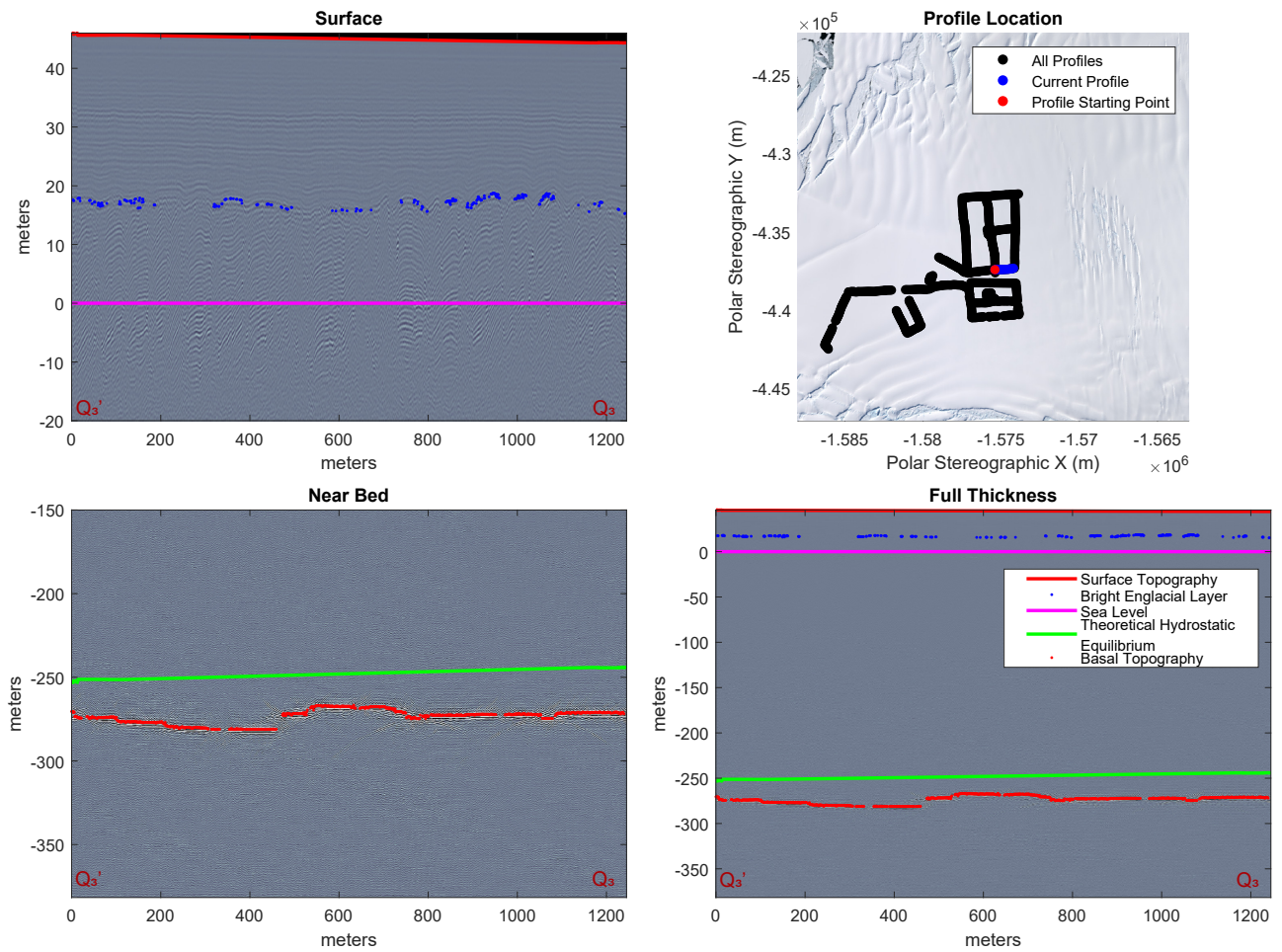


Fig. S31. Profile Q3. A migrated 100 MHz profile from the 2022/23 field season.

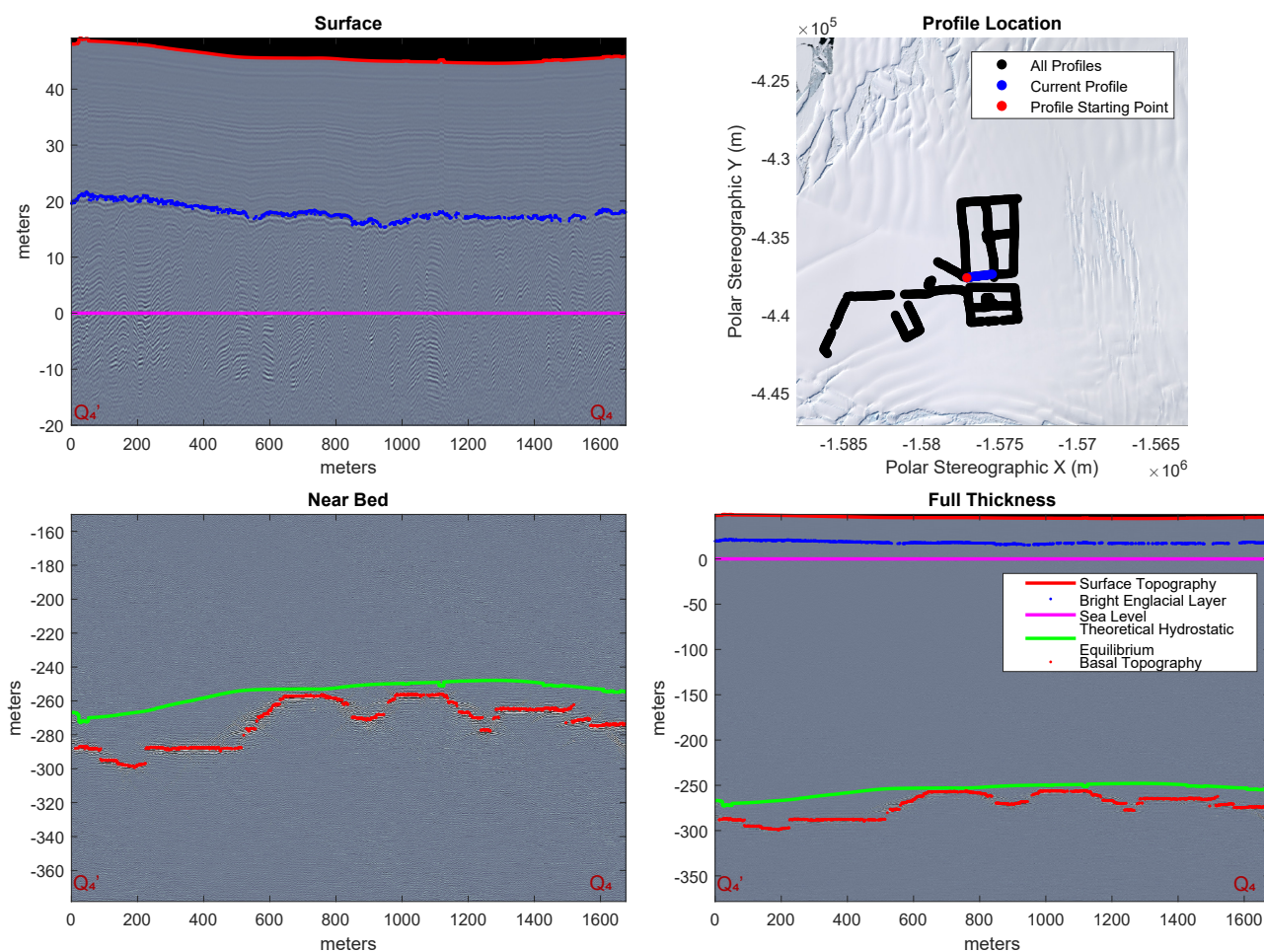


Fig. S32. Profile Q4. A migrated 100 MHz profile from the 2022/23 field season.

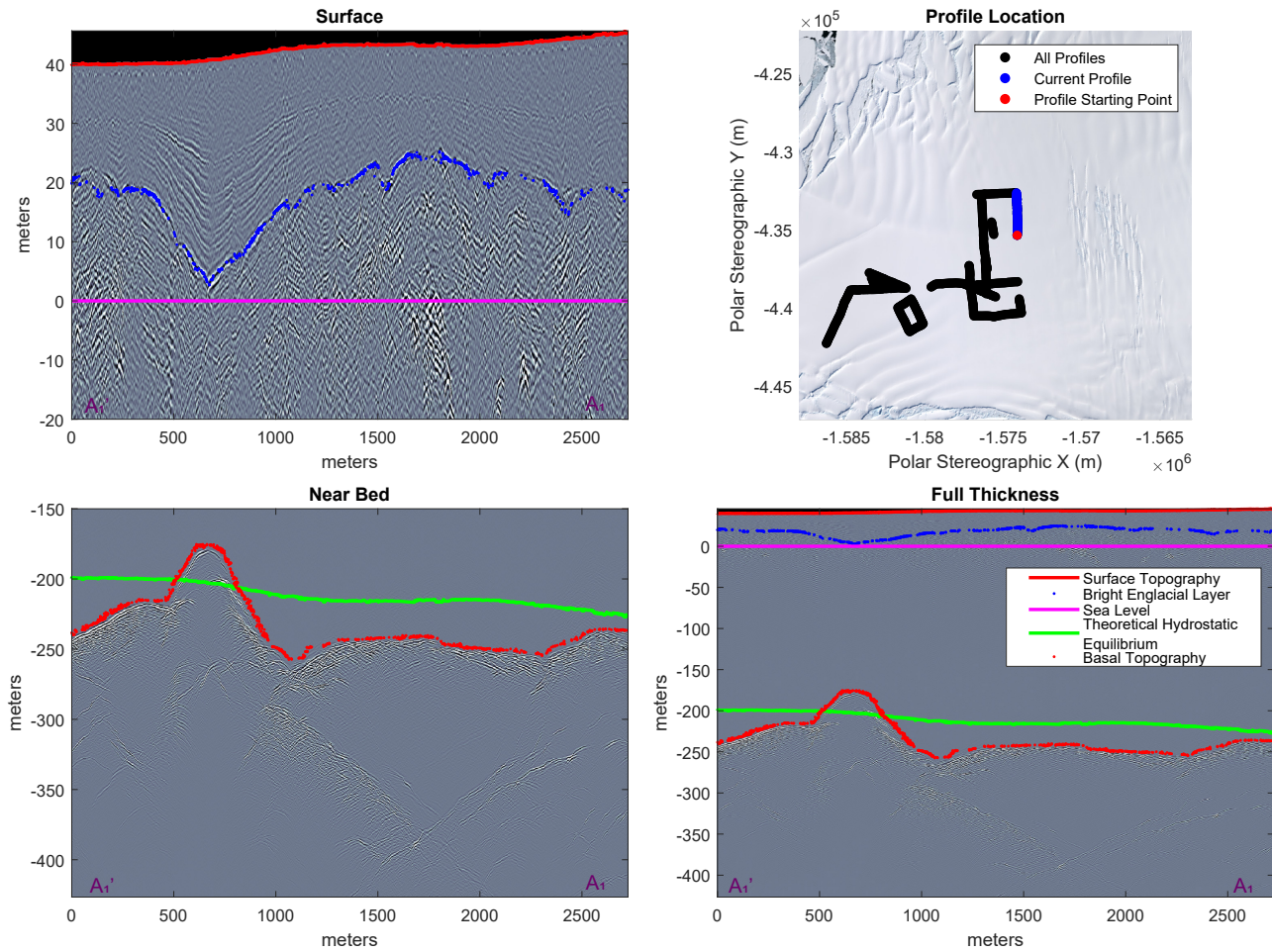


Fig. S33. Profile A1. A 20 MHz profile from the 2022/23 field season.

2022/23 Profiles (20 MHz)

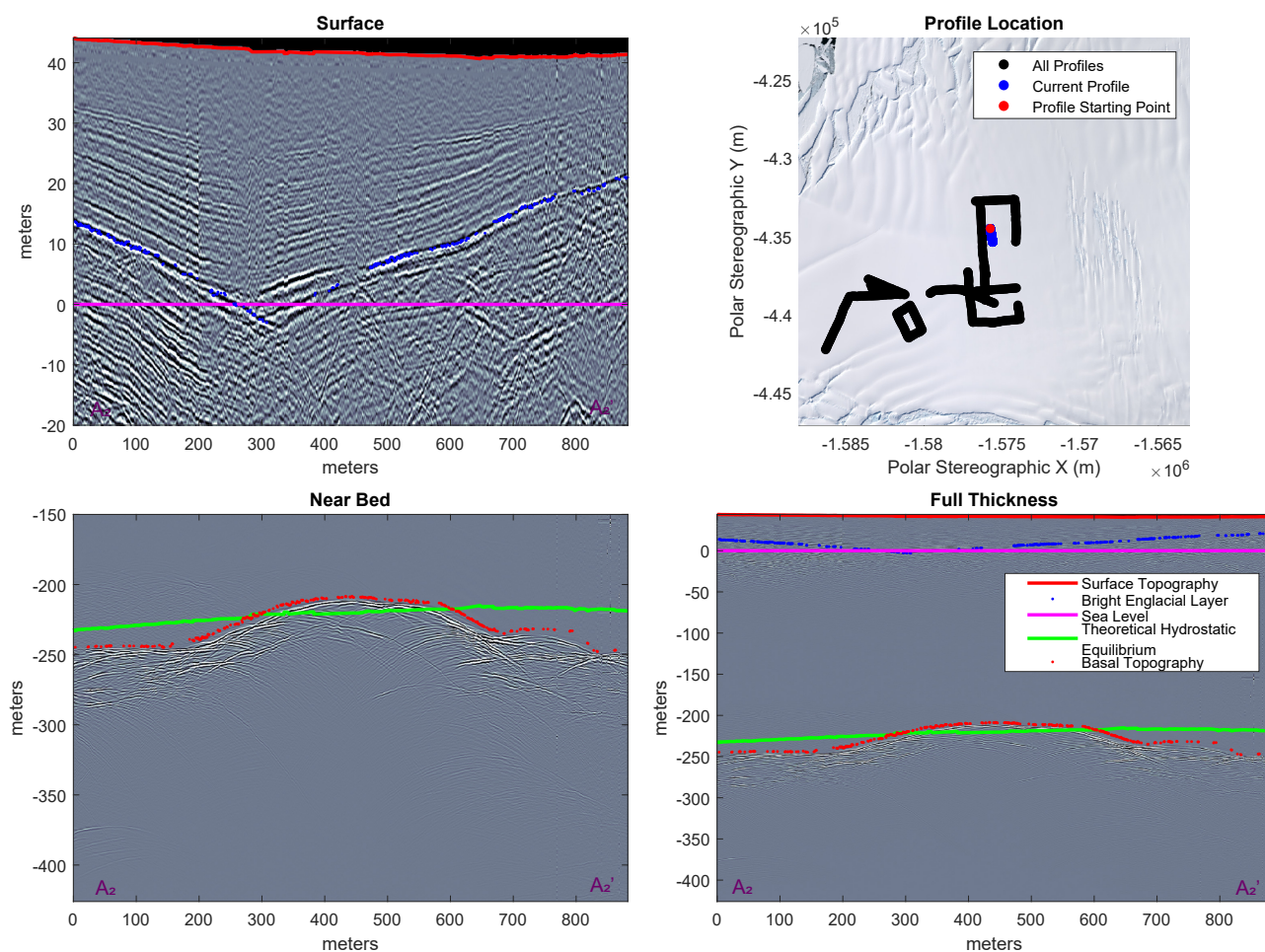


Fig. S34. Profile A2. A 20 MHz profile from the 2022/23 field season.

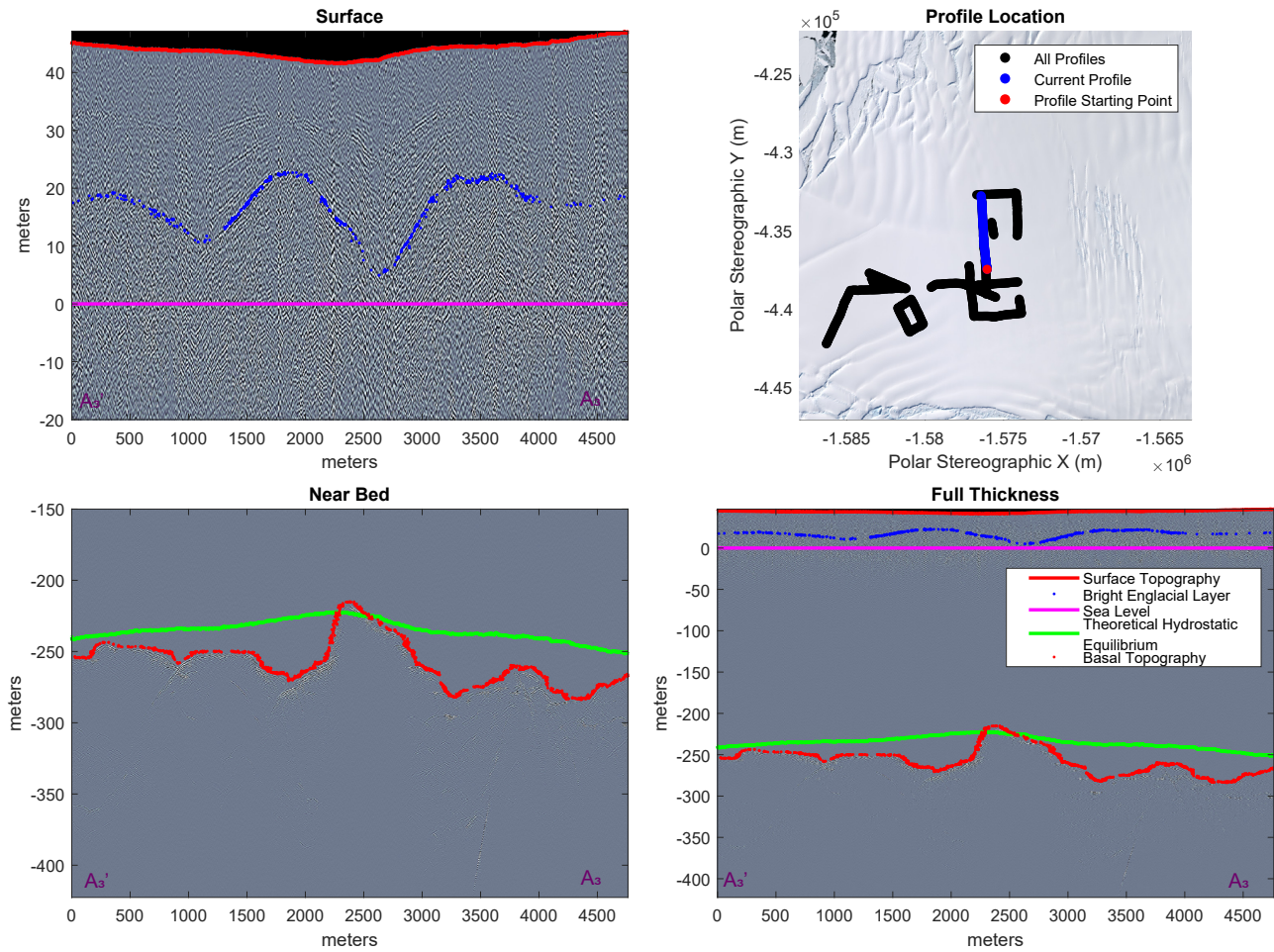


Fig. S35. Profile A3. A 20 MHz profile from the 2022/23 field season.

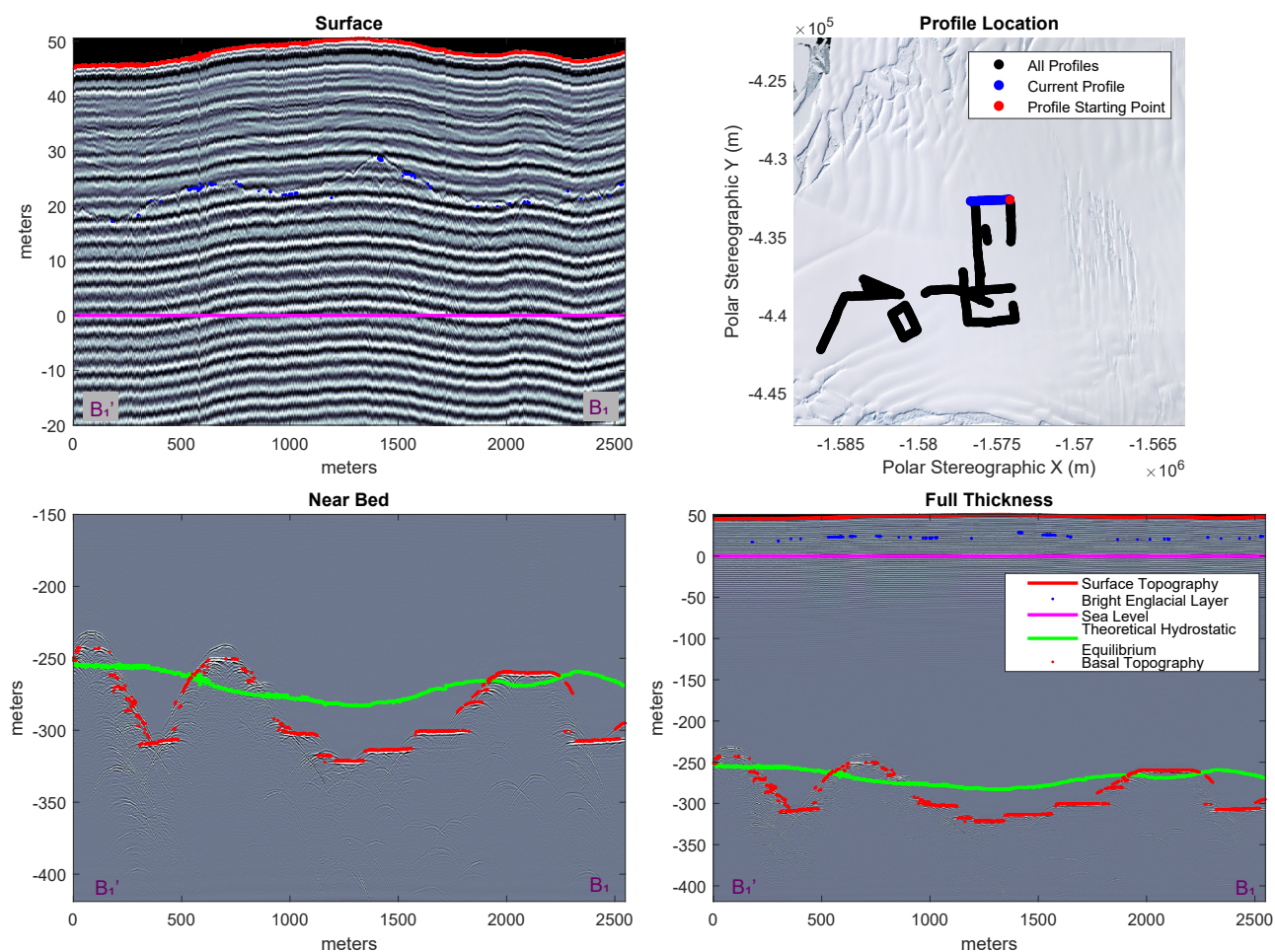


Fig. S36. Profile B1. A 20 MHz profile from the 2022/23 field season.

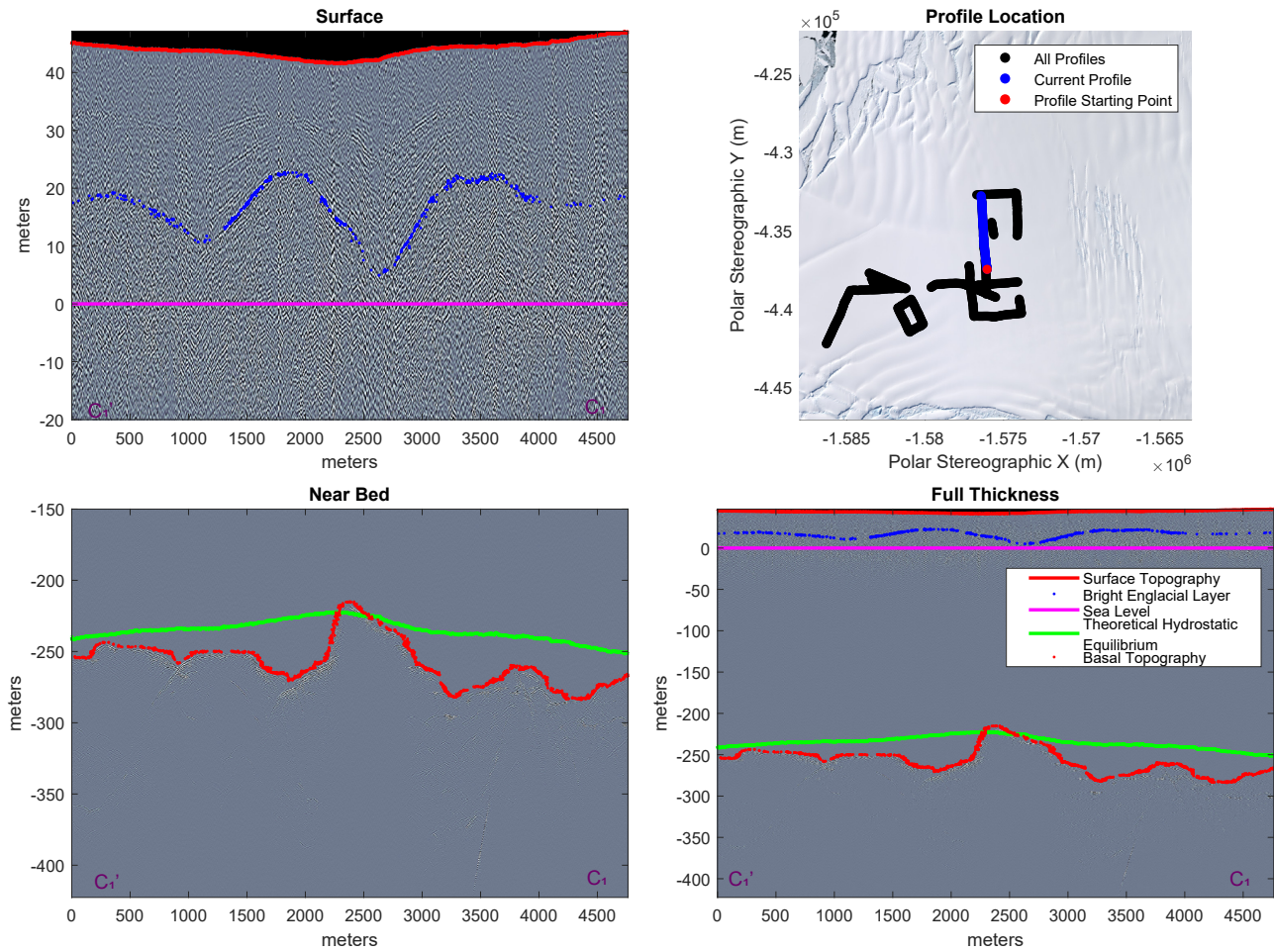


Fig. S37. Profile C1. A 20 MHz profile from the 2022/23 field season.

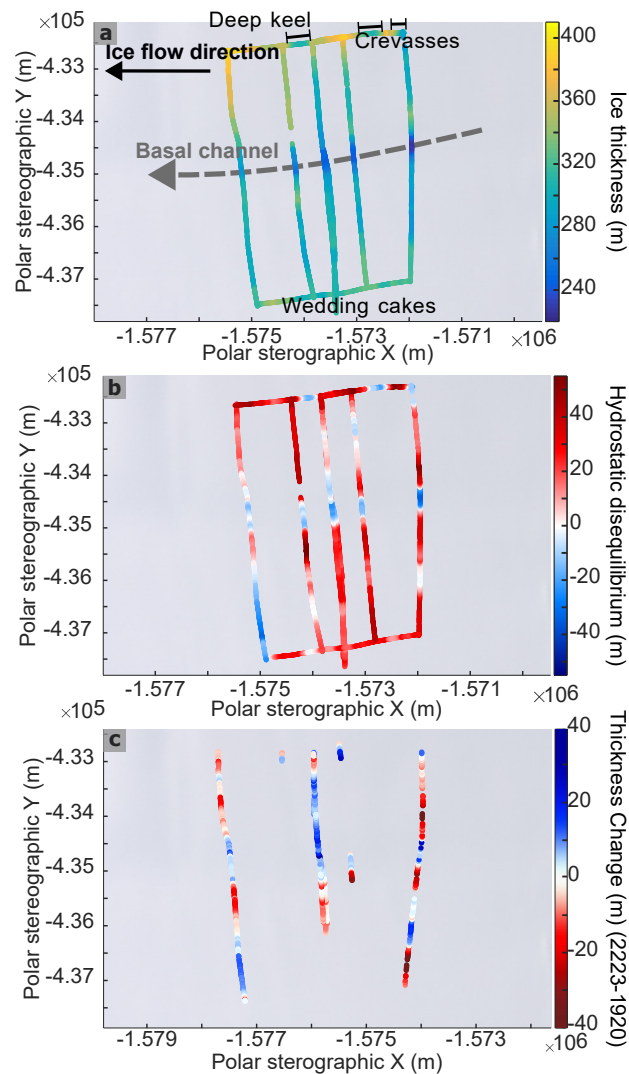


Fig. S38. Panel a shows ice thickness from the 2019/20 field season. Lower thickness areas generally correspond to either the channel or crevasses, and the highest thicknesses are located over the terrace features in the northeast (upper left) of the area. **b** displays hydrostatic disequilibrium from 2019/20. These values are obtained by using surface elevations to estimate where the base of the ice shelf should be if the shelf were in hydrostatic equilibrium, and differencing from the actual bed depths. Positive values correspond to where the bed is below equilibrium and wants to rise, and negative values correspond to where the bed is above equilibrium and wants to sink. **c** shows the change in thickness from the 2019/20 season to the 2022/23 season. Negative values show where the ice thinned (red), while positive values show where the ice thickened (blue). Due to basal crevasses and complex 3D topography, small differences in repeat radar track can lead to large apparent changes in thickness; therefore, details in these results should be used with caution. The blue in the lower left of both panels b and c, for example, is due to the radar tracks following the strike of a crevasse. The general pattern of blue within the channel (above hydrostatic equilibrium/thickening in panels b/c) within the channel on all profiles compared to primarily red (below hydrostatic equilibrium/thinning) outside of the channel is real.

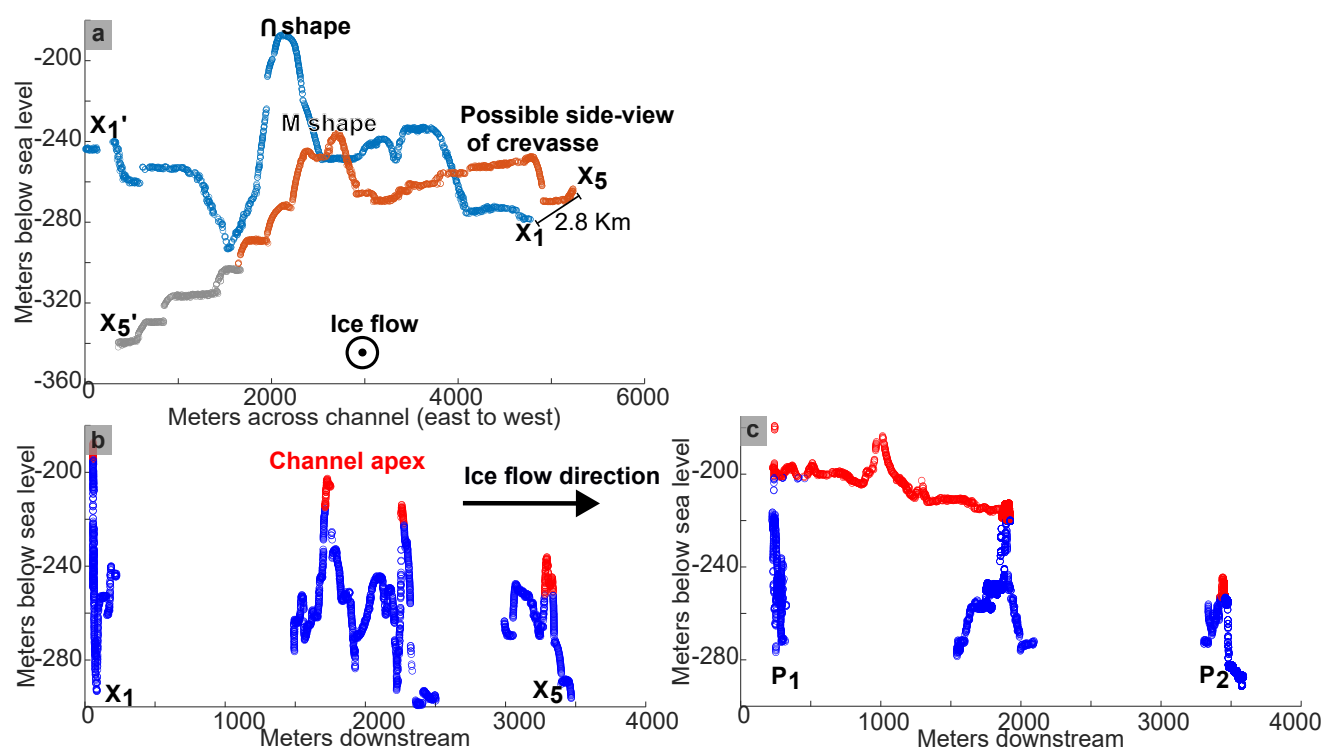


Fig. S39. Multiple perspectives on across-channel profiles. **a** (upstream view) shows the most upstream (in blue) and downstream (in orange) profiles from the 2019/20 season. **b** and **c** display side views of the across-channel profiles, from 2019/20 and 2022/23, respectively. In both panels, the most upstream profile is on the left, and the most downstream is on the right. The apex of the channel is highlighted in red for visual clarity. The gray points in panel **a** represent points that were removed from panel **b** for clarity.

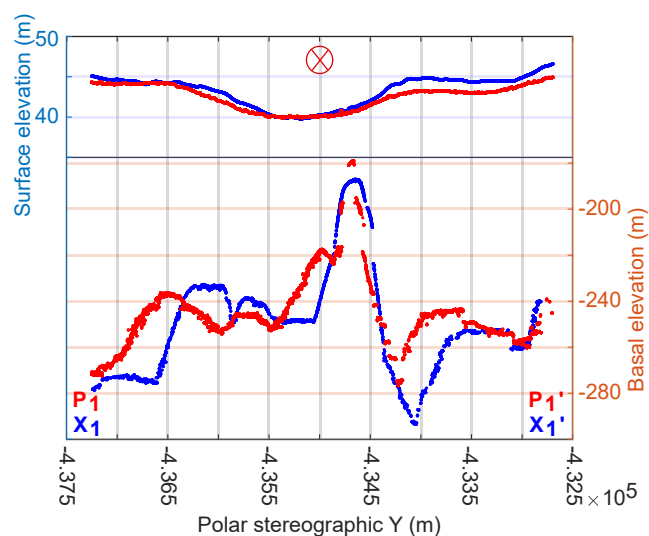


Fig. S40. Comparison of the furthest upstream profiles from the 2019/20 (X1, blue) and 2022/23 (P1, red) seasons. These are a Lagrangian pair, as described in Table S1. Surface topography is shown on top, and basal topography on the bottom. Particularly on the right (east) side of the surface topography, the surface lowers between field seasons as the ice thins due to dynamics and melt ([melt rates estimated between 1 and 2 m a⁻¹ outside the channel (Wild and others, 2024)

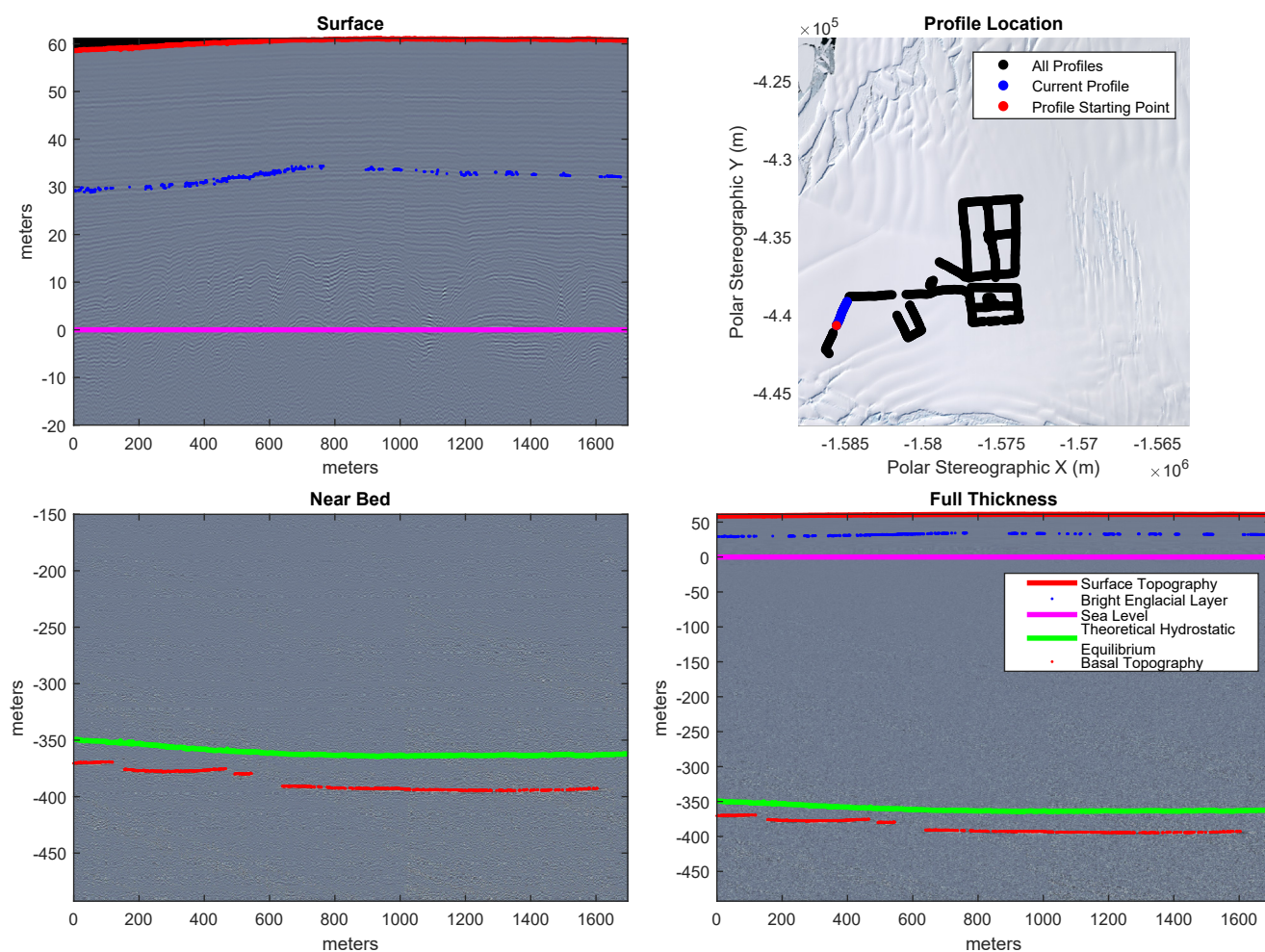


Fig. S41. Example 2019/20 season 100 MHz radar profile across thicker downstream area of TEIS. Showing smoother bed topography and no obvious folding of internal layers.