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Key Points:

- We establish a new method for dense array passive seismic reflection imaging of glacier beds using surface icequakes
- We retrieve a 2D ice thickness map involving no spatial interpolation and yielding vertical resolution similar to active seismic surveys
- We image an overdeepening likely favoring subglacial lake formation, consistent with prior surface elevation change observations

Supporting Information:

Supporting Information may be found in the online version of this article.

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Mapping Glacier Bed Topography in Crevassed Regions: A New Passive Seismic Reflection Method Applied to Isunnguata Sermia, West Greenland

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Abstract Glacier bed topography is crucial for modeling ice mass loss. However, heavily crevassed regions are challenging to survey using conventional methods, leading to data gaps. We establish a passive seismic approach exploiting naturally occurring crevassing events recorded with a dense seismic array to image the ice–bed interface without spatial interpolation. Using P-wave reflections from shallow events detected by matched field processing, we derive a two-dimensional ice-thickness map with ~15 m vertical resolution in a crevasse field on Isunnguata Sermia, West Greenland. Our measurements are consistent with ground-penetrating radar ice thickness picks and extend coverage to previously inaccessible areas. The resolved bed shows a ~100 m deep overdeepening along ice flow, poorly resolved in large-scale bed elevation products, co-located with a previously inferred active subglacial lake. This method provides a robust, logistically simpler tool than more conventional radar and active seismic surveys for high-resolution bed imaging in dynamically active glacier regions.

Plain Language Summary Mapping the topography of bedrock beneath glaciers is essential for predicting the evolution of ice sheets. However, mapping topography is challenging in crevassed areas: traditional methods such as ground-penetrating radar are ineffective due to logistical difficulties and signal loss. In this study, we image topography beneath a West Greenland glacier with a novel approach using natural seismic sources called “icequakes”, caused by crevasses opening at the ice surface. We record these signals with a dense network of seismic sensors and can detect reflections from the glacier bed, thus allowing detailed mapping of ice thickness and bed shape. Our results match and extend radar measurements. We identify a local topographic depression that is significantly deeper than predicted by large-scale mapping initiatives, highlighting the potential of this technique to yield new insight for dynamically active glacier systems.

1. Introduction

Knowledge of glacier bed topography is fundamental to understanding and predicting ice-sheet mass loss. Specifically, variations in bed topography exert a first-order control on ice dynamics: they drive deformation by inducing complex stress concentration patterns within the ice column (Law et al., 2021; Weertman, 1964) and modulate basal sliding speeds by regulating the storage and routing of subglacial water that governs basal lubrication (Creyts et al., 2013; Livingstone et al., 2022; Siegfried et al., 2016; Stearns et al., 2008; Zhan, 2019). Although airborne radar and gravimetry have transformed our understanding of Greenland's large-scale bed geometry (An et al., 2017; Lindbäck et al., 2014), significant challenges remain in resolving marginal areas where the ice is thin, warm and its surface often crevassed (Santin et al., 2026). Consequently, bed elevation is frequently reconstructed from spatially sparse observations in these regions. For example, BedMachine combines ice thickness measurements with a mass conservation approach to infer bed topography in regions lacking direct observations (Morlighem et al., 2017), whereas the Greenland Analogue Project (GAP) bed model relies on geostatistical interpolation (Lindbäck et al., 2014). As a result, short-wavelength features (10^1 – 10^2 m) remain unresolved in these gridded products, which are provided at 150 m (BedMachine v5) (Morlighem et al., 2017) and 250 m (GAP) (Lindbäck et al., 2014) horizontal resolution. Yet, these smaller-scale features (e.g., undulations,

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overdeepenings) can strongly influence both ice flow dynamics and subglacial hydrology (Law et al., 2021; Swift et al., 2021; Werder, 2016).

A range of conventional geophysical techniques can be employed to infer the smaller-scale bed topography, including surface ground-penetrating radar (GPR) and active seismic surveys. A first limitation is that these techniques are typically conducted along one-dimensional profiles; recovering two-dimensional bed maps requires multiple survey lines and complex interpolation (Hawkins et al., 2026; Morlighem et al., 2017, 2022) with the risk that features not aligned with the acquisition geometry remain poorly resolved. A second limitation is that acquisitions are challenging in highly crevassed, hydrologically active environments where surface meltwater pervades the ice column. However, these areas are crucial targets, as surface crevassing is generated by large lateral stress gradients that can be caused by pronounced bed topography changes (e.g., in ice stream shear margins (Raymond, 1996; Schoof, 2004) or bed overdeepenings (Gimbert, Gilbert, et al., 2021; Herzfeld et al., 2012)) promoting more complex ice deformation as well as englacial and subglacial hydrology dynamics (Law et al., 2021; McGrath et al., 2026; Werder, 2016). GPR signals are strongly attenuated and scattered by crevassing and liquid water (Gimbert, Nanni, et al., 2021; Jezek et al., 2011; MacGregor et al., 2015). In addition, high electrical conductivity prevents propagation through water layers, such that reflections may represent the ice–water interface rather than the bed. Seismic methods, being insensitive to electrical conductivity, are better suited for imaging through temperate ice and water-bearing layers. Active seismic surveys have been widely used to image glacier beds, but their performance in highly attenuating environments, such as temperate ice and crevassed areas, strongly depends on the source strength (Booth et al., 2020; Zechmann et al., 2018). Due to the challenging field logistics necessitated by working in harsh glacial environments, the source choice is often restricted to a sledgehammer, which may only generate detectable body waves over limited distances of a few hundred meters, preventing the detection of bed reflections (Paris, Gimbert, Le Bris, et al., 2026).

Passive seismic approaches based on natural seismic signals recorded by dense arrays of nodes provide a compelling complement to conventional active methods. Because passive deployments rely on compact, light-weight, and easily transportable seismic nodes, they can be carried out even under harsh surface conditions. Consequently, naturally occurring seismic sources that generate highly energetic broadband signals can be recorded in regions that are inaccessible to conventional active surveys (Gimbert, Nanni, et al., 2021). Methods such as horizontal-to-vertical (H/V) spectral ratio and ambient noise surface wave tomography have been used to estimate ice thickness (Picotti et al., 2017; Sergeant et al., 2020; Yan et al., 2018), but they provide limited horizontal and vertical resolution and involve a non-unique inversion of ice thickness (Gimbert, Nanni, et al., 2021).

In this study, we overcome these limitations by leveraging compressional (P) wave bed reflections generated by natural surface crevassing events and recorded by a dense array of seismic nodes. We focus our investigation on the ablation zone of Isunnguata Sermia, West Greenland, where three km-scale active subglacial lakes have been inferred from quasi-circular surface elevation anomalies observed in ArcticDEM data (Livingstone et al., 2019). Specifically, we target a crevasse field at the margin of the largest lake which hindered a previous GPR survey (Hawkins et al., 2026). Using a catalog of seismic sources relocated with matched field processing by Paris, Gimbert, Le Bris, et al. (2026), we construct a two-dimensional map of ice thickness and bed topography that extends into the crevasse field without spatial interpolation. We validate our results against independent GPR measurements at the edge of the crevasse field and discuss the reconstructed bed features given previous knowledge on glacier bed topography as well as the applicability of our method to other sites.

2. Materials and Methods

We use data from the GreenlandSeismic array 9F (Gimbert & RESIF, 2023), a dense seismic network deployed for one month in spring 2023 on the glacier tongue of Isunnguata Sermia, West Greenland (Figure 1a). This array was centered on a ~ 0.5 km² surface crevasse field where ground-based GPR surveys are difficult to conduct (Figure 1b; Hawkins et al. (2026)). It also partially overlies the largest inferred subglacial lake identified by Livingstone et al. (2019) (Figure 1c), referred to therein as Anomaly 2. The array comprised 82 three-component seismic nodes recording continuously at 250 Hz, arranged in a square grid covering approximately 1.2 km², with an average inter-station spacing of 125 m (Figure 1c). In the following subsections, we describe the seismic event catalog and outline the workflow used to image the ice–bed interface from these data.

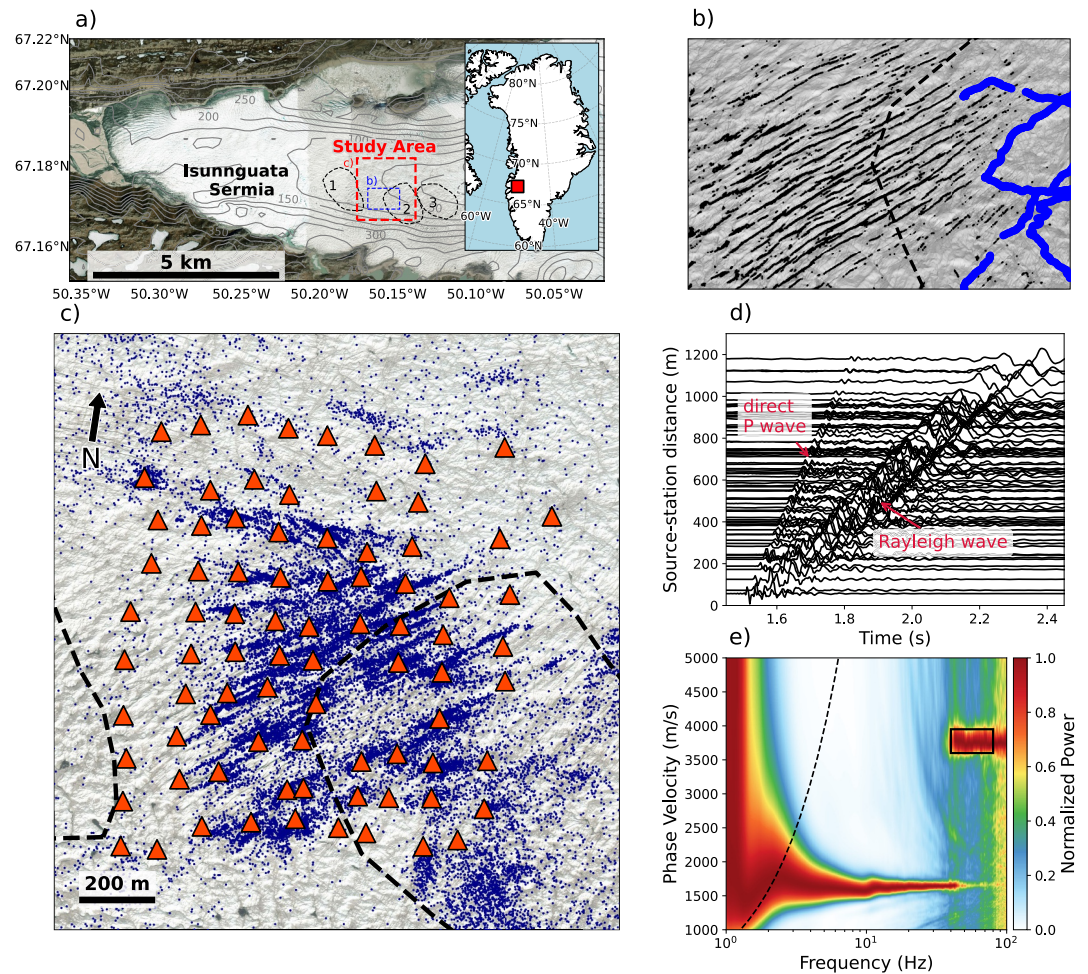


Figure 1. (a) Satellite image of the region around the study area. The black dashed lines correspond to the outlines of subglacial lakes identified in Livingstone et al. (2019). The gray lines show the bed elevation from BedMachine v5 (Morlighem et al., 2022). The blue dashed rectangle indicates the extent of the satellite image in (b). The red dashed rectangle indicates the location of the study area in panel (c). The red square in the inset indicates the location of the satellite image. (b) Pléiades satellite image of the crevasse field. The black lines highlight the crevasses. The blue lines indicate the locations surveyed during the GPR acquisition in Hawkins et al. (2026). (c) Satellite image of the study area. Dark blue dots show surface events (<10 m depth) from Paris, Gimbert, Le Bris, et al. (2026). Red triangles mark the locations of the seismic nodes. (d) Vertical-component unfiltered recordings of a natural seismic event across the node array, sorted by source-receiver distance. (e) Frequency-velocity diagram computed with surface events in Paris, Gimbert, Le Bris, et al. (2026). The black dashed line indicates the frequency-dependent resolution limit, given the maximum wavelength and sensor spacing $\lambda_{\max} = \Delta_{\max}/2$. The black rectangle highlights the velocity and frequency range with high power associated with high SNR P waves.

2.1. Catalog of Seismic Events

We use the 2023 seismic event catalog derived from the GreenlandSeismic array data, described in Paris, Gimbert, Le Bris, et al. (2026). Event detection was performed on the vertical component using matched field processing (MFP) (Nanni et al., 2021, 2022; Schippkus & Hadzioannou, 2022; Umlauf et al., 2021; Walter et al., 2015), an array-based inversion technique that simultaneously estimates source location and seismic velocity by minimizing the error between forward-modeled replicas and the recorded phase field (see additional details on MFP in Section S1 and Figures S1 and S2 of Supporting Information S1). MFP was performed in the 4–6 Hz frequency band, where Rayleigh waves dominate the wavefield. The resulting catalog contains 75,689 events, with a large population of surface sources attributed to surface crevasse, given their spatial distribution mimicking that of surface crevasses (Figures 1b and 1c).

Although detection was performed in the 4–6 Hz band, a subset of these events exhibits clear, higher-frequency, high signal-to-noise ratio (SNR) P-wave arrivals (Figure 1d). The frequency–velocity diagram from Paris, Gimbert, Le Bris, et al. (2026), computed with high-quality surface events (i.e., with high MFP output value) using the MFP algorithm without frequency averaging, shows that P waves propagate with an apparent velocity of $\sim 3,730$ m/s (black box in Figure 1e) and dominate the signal in the 40–80 Hz frequency range. Below ~ 40 Hz, surface waves dominate, while above ~ 80 Hz phases become harder to identify as SNR decreases significantly.

Imaging the basal interface through waveform stacking requires accurately located events with similar source characteristics. Because the uncertainty of the source location increases with depth (Paris, Gimbert, Le Bris, et al., 2026), we restrict our analysis to surface events, retaining only sources with depths shallower than 10 m. In this depth range, the vertical uncertainty (7–16 m) and horizontal uncertainty (1–5 m) (Paris, Gimbert, Le Bris, et al., 2026) are both small relative to the P-wave's wavelength (~ 62 m at 60 Hz), allowing for constructive interference (further discussion on the influence of source location uncertainty in Section S2 of Supporting Information S1). This selection yields a final data set of 30,972 events (dark blue dots in Figure 1c).

2.2. Two-Step Event Synchronization Template Generation

The use of multiple events to image the ice–bed interface requires accurate temporal synchronization so that reflected body waves emerge through waveform stacking. Because MFP is conducted within predefined time windows, it does not yield event origin times. Consequently, waveform alignment cannot rely on absolute timing, and we therefore adopt a two-step synchronization strategy. A schematic overview of the complete synchronization workflow is provided as a flowchart in Figure S4 of Supporting Information S1.

We first use the optimal velocity obtained from MFP to correct for source–receiver travel times across the array. The vertical-component waveforms are then bandpass filtered between 4 and 6 Hz, normalized by their maximum absolute amplitude, and averaged to produce a reference waveform for each event. Among these, the reference waveform with the largest amplitude is selected and hereafter referred to as the uber-reference. We compute the cross-coherence between the uber-reference and each individual reference waveform to estimate the relative time shifts required to synchronize all events in the 4–6 Hz frequency band. This procedure effectively aligns the Rayleigh wave arrivals, leading to their clear emergence in a section stacking all events filtered in the 4–6 Hz band (Figure S5a in Supporting Information S1). However, this synchronization is insufficient to coherently align the higher-frequency P-wave arrivals in the 40–80 Hz band, which are not observed in the stacked waveforms (Figure S5b in Supporting Information S1).

To preserve the P-wave arrivals, we coherently align waveforms in the 40–80 Hz frequency band, where P waves dominate the signal. Starting from the previously synchronized events, we generate new reference waveforms by correcting for P-wave travel times assuming a homogeneous P-wave velocity equal to 3,730 m/s identified from direct arrivals (see previous section and Figures 1d and 1e). The vertical-component waveforms are then bandpass filtered between 40 and 80 Hz using a zero-phase Butterworth filter, normalized by their maximum absolute amplitude, and stacked. We select the reference waveform of an event characterized by a high MFP output value and a clear, high SNR P-wave arrival, designating it as the uber-reference. This uber-reference is trimmed to a 0.26 s window starting 0.13 s prior to the peak amplitude to define a template. We compute the sliding correlation coefficient between this template and each reference waveform to determine the lag that maximizes correlation. To construct a more robust, global template, we stack the reference waveform of events with correlation coefficients exceeding 0.7 to generate an updated template. The sliding correlation procedure is then repeated using this refined template to obtain the final synchronization. Following this second step, direct P-wave arrivals emerge clearly in the 40–80 Hz band (Figure S5d in Supporting Information S1).

2.3. Seismic Reflection Processing

2.3.1. Reflection Location and Topography Correction Estimation

In conventional active seismic reflection workflows, the horizontal location of a reflector is typically approximated as the midpoint between the source and receiver, with topographic effects corrected using elevation differences measured by Real-Time Kinematic (RTK) GPS. These topographic corrections assume sub-vertical ray paths, which, in a homogeneous medium, is valid only when source–receiver offsets are small relative to the reflector depth. At our study site, BedMachine Greenland v5 (Morlighem et al., 2022) indicates an average ice

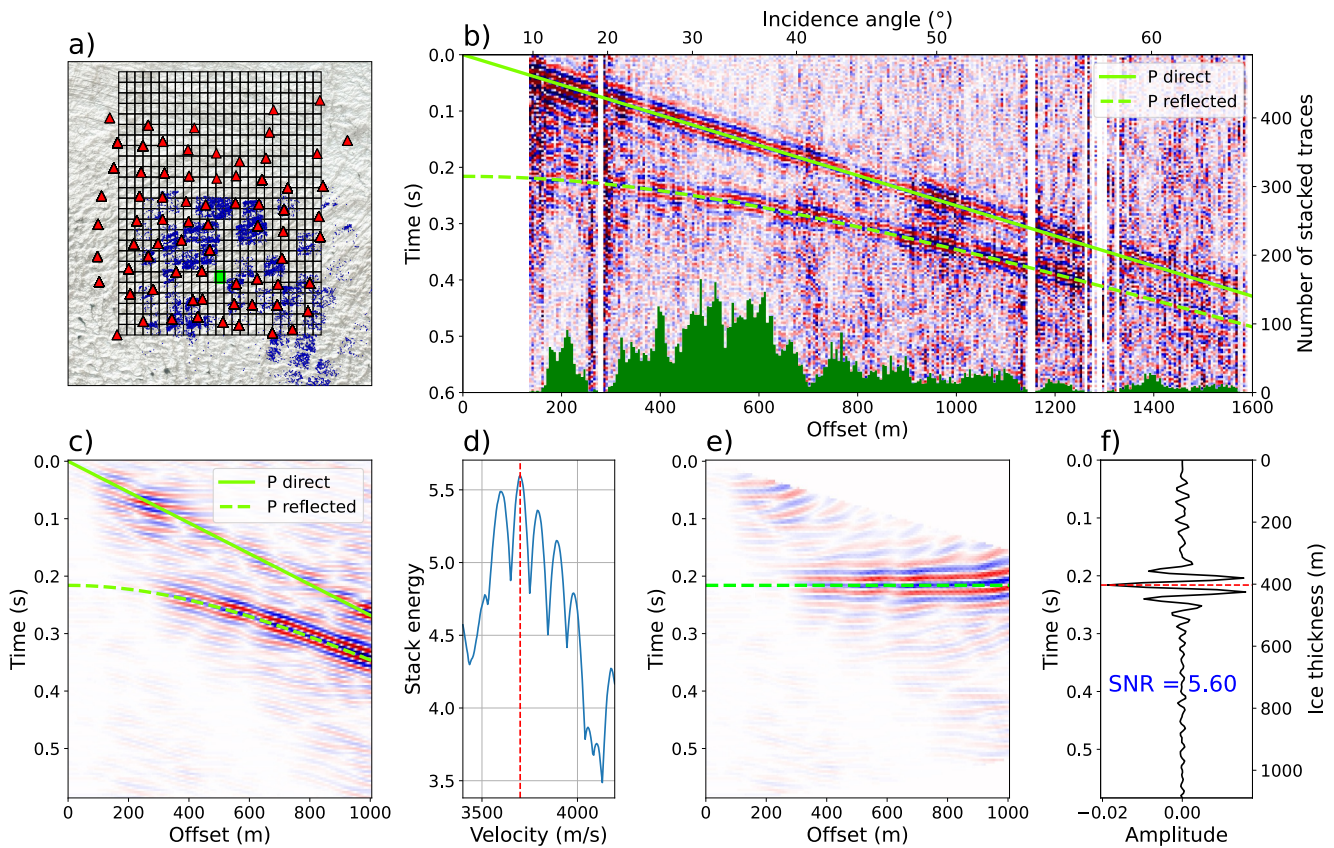


Figure 2. Seismic reflection processing for a cell of the grid. (a) Selected pairs of events (blue dots) and stations (red triangles) whose P waves reflect off the bed in the light green region. (b) CDP gather for reflections in the light green region in (a), filtered between 40 and 80 Hz with a zero-phase Butterworth filter. The light green solid and dashed lines mark the direct and reflected P-wave arrivals, respectively. The histogram (green bars) indicates the number of waveforms stacked in each 5-m offset bin. (c) F-K filtered CDP gather. (d) Velocity analysis result. The vertical red dashed line indicates the optimum NMO velocity. (e) NMO corrected gather. (f) Stacked NMO corrected gather along offset. The horizontal red dashed line indicates the picked two-way travel time.

thickness of ~ 400 m. Given the ~ 50 m east-west surface elevation difference across the array, this relatively shallow depth makes the large-scale surface slope non-negligible comparatively. Because the offset coverage provided by the array aperture can result in large incidence angles, standard topographic correction assumptions may be challenged where the source-receiver paths of the reflected arrivals are aligned with the direction of the surface slope.

To account for surface topography, we apply a correction using as prior information the large-scale ice thickness and bed elevation of BedMachine v5. We invert for the three-dimensional location of the specular reflection point that minimizes the total reflected travel path for each event–station pair. This approach explicitly incorporates elevation differences between source and receiver, as well as the large-scale bed topography from BedMachine. For each event–station pair, we then quantify the topographic correction as the travel-time difference between the topography-corrected and true ray paths, where the topography-corrected path is defined as a straight-ray path between a source and receiver placed at the midpoint elevation.

2.3.2. Common Depth Point Gather Processing and Ice Thickness Picking

We evaluate reflections on a grid of 25×25 cells covering the array. For each cell, we construct a common depth point (CDP) gather by selecting all event–station pairs whose estimated reflection point falls within the cell (Figure 2a) and sorting the corresponding traces by source–receiver offset. Each waveform is filtered between 40 and 80 Hz, time-shifted using the static topographic correction (see Section 2.3.1), and normalized by its maximum absolute amplitude. The traces are then averaged in 5 m offset bins, weighted by the correlation coefficient between the event reference waveform and the template, yielding a raw CDP gather. To suppress residual surface-wave energy and strongly attenuate the direct P-wave arrival, we apply an F-K (frequency-wavenumber)

filter, which separates seismic energy according to its frequency-wavenumber characteristics and suppresses arrivals with apparent velocities lower than 4,750 m/s.

To perform normal moveout (NMO) corrections, we window the CDP gather to keep only the data in a time window of 0–0.588 s and an offset range of 0–1,000 m. The offset range was selected to maximize the SNR of the offset-stacked CDP gather, since far offsets are noisier due to the limited number of waveforms available per bin. The time window was chosen to avoid contamination from later arrivals, which could bias the recovered NMO velocity. We use a range of test velocities between 3,400 and 4,200 m/s to correct for the hyperbolic moveout of the reflected arrival. For each test velocity we compute the energy of the NMO-corrected gather stacked along offset, and use it to define the optimal NMO velocity for NMO correction. Ice thickness is then calculated by picking the highest absolute amplitude peak in the stacked NMO-corrected gather and converting the corresponding two-way travel time to depth with the homogeneous P-wave velocity of 3,730 m/s. Finally, bed topography is obtained by subtracting the estimated ice thickness from the surface elevation, which is approximated by fitting a second-order polynomial to the surface elevations at the nodes.

3. Results

3.1. Example of CDP Gather and Ice-Thickness Estimate

In Figure 2b we show an example of CDP gather obtained from selecting all event–station pairs whose reflection points fall within the light green rectangle shown in Figure 2a. We observe two high-SNR arrivals with similar polarity corresponding to the direct P-wave arrival and the primary P-wave reflection at the ice–bed interface, along with strong surface wave energy at near offsets (<400 m). The offset data coverage is uneven, with no available waveforms from stations located closer than 100 m and limited coverage beyond 700 m (see also the contributing events shown as blue dots in Figure 2a). Despite the limited number of stacked waveforms at large offsets, the reflected arrival maintains a high SNR up to an incidence angle of $\sim 55^\circ$.

After applying the F–K filter, the energy associated with surface waves and the direct P wave is strongly attenuated, while the reflected arrival is preserved (Figure 2c). Velocity analysis of the F–K filtered and normal moveout (NMO)–corrected gather reveals a peak energy at a NMO velocity of 3,700 m/s (Figure 2d). Applying this optimal NMO correction flattens the reflected arrival across most offsets, although the last 50 m of the profile appears slightly overcorrected in travel time as shown by the upward curvature of the primary bed reflection (Figure 2e).

Stacking the filtered and NMO-corrected traces over offset produces a high-SNR zero-phase wavelet with a peak amplitude at approximately 0.21 s, consistent with the expected travel time of a P wave reflected from a depth of ~ 400 m (Figure 2f). The SNR of the picked reflection is quantified as the ratio of the peak absolute amplitude to the standard deviation of the stacked waveform.

3.2. Ice Thickness Map

In Figure 3a we show the ice thickness estimates obtained over all CDP gathers and under topographic correction. These estimates are generally similar to those obtained without topographic correction (see Figure S6 in Supporting Information S1), owing to the two-dimensional geometry of the array, which yields source–receiver pairs spanning all azimuths and therefore includes paths not aligned with the surface slope direction; this indicates the technique remains applicable even where prior knowledge of bed and surface topography is unavailable. We observe that ice thickness varies smoothly across neighboring cells (between 340 and 500 m) in the southern half of the array (Figure 3a), but exhibits strong spatial variability in the northern and particularly northeastern portions of the array. The increased variability occurs in regions with few detected surface seismic events (Figure 1c). This results in CDP gathers containing few waveforms, leading to sparse offset coverage and unreliable ice thickness estimates.

To keep only reliable ice thickness estimates, we first discard cells with optimal NMO velocities outside the physically plausible range of 3,550–3,950 m/s (Figure 3b), defined from the width of the P-wave phase in the frequency–velocity diagram (black box in Figure 1e). We also exclude cells with SNR <4.2 because NMO velocities below this threshold are frequently non-physical (Figure 3c), suggesting that the bed reflections in these cells are not sufficiently coherent to produce reliable thickness estimates. Finally, we remove the remaining isolated cells that lack neighboring estimates, which have not been filtered because their CDP gathers contain spurious arrivals due to limited stacking.

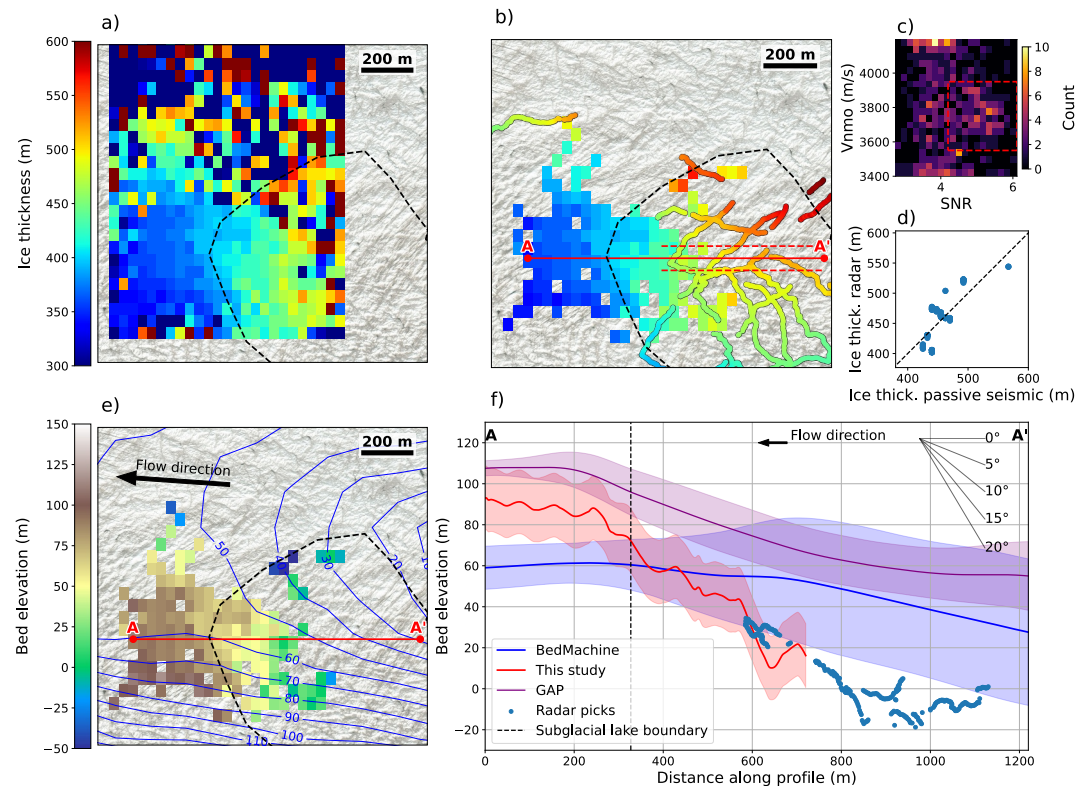


Figure 3. (a) Ice thickness estimates derived from passive seismic reflections. The black dashed outline marks surface elevation Anomaly 2 reported in Livingstone et al. (2019). (b) Same data set as (a) after applying quality filtering based on NMO velocity and signal-to-noise ratio (SNR), with thresholds defined in (c). Colored thick lines indicate ice thickness measured from ground-penetrating radar surveys in Hawkins et al. (2026). (c) Two-dimensional histogram of NMO velocity versus P-wave reflection SNR used to define the filtering criteria applied in (b). (d) Comparison between ice thickness estimates from passive seismic and radar measurements at coincident locations. The black dashed line indicates the 1:1 relationship, where both approaches yield identical values. (e) Bed elevation derived from passive seismic imaging. Blue contours show bed topography from BedMachine v5. The black arrow indicates the glacier flow direction, and the red line marks the location of the profile shown in (f). (f) Bed elevation along the glacier flowline from passive seismic imaging (red), BedMachine v5 (blue), and GAP (purple), with associated uncertainty envelopes (shaded). Radar-derived picks within the region bounded by the red dashed lines in (b), converted to bed elevation using ArcticDEM and projected onto the profile, are shown as blue points. The vertical dashed black line indicates the boundary of surface elevation Anomaly 2.

The remaining ice thickness estimates derived from P-wave reflections are generally similar to those obtained from GPR surveys (Hawkins et al. (2026); colored lines in Figure 3b). Although passive seismic estimates follow the same overall spatial trend as those from radar, they display reduced amplitude variations: where radar-derived thicknesses are minimal, P-wave-derived thicknesses are slightly higher, and where radar thicknesses are maximal, P-wave-derived thicknesses are slightly lower (Figure 3d). We attribute this discrepancy primarily to the lower horizontal resolution of the passive seismic method, which has a footprint governed by the Fresnel zone radius (Chopra et al., 2006), estimated to be on the order of ~ 120 m for a reflector at 450 m depth and a 60 Hz central frequency. Although the radar profiles cover the edge of the region with stable and reliable passive estimates, the P-wave-derived thicknesses generally remain within 20 m of the radar values, comparable to the vertical resolution of $\lambda/4 \approx 15$ m (Chopra et al., 2006), and all estimates fall within 35 m of the radar measurements (Figure 3d).

3.3. Bed Topography Map

We compare the bed topography derived from our passive seismic imaging with that from BedMachine v5 and GAP (Figures 3e and 3f). Our results reveal an east to west increase in bed elevation along the ice flow direction followed by a relatively flat plateau in the western part of the mapped area. Although BedMachine also shows along-flow variations over this region (contour lines in Figure 3e), its dominant gradient is oriented SW–NE,

which differs from our observations. This discrepancy is further illustrated on a W–E along-flow profile (Figure 3f). Along this transect, our bed elevation decreases against the flow direction from ~90 m to ~20 m, with a clear break in slope at ~230 m, ~100 m downstream of the Anomaly 2 boundary, leading to a local slope of ~9°. These estimates are consistent with nearby (<50 m) radar picks, which extend the observed trend farther east, where bed elevation decreases to ~–10 m, defining an overdeepening of ~100 m. In contrast, the BedMachine and GAP profiles differ from each other by up to ~50 m, exhibit limited elevation variations (30–55 m), and show either no clear slope break (BedMachine) or only a subdued one (GAP), corresponding to gentler slopes of less than 5°. This reduced amplitude likely reflects the SW–NE orientation of the slope in these products, which does not align with the along-flow structure resolved by the seismic and radar observations. The seismic observations also reveal small-scale (~100 m wavelength), low-amplitude bed undulations absent from both BedMachine and GAP. Although these could be real, we refrain from interpreting them further since their amplitudes fall below the vertical resolution of the measurements.

4. Discussion

4.1. Toward Higher-Resolution Glacier Bed Imaging in Crevassed Areas

Abundant crevassing-induced seismic sources, combined with dense seismic array deployments, allow imaging of the ice–bed interface in two dimensions without spatial interpolation in a heavily crevassed region inaccessible to more conventional surveying methods, particularly GPR. In contrast to passive seismic techniques based on ambient noise, which primarily exploit surface waves, our method leverages opportunistic sources to recover high-frequency (40–80 Hz) P-wave reflections. This provides more direct constraints on bed depth and improves vertical resolution to 15 m, approaching that of active seismic surveys (Booth et al., 2020).

A key strength of this passive seismic approach is the relationship between source density and seismic array configuration. Because the method relies on stacking reflections from multiple source–receiver pairs, the quality of the reconstructed bed topography map does not only depend on the number of stations, but also on the number of sources. In highly crevassed regions, where seismic sources are abundant, comparable stacking quality may be achieved with fewer deployed sensors. This makes the method particularly resilient in environments where conventional surveys are most challenging. Array design may be optimized beforehand based on crevasse mapping from high-resolution surface elevation data (Chudley et al., 2025) to infer regions with high source density and maximize source–receiver coverage while minimizing the number of required stations. Such a strategy is particularly promising for large-scale deployments on shear margins, grounding zones, surge-type glaciers, and other dynamically active regions, where intense crevassing is associated with strong spatial gradients in ice flow and basal conditions.

In less crevassed areas, where there are fewer sources, such as the northeastern part of our array, retrieving reliable CDP gathers is more challenging. Here, this limitation could be mitigated by expanding the array further north to increase both the number of well-located sources detected with MFP and the offset coverage in the CDP gathers. This dependence on natural source distribution highlights the complementarity of our passive method to conventional active methods (seismic and radar), which target different observational gaps and can be combined to improve overall high-resolution bed topography constraints.

4.2. Bed Overdeepening Co-Located With a Previously Inferred Subglacial Lake

The reconstructed bed topography of the downstream part of Anomaly 2 reveals a pronounced bed overdeepening whose western margin forms a ~100 m relief ridge. The extent of this ridge closely matches that of the anomaly, previously interpreted as an active subglacial lake (Livingstone et al., 2019). This overdeepening is not as well resolved by large-scale bed topography products such as BedMachine and GAP, which underestimate its amplitude and misrepresent its orientation. We suggest the observed bed overdeepening generates strong longitudinal strain rates (Gimbert, Gilbert, et al., 2021; Herzfeld et al., 2012) favoring the formation of crevasses in this area. Consequently, regions most likely to host subglacial water storage could be those where conventional geophysical surveys are most challenged. As a result, sparse observations and interpolation in large-scale models may lead to systematic under-detection and poor characterization of these features.

4.3. Potential for Constraining Basal Interface Properties

Beyond bed topography, seismic reflections also contain information about the physical properties of the basal interface, including contrasts in density and seismic wave velocity. In reflection seismology, these properties are investigated using amplitude-versus-angle (AVA) analysis, which exploits the variation of reflection amplitude with incidence angle (Booth et al., 2012; Castagna & Backus, 1993; Ostrander, 1984). Conducting such an analysis with our passive reflection method requires further developments dealing with non-normalized waveforms, as well as source amplitude and wave attenuation estimates. These improvements are beyond the scope of this study, but could provide complementary constraints to those obtained from active seismic reflection surveys given their broader offset coverage. Incorporating these large-offset measurements would be particularly valuable for discriminating between soft (till) and hard (basement) bed materials, as well as providing evidence for the existence of a water layer (Agnew et al., 2025; Ju et al., 2026; Peters et al., 2008).

In addition, the ice thickness measurements from passive P-wave seismic reflection provide valuable independent constraints for seismic wave velocity models derived from ambient noise surface wave tomography. Incorporating these constraints would reduce the non-uniqueness inherent to 1D dispersion-curve inversion, yielding better-resolved velocity structures, thereby improving our ability to characterize basal material, as well as englacial and subglacial water storage.

5. Conclusion

We establish a new workflow that enables high-resolution mapping of ice thickness in highly crevassed regions with limited field effort. Our method exploits abundant natural seismic sources generated by crevassing events previously detected with matched field processing using a dense seismic node array. This approach enables direct imaging of the bed without the need for spatial interpolation, while achieving a vertical resolution comparable to that of active seismic surveys.

We apply our method to data acquired from a dense seismic node array deployed on Isunnguata Sermia, an outlet glacier in West Greenland where previous studies have suggested the presence of active subglacial lakes. The resulting ice thickness estimates are consistent with independent ground-penetrating radar measurements and extend them downstream into areas where surface crevassing prevented GPR acquisition. The resolved bed shows a pronounced along-flow overdeepening of ~100 m that is poorly resolved in large-scale bed elevation products. The downstream limit of this overdeepening forms a clear ridge whose extent is close to that of Anomaly 2 observed in Livingstone et al. (2019). This supports the capacity of this region to host a subglacial lake.

Our workflow offers strong potential for improved imaging of marginal regions around Greenland and Antarctica (e.g., shear margins, grounding zones, surge-type glaciers), which are challenging to access with conventional surveying methods and can be difficult to resolve using airborne imagery due to high signal attenuation. The energy released by crevassing events is high enough that primary P-wave reflections can be recorded with high SNR across a wide range of incidence angles. These reflections provide an opportunity for AVA analysis to constrain the physical properties of basal materials, which play a significant role in glacier dynamics.

Acronyms

CDP	Common Depth Point
NMO	Normal Moveout
SNR	Signal-to-Noise Ratio
AVA	Amplitude versus Angle
MFP	Matched Field Processing
GPR	Ground Penetrating Radar

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

We used data from the “9F” network (Gimbert & RESIF, 2023) which will be available under the FDSN network code “9F” (10.15778/RESIF.9F2023) by January 2027 at the Epos-France datacenter (Péquegnat et al., 2021) and through the European Integrated Data Archive (EIDA, Strollo et al. (2021)). The event catalog, waveforms, and code to reproduce the results are available in Paris, Gimbert, Roux, et al. (2026).

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