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Ice-core-dated englacial stratigraphy across Wilkes Subglacial Basin, East Antarctica: Implications for ice-sheet behaviour

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Abstract

Wilkes Subglacial Basin (WSB) stores enough ice to raise global sea levels by three to four metres, and comprises a region of East Antarctica potentially vulnerable to climate-induced ice loss. We investigate WSB's glacial history back to just after the Last Interglacial (~130–115 ka) using its englacial stratigraphy. We trace eight radio-echo-sounded internal reflection horizons (IRHs) across WSB, dating them at intersections with Talos Dome and EPICA Dome C ice cores, with one IRH englacially connecting the ice cores. The WSB IRHs span ages from 13.1 ± 0.5 to 90.3 ± 2.3 ka, with an older 122.5 ± 2.0 ka IRH traced across the Dome C plateau. With these IRHs, we make regional inferences about past and present ice behaviour, including identifying areas of IRH drawdown relating to ice velocity, accumulation and geothermal heat flow. Approximately 300 km upstream of the WSB grounding line, 20% of the ice lies below and hence is older than 90.3 ± 2.3 ka, suggesting ice presence across this region throughout the Last Interglacial. Several of our IRHs have been mapped extensively elsewhere across the East Antarctic Ice Sheet, and this study therefore contributes to the wider objective of mapping Antarctica-wide englacial stratigraphy.

Introduction

The East Antarctic Ice Sheet (EAIS) contains a volume of ice equivalent to a global sea-level rise of 52 m (Morlighem and others, 2020). There has been extensive ice cover over the East Antarctic continent since about 34 Ma (Colleoni and others, 2022) and, in the EAIS interior, palaeoclimate records have been cored to an age of ~800 ka (Wolff and others, 2010), and may reach > 1.5 Ma (Chung and others, 2023). Compared to its more dynamic neighbour, the West Antarctic Ice Sheet (Sugden, 1996; Gardner and others, 2018), which has experienced larger variability and sensitivity to climate change over time and is currently experiencing significantly faster flow and ice mass loss, it is often assumed that the EAIS was less responsive to Quaternary climate cycles, indicating a potentially higher resilience to anthropogenic warming. However, over the past few years, evidence has been accumulating to suggest that parts of the EAIS may be more strongly connected to climate, and hence more vulnerable to deglaciation, than previously assumed (Stokes and others, 2022; Patterson and others, 2026). A major focus has been on regions of the EAIS now known to occupy large marine basins, which may be vulnerable to marine ice-sheet instability (Mercer, 1978; Joughin and others, 2014; DeConto and Pollard, 2016; Scambos and others, 2017).

Wilkes Subglacial Basin (WSB) is one of several marine basins in East Antarctica, and stores enough ice to raise global sea levels by 3–4 m (Morlighem and others, 2020). Satellite observations over the last few decades have shown a number of WSB outlet glaciers accelerating and thinning (Rignot and others, 2019), while several recent studies based on ice-sheet modelling have shown that the ice sheet in EAIS marine basins may be subject to substantial retreat and sea-level contribution on millennial time-scales, under higher future anthropogenic climate change scenarios (Golledge and others, 2015; DeConto and Pollard, 2016; Stokes and others, 2022). Furthermore, a growing body of research underpinned by marine-sediment and ice-core records suggests variable extents of deglaciation or grounding line retreat in WSB during past warm interglacials (Cook and others, 2013; Mengel and Levermann, 2014; Bertram and others, 2018; Wilson and others, 2018; Blackburn and others, 2020; Sutter and others, 2020;

2021; Stokes and others, 2022; Iizuka and others, 2023). This includes evidence of a dynamic, more climate-sensitive, EAIS in WSB during several warm periods in the Miocene (23–5.3 Ma), Pliocene (5.3–2.6 Ma) and Pleistocene (2.6 Ma–11.7 ka), with more recent retreat of WSB indicated during the Last Interglacial Period (LIG), 130–115 ka (Cook and others, 2013; Sutter and others, 2020; Crotti and others, 2022; Iizuka and others, 2023).

In this paper, we further expand the englacial exploration of the Antarctic Ice Sheet to a new region, and simultaneously provide a new perspective on understanding ice-sheet evolution in WSB by exploring the englacial stratigraphy that spans the upstream reaches (~ 600–300 km from the grounding line) of WSB. We trace internal reflection horizons (IRHs) imaged through the ice in airborne radio-echo sounding (RES) data collected across WSB. IRHs are caused by dielectric contrasts in the ice from changes in density, conductivity or crystal orientation (Robin and others, 1969; Bingham and others, 2025). In this study, where we trace IRHs well below the ice surface and well above the ice base, we attribute the IRHs to variations in ice conductivity, stemming from atmospheric variability affecting the acidity of the snow that initially accumulated on the ice sheet (Siegert, 1999). The IRHs can therefore be treated as isochrones, providing spatiotemporal records of age–depth structure that complement the point-based measurements of ice cores (Bingham and others, 2025). In tracing IRHs across WSB, we generate a new and valuable dataset of IRHs and englacial stratigraphy across WSB and the Dome C Plateau, which can act as useful observational anchors for future three-dimensional ice-sheet modelling studies investigating past and future evolution of the ice sheet in this region. With one of our IRHs, we make an important englacial connection between EPICA Dome C and Talos Dome ice cores, which extends further across EAIS to the pan-continental ice-sheet radiostratigraphy that is being compiled by the Scientific Committee for Antarctic Research (SCAR) AntArchitecture initiative (Bingham and others, 2025). Furthermore, we are able to extract information on IRH continuity, depth and deformation, from which we make interpretations about the internal structure and flow of the EAIS in WSB in relation to bed topography, ice velocity and geothermal heat flow (GHF). These insights, in turn, facilitate future studies to combine observational and modelling methods to investigate evolution and stability of the EAIS in WSB. Finally, in tracing and dating these IRHs, we reflect on lessons learned for extending this initiative over other regions with potentially more disrupted englacial stratigraphy.

Methodology

Radio-echo sounding data

We use airborne RES data acquired across WSB during the 2005/06 austral summer season jointly by the British Antarctic Survey (BAS) and the Italian Programma Nazionale di Ricerche in Antartide (PNRA) (Corr and others, 2007; Ferraccioli and others, 2021). The dataset contains over 61,000 km of flight line data. The densest coverage was centred through WSB between latitudes 68 and 78 °S, but with profiles extending westwards to Dome C and eastwards over Talos Dome in North Victoria Land (Fig. 1). The data were acquired with the BAS Polarimetric Airborne Survey INstrument (PASIN) (version 1), a bistatic radar that operated at 150 MHz centre frequency with 10 MHz bandwidth. PASIN acquired data with a shallow-sounding pulse

(0.1 microsecond) with a range resolution of 8.4 m, interleaved with a deep-sounding linear chirp (4 microsecond, 10 MHz), with a range resolution of 12.9 m (Ferraccioli and others, 2021). The data were processed through chirp decompression, incoherent stacking and 2-D Synthetic Aperture Radar (SAR) processing to improve along-track resolution and signal-to-noise ratio (Ferraccioli and others, 2021).

Tracing internal reflection horizons

To optimise the RES data for tracing IRHs, we first processed them to improve IRH traceability. This comprised removal of the aircraft-to-surface gap, using the surface pick from the aircraft radar altimeter on the radar data, and the application of a bandpass filter and gain function that significantly enhances the layering, as previously employed on PASIN RES data (Ashmore and others, 2020; Bodart and others, 2021). We then loaded the data, in SEG-Y format, together with previously traced bedpicks (Ferraccioli and others, 2018), into the seismic-data software package Schlumberger Petrel (hereafter: Petrel), which we used to trace IRHs using a combination of semi-automatic and manual picking tools (Fig. 2). The semi-automatic picker used a seeded 2-D auto-algorithm; a point on the radargram was manually selected, and the algorithm followed this pattern of reflectance for as long as it could. Depending on the consistency and visibility of the IRHs, the semi-automatic picker could follow them from between hundreds of kilometres (typically in areas of slow-flowing, stable ice) to just a few metres (typically where IRH geometry was complicated by ice dynamics, which we discuss in the Results section). When consistency and visibility of englacial layers became too low, and the semi-automatic picking failed, but IRHs were still visible enough to trace, we applied manual picking tools to fill gaps and extend IRHs further. We did not ‘pick’ IRHs across gaps or regions where they could not be seen in the RES data. This meant that, due to limitations in visibility and consistency of IRHs over certain areas and in some radar lines, different IRHs were traced over various distances, and no IRH could be traced over the entire dataset. In total, we traced various IRHs over approximately 70% of the dataset. Throughout this procedure, and following Ashmore and others (2020), we alternated between tracing in pulse and chirp radargrams, depending on the reflectivity of IRHs. As the pulse data better display IRHs in the upper sections of the ice column, and chirp data better display IRHs in the deeper sections of the ice column (Frémand and others, 2022), switching between the two allowed us to trace more effectively throughout the entire ice column.

Eight individual IRHs were traced, all connecting to one or both of Talos Dome and Dome C. This facilitated age–depth assignment with both the Talos Dome Ice Core (TALDICE) (Urbini and others, 2008; Bouchet and others, 2023) and EPICA Dome C Ice Core (EDC) (Wolff and others, 2010; Bouchet and others, 2023). We selected these IRHs on the basis that each of them is expressed with a relatively bright reflectivity where they intersect with either TALDICE or EDC. Overall, the IRHs span ~ 30–95% and 10–50% of the ice column at intersection with each of the ice cores, respectively, hence giving a structural overview through most of the ice depth. In addition, two of the IRHs were selected as they appeared to be equivalent to bright reflectors that Winter and others (2019) traced to other locations across the EAIS.

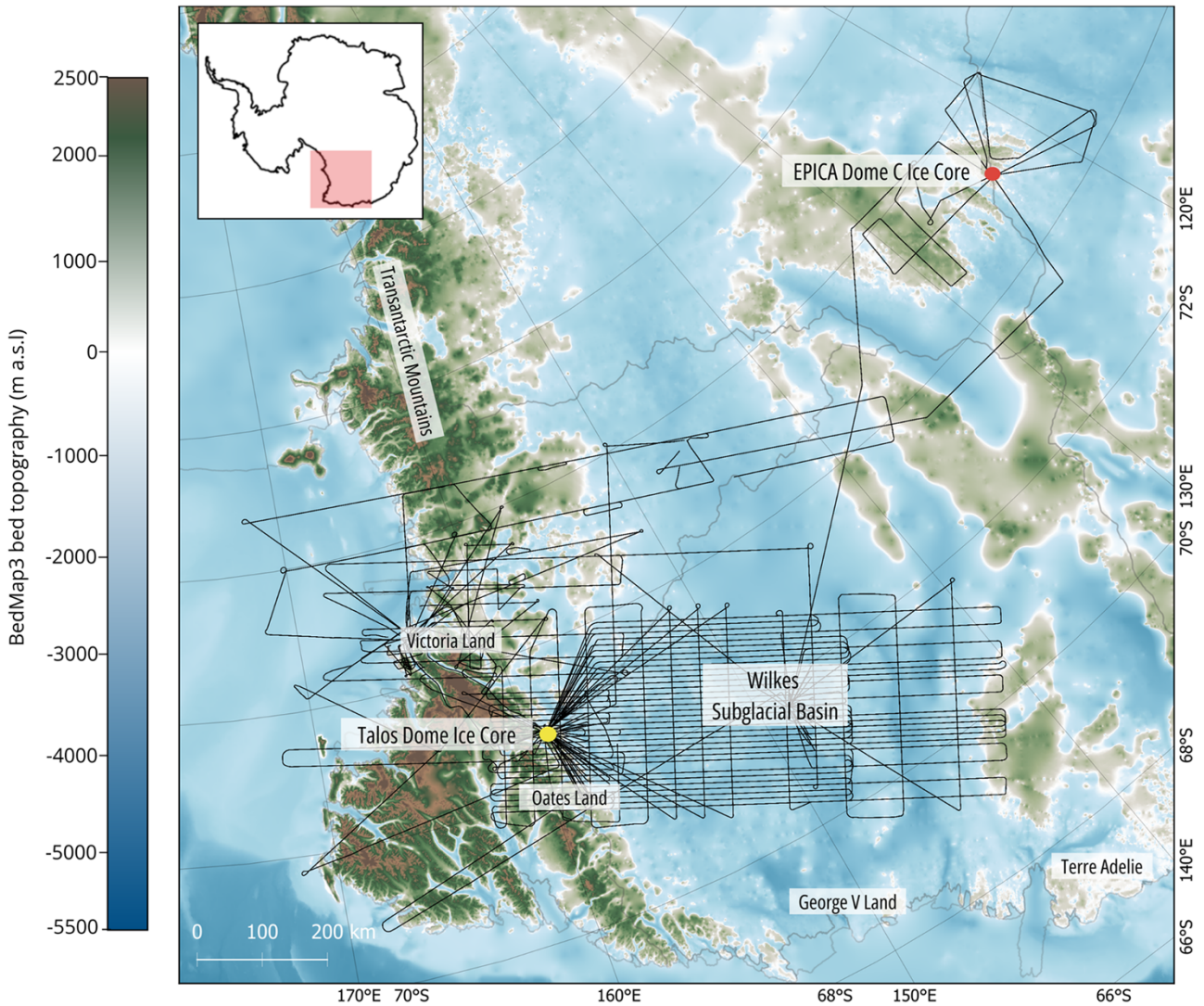


Figure 1. Dataset coverage over Wilkes Subglacial Basin. Location within Antarctica shown in inset, top left. Prominent black lines depict RES tracks acquired by BAS/PNRA in 2005/06, overlain on Bedmap3 bed topography (Pritchard and others, 2025), with drainage basin outlines in grey from Zwally and others (2012).

IRH conversion to depth and dating

We initially converted the two-way travel time of the radargrams into depth d (m), using:

$$d = \frac{v_{ice} * t}{2} + z_f$$

where v_{ice} is assumed to be a constant electromagnetic wave propagation velocity of $168.5 \text{ m } \mu\text{s}^{-1}$ (Fujita and others, 2000), t is the two-way travel time and z_f is the firn correction. We used a firn correction of $9.5 \pm 1.7 \text{ m}$ for IRHs dated at intersections with TALDICE, after Bodart and others (2026), and a firn correction of $14.6 \pm 1.4 \text{ m}$ for IRHs dated at intersections with EDC, after Cavitte and others (2016). We then assigned ages to our traced IRHs by linking them to the most recently updated available ice-core chronologies for TALDICE and EDC (Bouchet and others, 2023). In order to assign ages to our traced IRHs, we used the average depth of all points picked within a set radius of the point of intersection between IRH and ice-core location (Fig. 3). For IRHs intersecting TALDICE, we took the average depth for

all the picked points of the IRH in a radius of 50 m around the ice-core location. These picked points came from flightlines 17 and 18. For EDC, because the ice core does not precisely intersect any RES flightlines, we took the average depth for all IRH points picked within a radius of 750 m around EDC to obtain an average IRH depth. These picked points came from flightlines 36 and 37.

We calculated depth uncertainties of our IRHs by accounting for error sources derived from three key factors: (1) variability in the propagation of electromagnetic waves through ice, (2) the firn correction and (3) the range precision of the radar (Cavitte and others, 2016). Variation in the propagation of electromagnetic waves derives from impurities in the ice, temperature changes and anisotropy, which vary through the ice column (Fujita and others, 2000). Here, we collectively estimated these by assuming an error in the electromagnetic wave propagation through ice of $\pm 2 \text{ m } \mu\text{s}^{-1}$ (after Fujita and others, 2000). For firn, we addressed its depth variability across WSB and the Dome C region by ascribing two separate firn corrections to the different ice-core locations, ascribing an uncertainty of $\pm 1.7 \text{ m}$ to TALDICE and an uncertainty

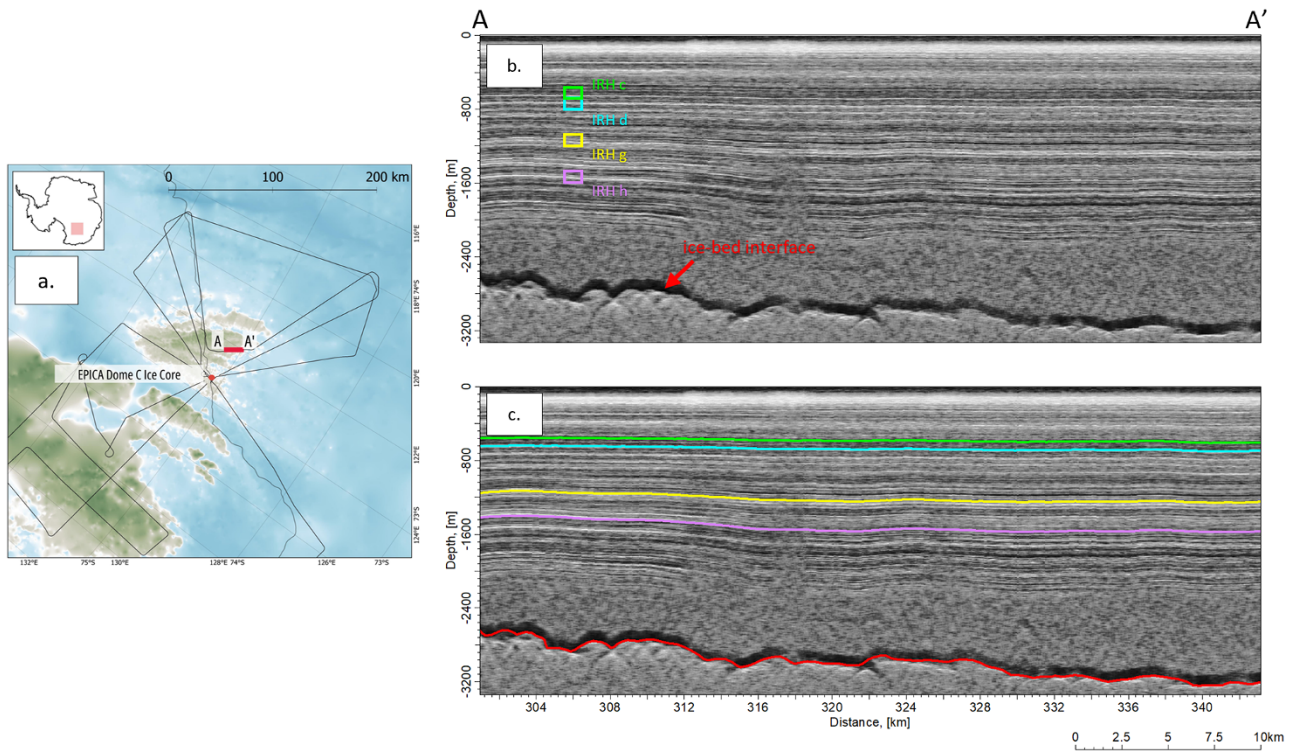


Figure 2. Example of radargram and IRH traceability. (a) Location of example radargram A-A'. (b) Radargram A-A' with IRH-tracing seed points annotated. (c) Radargram A-A' with traced IRHs superimposed.

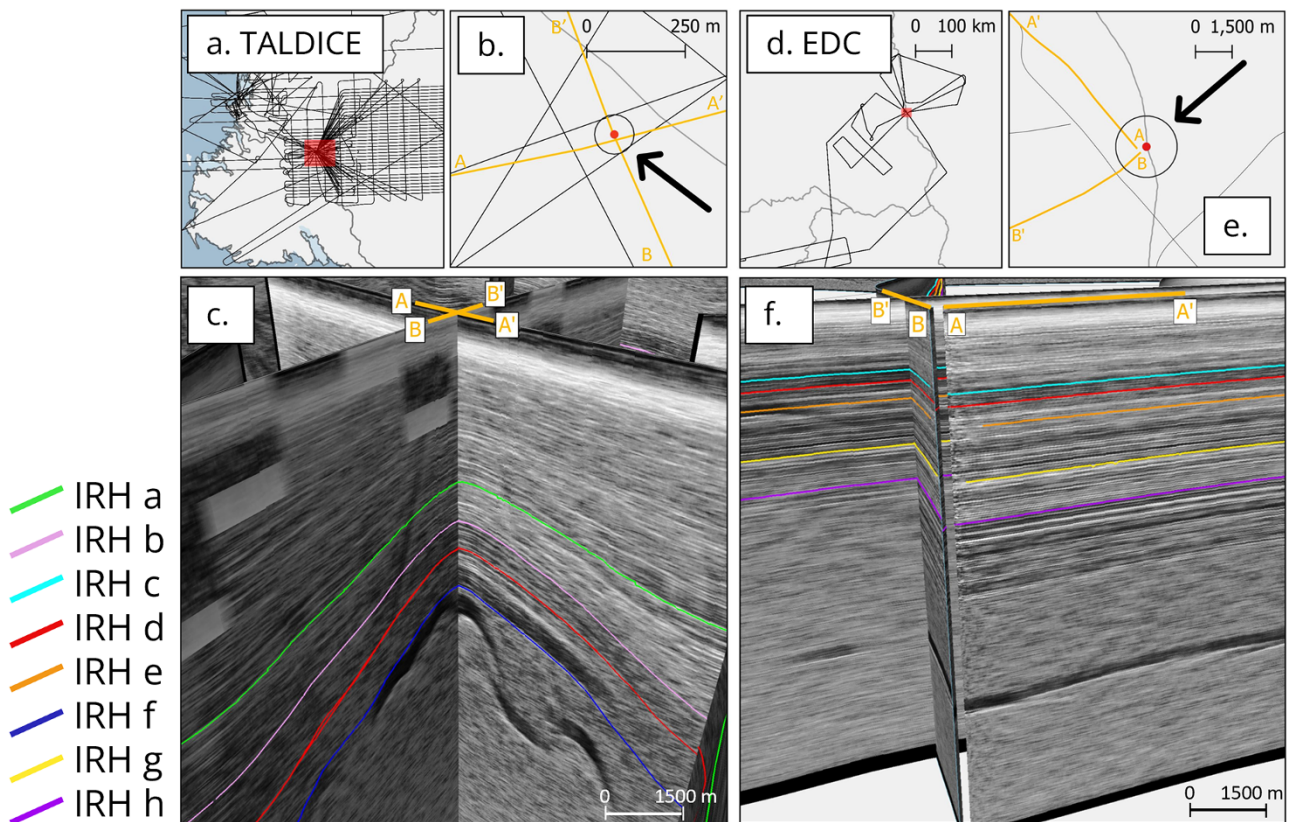


Figure 3. Flight line intersection with ice core locations and method of date assignment for all IRHs. Panels (a)–(c) show flightline intersections at TALDICE, and panels (d)–(f) show flightline intersections at EDC. Panels (a) and (d) show the overview of the area. Panels (b) and (e) show the area within the red box from panels (a) and (d), respectively, with the black arrow highlighting the viewing direction in panels (c) and (f), respectively.

Table 1. Age and depth range for all eight IRHs traced and discussed in this study, with the uncertainty given at the intersection with an ice core. For IRH d that intercepts both TALDICE and EDC, ages were derived from each ice core, as displayed in rows 4 and 5, but for further analysis we use an average age and uncertainty, as provided in row 6. For the fractional depth, the range given is the overall range in fractional depth for the entire area of the traced IRH across the dataset.

IRH	Age (ka)	Uncertainty (ka)	Depth range below surface (m)	Uncertainty range (m)	Fractional depth range	Dating location
a	13.1	0.5	542.8–1326.6	14.5–20.4	0.29–0.53	TALDICE
b	26.6	1.7	749.0–1721.3	15.8–24.2	0.33–0.75	TALDICE
c	30.0	1.4	240.7–1029.8	13.3–17.9	0.10–0.38	EDC
d	39.5	1.3				TALDICE
d	38.0	1.3				EDC
d	38.7	2.6	320.2–2014.2	13.5–27.2	0.13–0.82	TALDICE, EDC
e	51.7	1.5	568.2–1447.5	14.7–21.6	0.17–0.50	EDC
f	63.2	4.8	1258.5–2595.4	19.8–33.4	0.53–0.95	TALDICE
g	90.3	2.3	965.8–2679.1	17.3–34.4	0.30–0.85	EDC
h	122.5	2.0	969.5–2871.0	17.4–36.5	0.38–0.75	EDC

of ± 1.4 to EDC, after Bodart and others (2026) and Cavitte and others (2016), respectively. Lastly, the range precision of the radar affects the vertical resolution of the radargrams and, for PASIN, this is given as 8.4 m for pulse-radar and 12.9 m for chirp-radar (Ferraccioli and others, 2021). As our IRH picking has varied between pulse and chirp radar, we chose here to use the larger value for our uncertainty calculations. We thus calculated depth uncertainties for each IRH pick by summing these uncertainty contributions through quadrature.

To obtain age uncertainties for our IRHs, we took the root-mean-square error (RMSE) of (1) uncertainty in the dating of the ice core, at the depth of mean IRH intersection with the ice core, and (2) uncertainty given by the range in ages from the uncertainty in IRH depth, following Bodart and others (2026). Additional confidence in our derived ages and uncertainties comes from comparison with intersecting traced and dated IRHs from Bodart and others (2026), which we consider in this paper's discussion section.

Results

Extent and geometry of IRHs

Table 1 presents the depths and ages of the eight IRHs we traced across WSB, and their respective uncertainties. We have labelled the IRHs a–h from shallowest to deepest in the ice column. The spatial extent to which each IRH could be traced varied considerably, with IRHs d and g proving to be particularly ubiquitous (Fig. 4). The other IRHs were typically restricted to regions around either Dome C or Talos Dome.

Figures 4 and 5, respectively, map absolute depth and fractional depth (IRH depth relative to ice thickness) of the traced IRHs across WSB. In total, IRH depth varies between 542 and 2871 m below the ice surface, and fractional depth varies between 0.10 and 0.95 across the entire dataset. All IRHs traced around both TALDICE and EDC, and across WSB, generally transition from shallower absolute and fractional depths in the EDC area to deeper and much deeper fractionally in western WSB and over TALDICE. There is no relationship between the spatial traceability of IRHs and their depth (Fig. 4). Most of the limitations with traceability of IRHs traced from TALDICE occur heading into the centre of upper WSB (Fig. 4a, b, f). IRH tracing out of the vicinity of EDC was often compromised when crossing from the high topography of the Dome C Plateau into the deeper topography of WSB (Fig. 4c, e, h). Figure 6 provides a few examples of decreasing traceability in relation to changing bed topography, where particularly bright IRHs are traceable throughout some of these areas, but not throughout

other areas. We explore potential reasons behind these traceability issues in the Discussion. These issues could be combated while tracing within WSB for some of the IRHs due to the high coverage of radar lines. However, only two radar lines linked the WSB area with EDC, which meant that possibilities to trace IRHs between WSB and the Dome C Plateau were limited.

Age of IRHs and dating uncertainties

We dated our traced IRHs to between 13.1 ± 0.5 ka and 122.5 ± 2.0 ka, thus spanning a period from approximately the middle of the LIG until just before the start of the Holocene (Table 1, Fig. 4, 5). The uncertainty in dating of the IRHs varies between 0.5 ka, for IRH a, and 4.8 ka, for IRH f. IRH a is the shallowest and youngest IRH, and its small age uncertainty can likely be attributed to this. IRH f, despite not being the oldest IRH dated in this study, has the largest uncertainty. This could be due to a combination of several reasons. Differences in ice-core resolution, both at different depths of the same ice core, or between two separate ice cores, could affect the uncertainty. Furthermore, IRH f is derived at the greatest fractional depth of ice at intersection with an ice core (at 79% of ice depth at TALDICE; Fig. 5), which also affects the uncertainty. The older IRHs g and h, respectively, have medium uncertainties of 2.3 and 2.0 ka because they are derived from the highest fractional depths at EDC in this study (39% and 48% of ice depth; Fig. 5).

Six of our traced IRHs (a, d, e, f, g, h) correspond in depth, age (see Table 2) and uncertainty range with IRHs traced in WSB and to other parts of the EAIS by others as part of the wider SCAR AntArchitecture initiative (Table 2). This allows us to reconcile these IRHs with the wider englacial stratigraphy and use some of these complementary data to inform our data interpretation, as we expand upon in our later Discussion.

The spatial pattern of IRHs across upper WSB

Using Fig. 7, which focuses on the 90.3 ± 2.3 ka IRH, one of our deepest and most extensively continuous IRHs, we now begin to explore the spatial pattern of IRHs across selected zones of interest in WSB. The figure displays the extent and fractional depth of the 90.3 ± 2.3 ka isochrone across WSB, combining the results from this study with additional tracing of this IRH conducted by Bodart and others (2026). Zones A and B represent RES profiles through which the 90.3 ± 2.3 ka IRH has been traced into the onset region of two outlet glaciers, David and Cook Glaciers. Both cases show reducing amount of ice below the IRH and increasing fractional

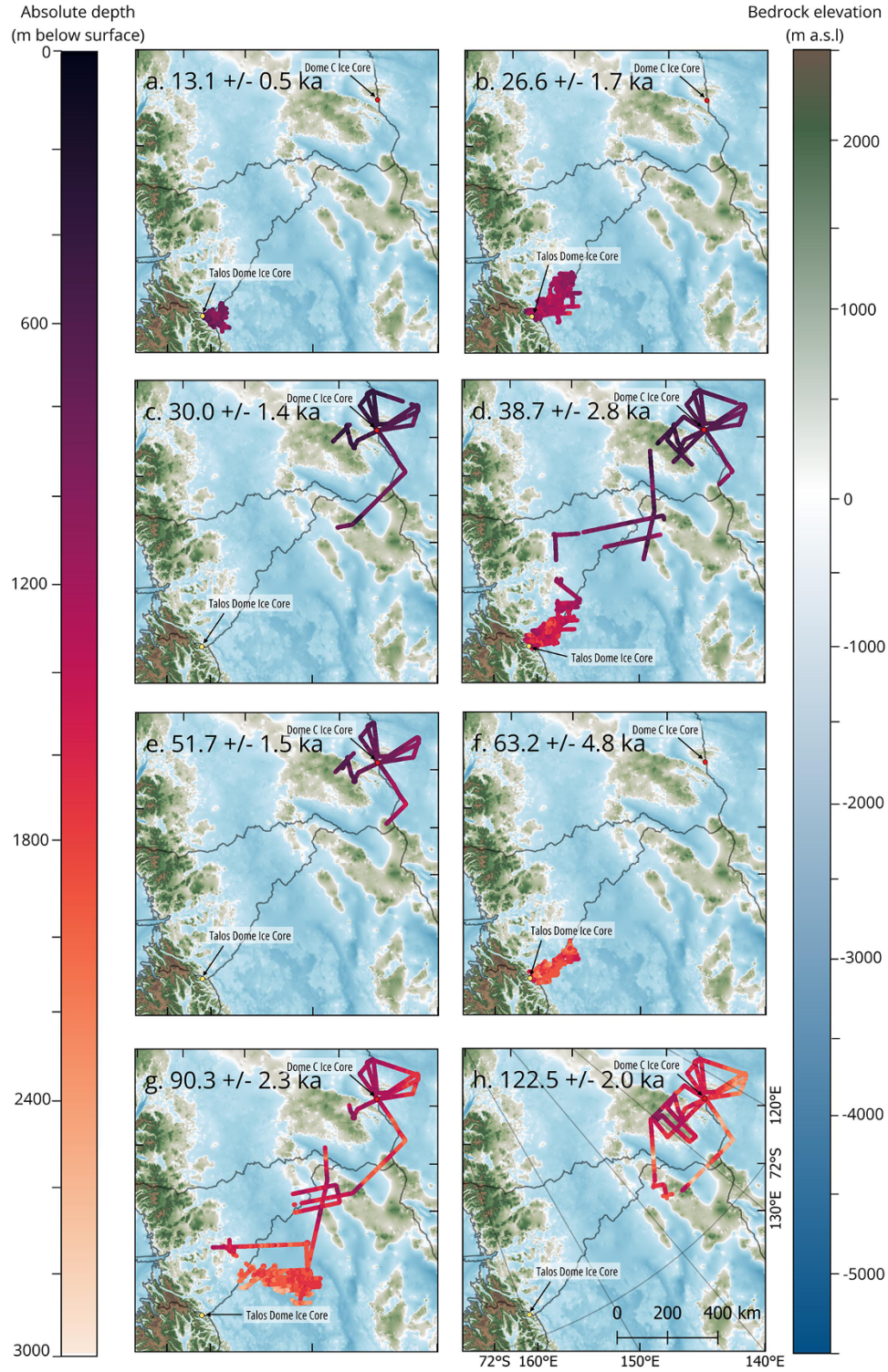


Figure 4. Depth of traced IRHs displayed spatially and overlain over Bedmap3 bed topography (Pritchard and others, 2025), with drainage basin outlines from Zwally and others (2012). The figure displays where an IRH of a certain age has been traced. Spatial variability in depth given by the colour of the line, and the exact ranges in depth can be found in Table 1.

depth as ice flows into the main glacier trunks, where ice velocity is also increasing. Zone C highlights an area of central upper WSB where the IRHs, relative to surrounding regions, remain relatively shallow in the ice and hence there is proportionally a larger volume of ice older than 90.3 ± 2.3 ka lying below the deepest dated

IRH (Fig. 7). This zone is located where the bed is deep and ice flows slowly, and the shallow fractional depth of the IRH may indicate relatively little basal melt, or long-term lower accumulation rates. Zone D marks an area of reducing ice depth below the 90.3 ± 2.3 ka IRH, without a coupled decrease in fractional depth, as the IRH is

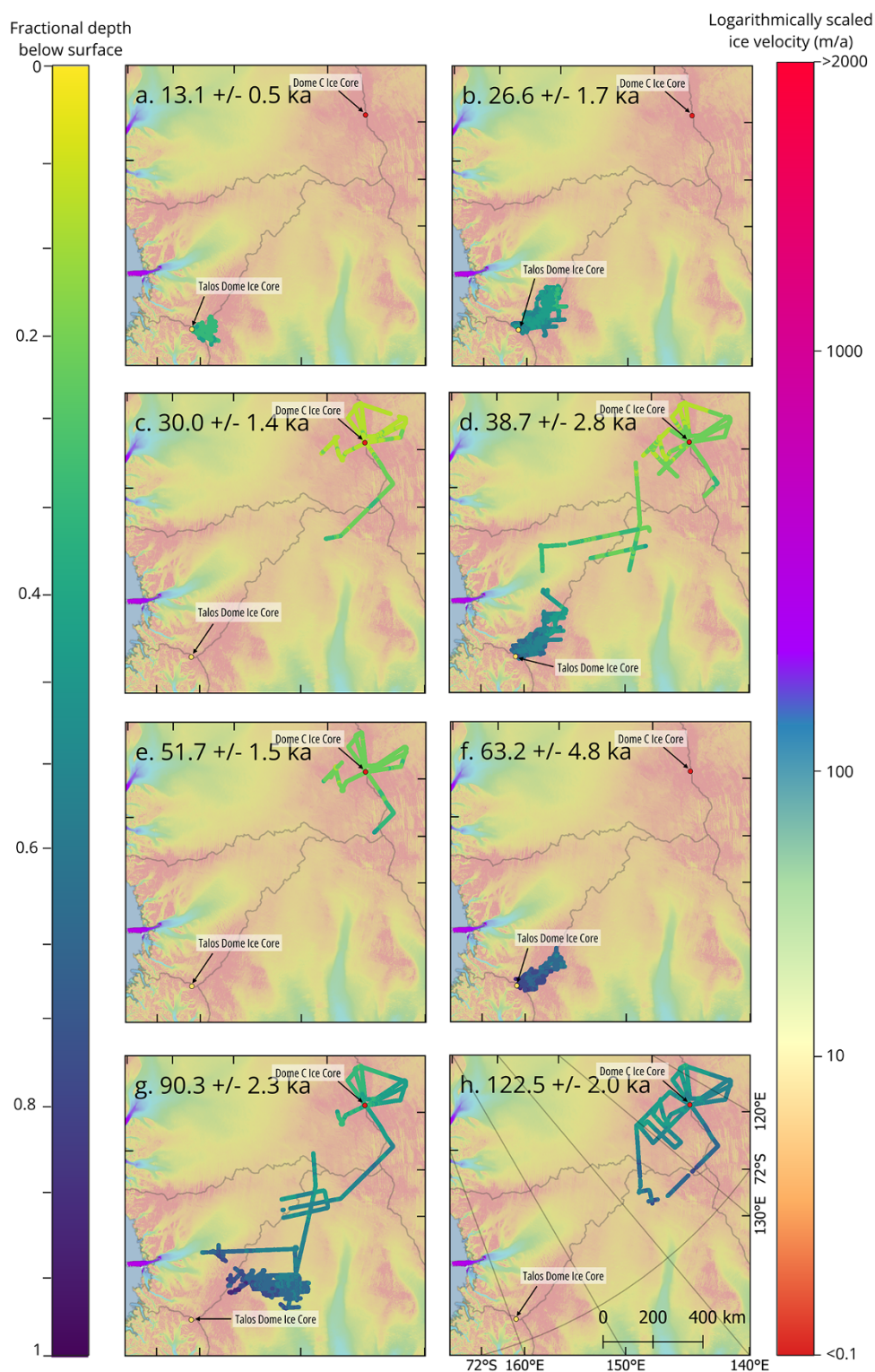


Figure 5. Fractional depth of traced IRHs displayed spatially and overlain over ice sheet velocity from Rignot and others (2017), with drainage basin outlines from Zwally and others (2012). The figure displays where an IRH of a certain age has been traced. Spatial variability in fractional depth given by the colour of the line, and the exact ranges in fractional depth can be found in Table 1. Across the entire area and the entire dataset, the traced IRHs cover over 85% of the ice column in fractional depth.

traced over higher bed topography. Finally, Zone E marks a region where IRHs dip sharply towards the ice-sheet base, with little ice remaining below the 90.3 ± 2.3 ka IRH. Cumulatively, these zones show how the spatial arrangement of the englacial stratigraphy can be used to pinpoint areas of deep, and thus older, ice. Such regions

may correspond to locations where ice dynamical processes, such as ice velocity, mass loss, basal melting and sensitivity to climate-induced perturbations, have a reduced influence. In the following section, we discuss our interpretation of these zones of interest across the upper WSB.

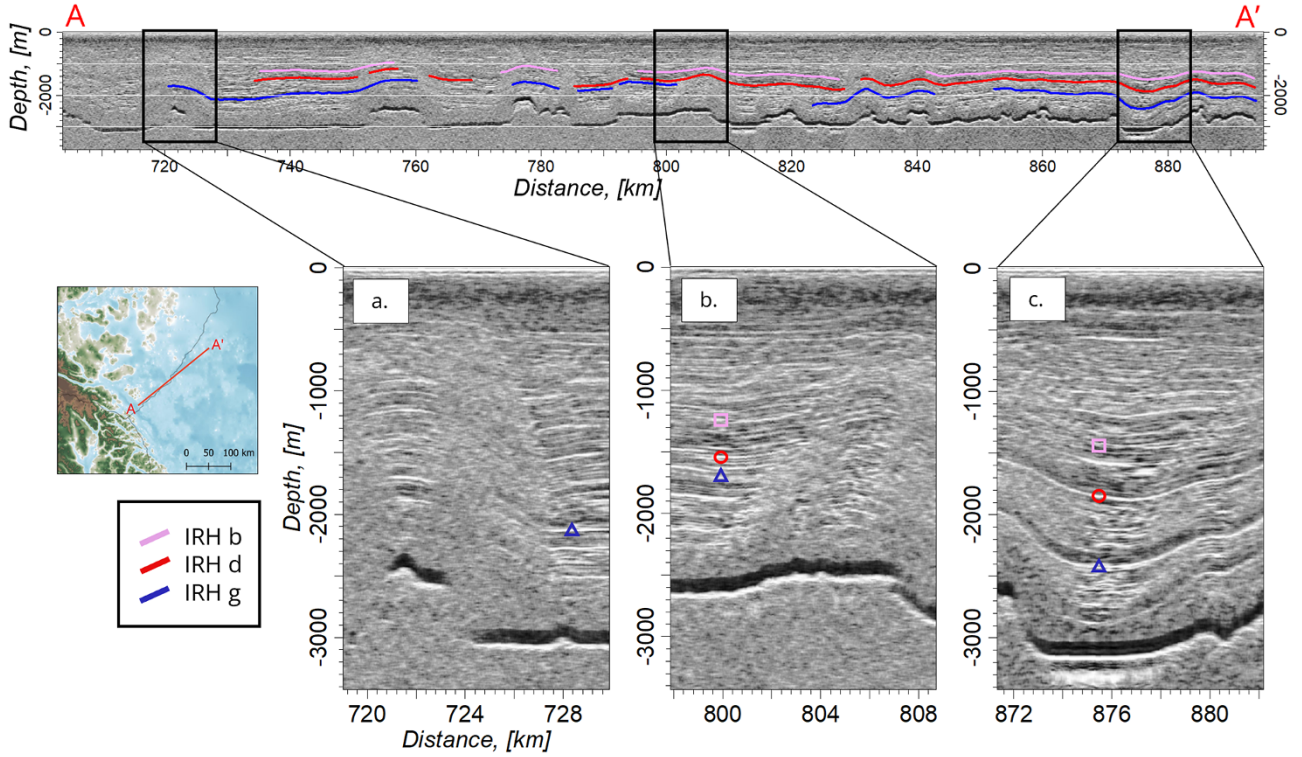


Figure 6. Variations in traceability of three IRHs over a radargram crossing WSB and paralleling the ice divide leading down towards Talos Dome (A' area) as shown in the inset map. Insets show (a) disappearing traceability for all three IRHs, (b) IRH b and d continuously traceable with IRH g disappearing and (c) continuous traceability for all three IRHs.

Table 2. IRH dates from this study, and which other studies have IRH dates which fall within the uncertainty bounds of these. For Winter et al. (2019), there are two separate ages for each IRH, as these were traced in two separate areas: around EDC and around EPICA Dronning Maud Land.

This study	Bodart et al., 2026	Franke et al., 2025	Sanderson et al., 2024	Chung et al., 2023	Wang et al., 2023	Cavitte et al., 2021	Beem et al., 2021	Winter et al., 2019
13.1 ± 0.5 ka	13.2 ± 0.4 ka							
38.7 ± 2.8 ka	38.0 ± 0.8 ka	38.1 ± 0.02 ka	38.2 ± 2.2 ka	39.2 ± 1.1 ka	36.3 ± 3.6 ka	38.1 ± 0.7 ka	37.6 ka	38.2 ± 0.6/± 1.3 ka
51.7 ± 1.5 ka		52.2 ± 0.07 ka					51.4 ka	
63.2 ± 4.8 ka	64.4 ± 3.0 ka			62.2 ± 2.4 ka				
90.3 ± 2.3 ka	90.2 ± 1.9 ka	91.0 ± 1.1 ka	90.2 ka ± 3.6	97.7 ± 2.1 ka	95.5 ± 7.5 ka	96.5 ± 1.8 ka	93.9 ka	90.2 ± 1.6/90.4 ± 1.7 ka
122.5 ± 2.0 ka	121.1 ± 1.8 ka					121.1 ± 1.8 ka		

Discussion

Overall, we have traced and dated eight IRHs across the TALDICE-WSB-EDC region of the EAIS, spanning from the middle of the LIG (122.5 ± 2.0 ka) to just before the onset of the Holocene (13.1 ± 0.5 ka). As a standalone dataset, these IRHs can provide important three-dimensional observational input to modelling studies investigating ice-sheet evolution in WSB. However, they also provide insights into past and current ice-sheet behaviour in the region, as this section discusses. Here, firstly, we use the IRHs to explore how ice flow and deformation, and other geophysical influences alter the position of isochrones in the ice in upper WSB, with respect to ice flow, past accumulation rates and relationship to basal topography and geothermal heat flow. We then consider wider stratigraphic connections between this region and the wider EAIS, using the 38.7 ± 2.8 ka IRH as our exemplar, as this IRH was used to successfully englacially connect EDC with TALDICE in this study, and has been widely traced in other studies elsewhere across the EAIS (see Fig. 9). Finally, we make some suggestions for future practice in tracing and dating englacial stratigraphy where

direct tracing can be compromised, as we have experienced here, by ice dynamics or other causes of englacial disruption.

Spatial patterns of change highlighted by the englacial stratigraphy

The drawdown of englacial stratigraphy at the onset regions of David and Cook Glacier (Zones A and B in Fig. 7, and fractional depth changes in Fig. 8a) matches drawdown patterns of IRHs into outlet glaciers and ice streams as observed elsewhere around Antarctica (Bingham and others, 2025). For Zone A, previous ice-surface reconstructions have revealed mid-Holocene thinning of a large part of the drainage catchment that flows into David Glacier, which continued for about 2 ka (Stutz and others, 2021). For Zone B, this drawdown may be linked to either or both the contemporary ice-dynamic regime and inland thinning, which has been invoked to explain recent increases in velocity observed where the glacier flows into the sea (Miles and others, 2018). Changes in the ice-dynamic regime, such as velocity increase in outlet

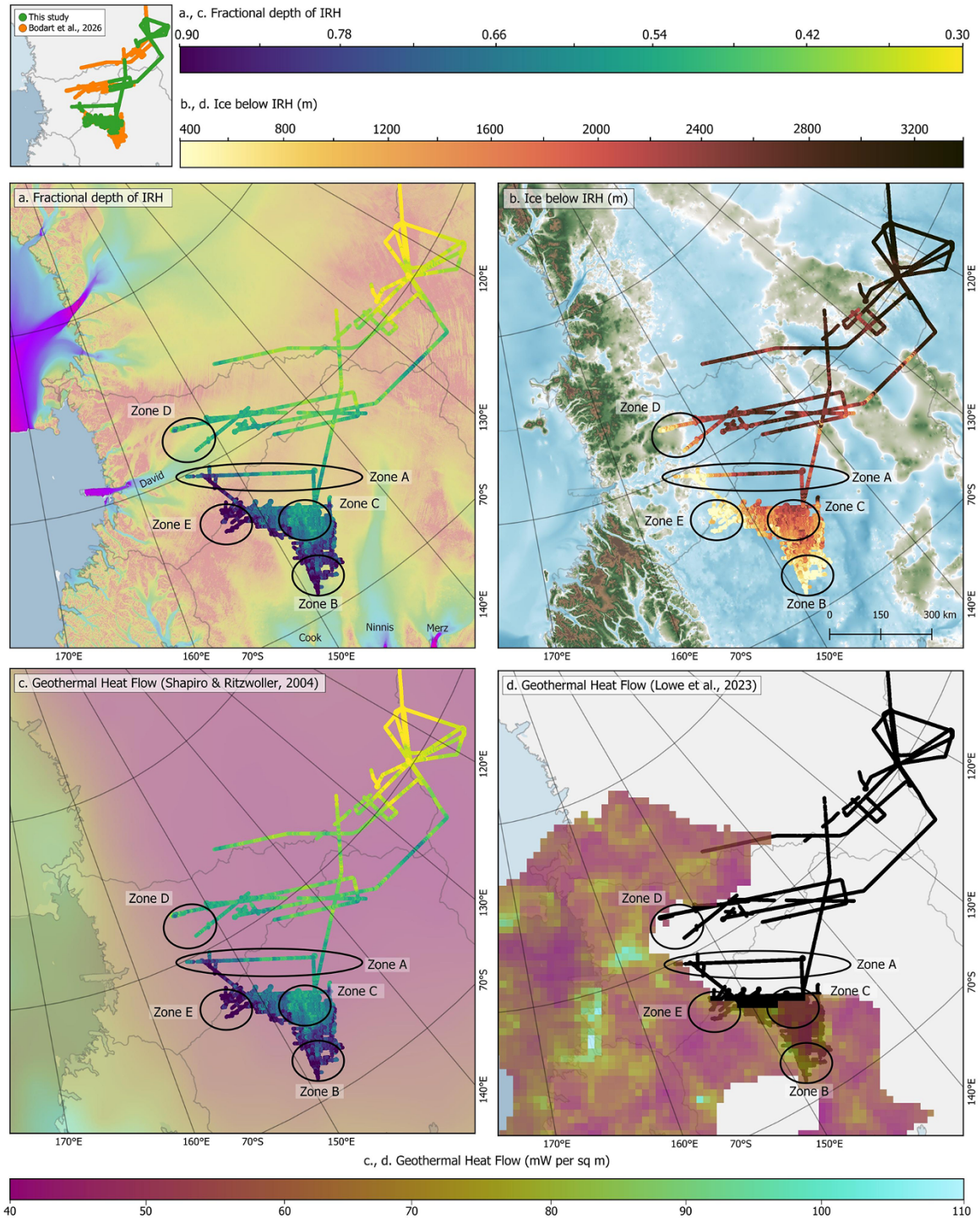


Figure 7. The 90.3 ± 2.3 ka IRH in terms of (a, c) fractional depth, (b, d) amount of ice below IRH. The backgrounds show (a) logarithmically scaled ice-sheet velocity (Rignot and others, 2017), (b) bed topography (Pritchard and others, 2025), (c) seismically derived geothermal heat flow (Shapiro and Ritzwoller, 2004), (d) geothermal heat flow derived from the joint inversion of gravity and magnetic data (Lowe and others, 2023). Top-left insert shows coverage of IRH tracing from this study (green) and Bodart and others (2026) (orange). Colourmap references for (a) and (b) are available from previous Fig. 4 and 5, respectively.

glaciers, can cause increased ice flow and thinning into inland regions of the catchment draining the glacier (Fahnestock and others, 2001; Schoof, 2007; Beem and others, 2018), which would cause drawdown of IRHs as visible in Zone B. Our study has provided a dataset that could be used to constrain whether the current geometry results from largely stable current IRH drawdown

patterns, or contains a signal of past dynamic thinning from this region.

Zone C, where we have highlighted that there is proportionally a larger volume of ice older than 90.3 ± 2.3 ka lying below the deepest dated IRH (Fig. 7), suggests that central WSB could contain some of the oldest ice in WSB, the investigation of which

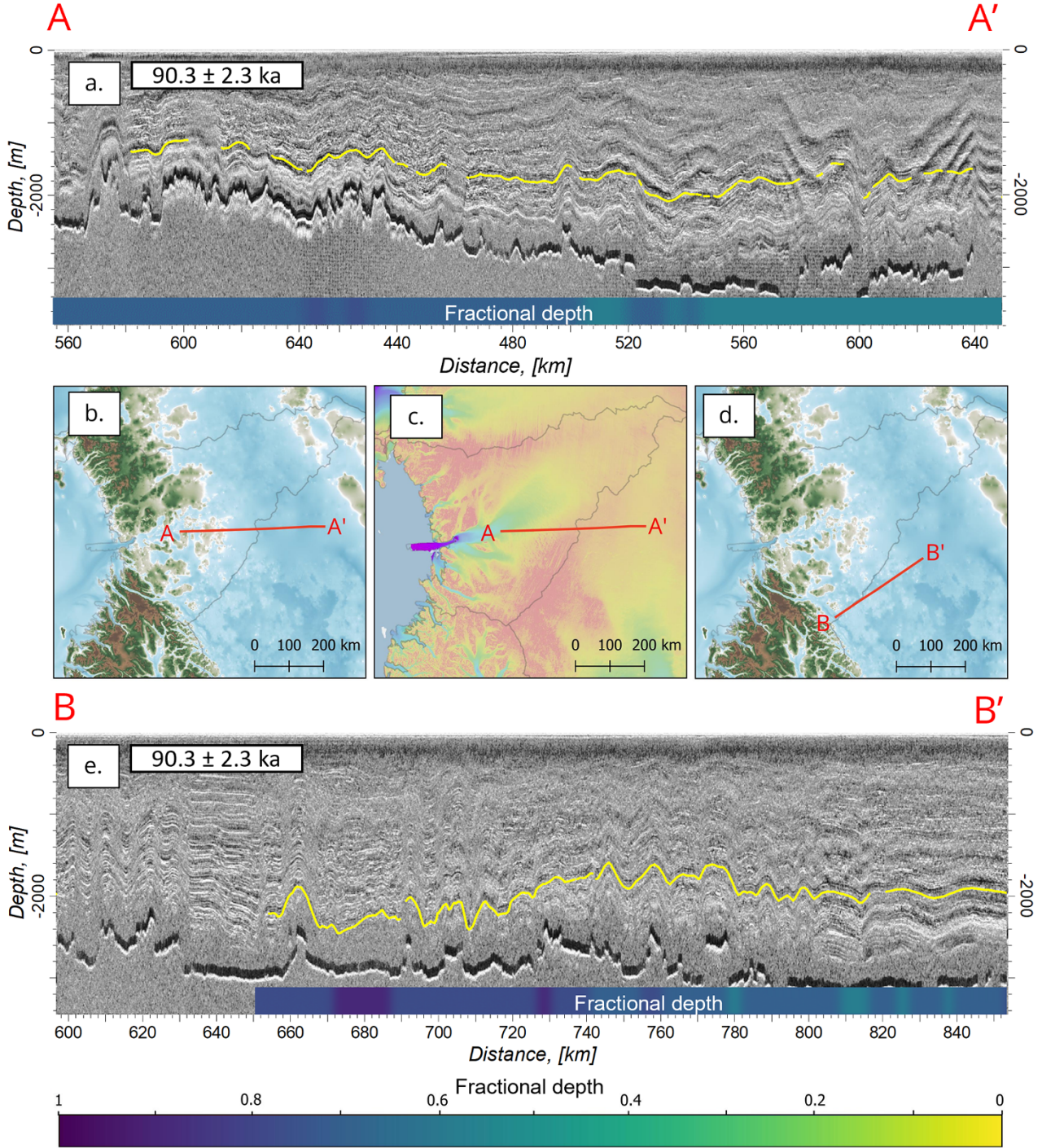


Figure 8. Drawdown of the 90.3 ± 2.3 ka IRH over distance, shown in radargrams from Zones A (A-A') and E (B-B'). The equivalent fractional depth at the given distance for the traced IRH. Here displayed over bed topography from Bedmap3 (Pritchard and others, 2025) (c, e), and ice velocity from Rignot and others (2017) (d).

could shed further light into the deeper glacial history of WSB. This zone is located in an area with deep bed topography and slow-flowing ice, and the shallow fractional depth of the IRH could indicate relatively little basal melt, or long-term lower accumulation rates. Meanwhile, Fig. 7 shows that ice flowing through Zone D is decreasing in absolute depth, but staying at a constant fractional depth, as it encounters and flows over the Transantarctic Mountains. This shows that while ice is thinner in this area, there is still a significant proportion of ice below, and thus older than,

the 90.3 ± 2.3 ka closer to the ice surface than in central WSB and therefore potentially more accessible to borehole sampling. However, such old ice will also have been subjected to significant stresses from the overlying ice weight and flow, such that the continuity of old-ice records may be compromised by vertical thinning or warping of individual isochrones.

The sharp drawdown of IRHs towards Talos Dome (Zone E in Fig. 7, and see Fig. 8), unlike for Zones A and B, does not occur at a contemporary glacier onset zone, and is also not associated

with marked changes in ice flow, ice thickness or bed topography. There are, however, potential transitions in snow accumulation and GHF at this site compared with elsewhere across upper WSB. Accumulation rates at Talos Dome are significantly higher than at any other dome in East Antarctica, with a mean accumulation rate of 80 mm yr^{-1} (Stenni and others, 2002), compared with around 25 mm yr^{-1} on the Dome C Plateau (Stenni and others, 2002; Cavitte and others, 2018; Medley and Thomas, 2019). Ice-sheet-wide maps from RACMO2 (Van Dalum and others, 2025) indicate slightly higher accumulation around Zone E compared with further upstream in upper WSB. Moreover, Zone E occurs where GHF may be notably higher than further inland in the WSB, which would also draw its IRHs down towards the ice base—this is supported by elevated GHF at Zone E modelled by Lowe and others (2023) (Fig. 7). Other studies (Shapiro and Ritzwoller, 2004; Fox Maule and others, 2005; Burton-Johnson and others, 2020; Hazzard and Richards, 2024) also signal elevated GHF in this region, compared with the inland portion of WSB.

The above has demonstrated that englacial stratigraphic patterns can be used to make regional inferences on past and present ice dynamics, accumulation and basal melting directly from observations, which adds details to our understanding of ice-sheet behaviour in the region. A further important aspect of the $90.3 \pm 2.3 \text{ ka}$ IRH is its ubiquity and easy traceability throughout much of WSB. Previous studies have shown that for IRHs to remain traceable, ice must normally remain continuously slow-flowing and unperturbed over time (e.g. Rippin and others 2003; Bingham and Siegert 2007; Sanderson and others 2023; Bingham and others 2025). Therefore, we conclude that the large-scale preservation of the $\sim 90 \text{ ka}$ IRH indicates that the ice sheet in upper WSB, where this study has focused, must very likely have been slow-flowing and ice-dynamically unchanged since at least 90 ka , covering a large part of the late Pleistocene and Holocene since the LIG. Given that below the deepest points of the $90.3 \pm 2.3 \text{ ka}$ IRH lies another 15% of ice depth, it is probable that the upper WSB was also glaciated throughout the LIG, $\sim 130\text{--}115 \text{ ka}$, thus confirming findings of previous modelling work (e.g. Sutter and others 2020), that the catchment did not experience a complete collapse during or since the LIG. However, this assumes that the ice stratigraphy below the $90.3 \pm 2.3 \text{ ka}$ IRH is continuous, which we have not been able to confirm. Future work with ice-sheet modelling constrained by surrounding IRH geometry (Sutter and others, 2021), or drilling into the deepest ice at this location, would be required to constrain definitively the LIG (and Quaternary) glacial history of this region.

Stratigraphic connections across East Antarctica

This study has connected EDC with TALDICE through our $38.7 \pm 2.8 \text{ ka}$ IRH, thereby contributing to the production of a wider RES-derived englacial stratigraphy across the EAIS that now connects TALDICE and WSB to South Pole via previous englacial links to Dome A, Vostok and EDC (Fig. 9 and Cavitte and others 2016; Winter and others 2019; Beem and others 2021; Sanderson and others 2024; Bodart and others 2026). Similar englacial stratigraphy, including an IRH also dated to $\sim 38 \text{ ka}$, has been traced across Dronning Maud Land to Dome Fuji (Fig. 9 and Winter and others 2019; Wang and others 2023; Franke and others 2025). The widespread recognition and tracing of these IRHs across much of East Antarctica makes them particularly useful for understanding ice sheet-wide patterns in englacial stratigraphy and as observational data in modelling studies (e.g. Sutter and others 2021; Bodart and others 2026); we recommend that future studies tracing IRHs

into new regions of East Antarctica target these isochrones to further build on this growing database.

Observing the full spatial perspective afforded by Fig. 9, and using the $\sim 38 \text{ ka}$ IRH as indicative of overall IRH geometry, the shallowest IRHs (by fractional depth) are generally found in the deep interior, while clear IRH drawdown occurs around the flanks of the EAIS, for example in Dronning Maud Land, Aurora/Sabrina Subglacial Basins and eastern WSB towards Talos Dome, as well as towards the Transantarctic Mountains from the South Pole. In Dronning Maud Land, Franke and others (2025) attribute IRH drawdown to the onset of fast ice flow through outlet streams, causing thinning and associated with some basal melt, as we also infer for WSB Zones A and B, respectively, the contemporary onset zones for David and Cook Glaciers. In the Aurora/Sabrina Subglacial Basins, there is a notable drawdown in IRH depth as ice traverses the subglacial ridge separating Aurora and Sabrina basins (Fig. 9), highlighting a potentially significant role for this topographic threshold in separating what could be different behavioural regimes in the ice. This could be a result of variable accumulation rates in the more coastal Sabrina Subglacial Basin (Van den Broeke and others, 2016), or contrasting conditions such as changing ice flow or ice-bed interactions causing increasing drawdown in the Sabrina Subglacial Basin. In the South Pole region, IRH drawdown has been noted previously and potentially attributed to locally high GHF (Jordan and others, 2018; Sanderson and others, 2024); this may provide an analogy for the IRH drawdown we have noted at Feature E in WSB, which also corresponds to a region where modelled GHF is relatively high for the region (Fig. 7d).

The primary pattern of the $\sim 38 \text{ ka}$ IRH across East Antarctica comprises deepening of IRHs by fractional ice depth from interior to coast, over which is superimposed a secondary pattern of regional anomalies influenced by ice dynamics, climate and ice-bed conditions. This ice-sheet-wide picture supplies important context for our understanding of the variations in these conditions across the ice sheet at large, as well as providing important information for ice-sheet modelling studies. In addition to the $\sim 38 \text{ ka}$ IRH, five more of the eight IRHs that we chose to trace across WSB (recall that they were chosen for their initial characteristics of distinctive brightness or continuity) are the same IRHs also selected for tracing by others across the EAIS when age and depth uncertainties are considered (Table 2). In particular, IRHs of similar ages to our $38.7 \pm 2.8 \text{ ka}$ and $90.3 \pm 2.3 \text{ ka}$ IRHs have been traced in all seven other of the tabulated studies (Table 2).

Given their ubiquity, it is highly likely that the two IRHs ($\sim 38 \text{ ka}$, $\sim 90 \text{ ka}$) that have been so uniformly recognised across different RES systems, and so widely traced across the EAIS, represent IRHs of notably high reflectivity and continuity relative to the majority of IRHs across the ice sheet. It is well explored that changes to relative reflectivity in different IRHs often derive from variations in the chemical composition of snow as it initially accumulates onto the ice sheet (Miners and others, 1997; Siegert, 1999; Winter and others, 2017). Particularly strongly reflective IRHs, typically associated with peaks in ice conductivity, have been linked to major debris-deposition events significantly increasing the acidity of the associated IRHs, most commonly from volcanic eruptions (Millar, 1981; Winter and others, 2019; Bodart and others, 2021; Franke and others, 2025). For instance, Franke and others (2025) attribute the $\sim 38\text{-ka}$ IRH they traced in Dronning Maud Land to a large-magnitude bipolar eruption documented by Lin and others (2023) as having occurred in the lower latitudes or Southern Hemisphere around 38.13 ka . This event falls within the age bounds of our $38.7 \pm 2.8 \text{ ka}$ IRH. Bodart and others (2026) attribute the $\sim 90\text{-ka}$ IRH

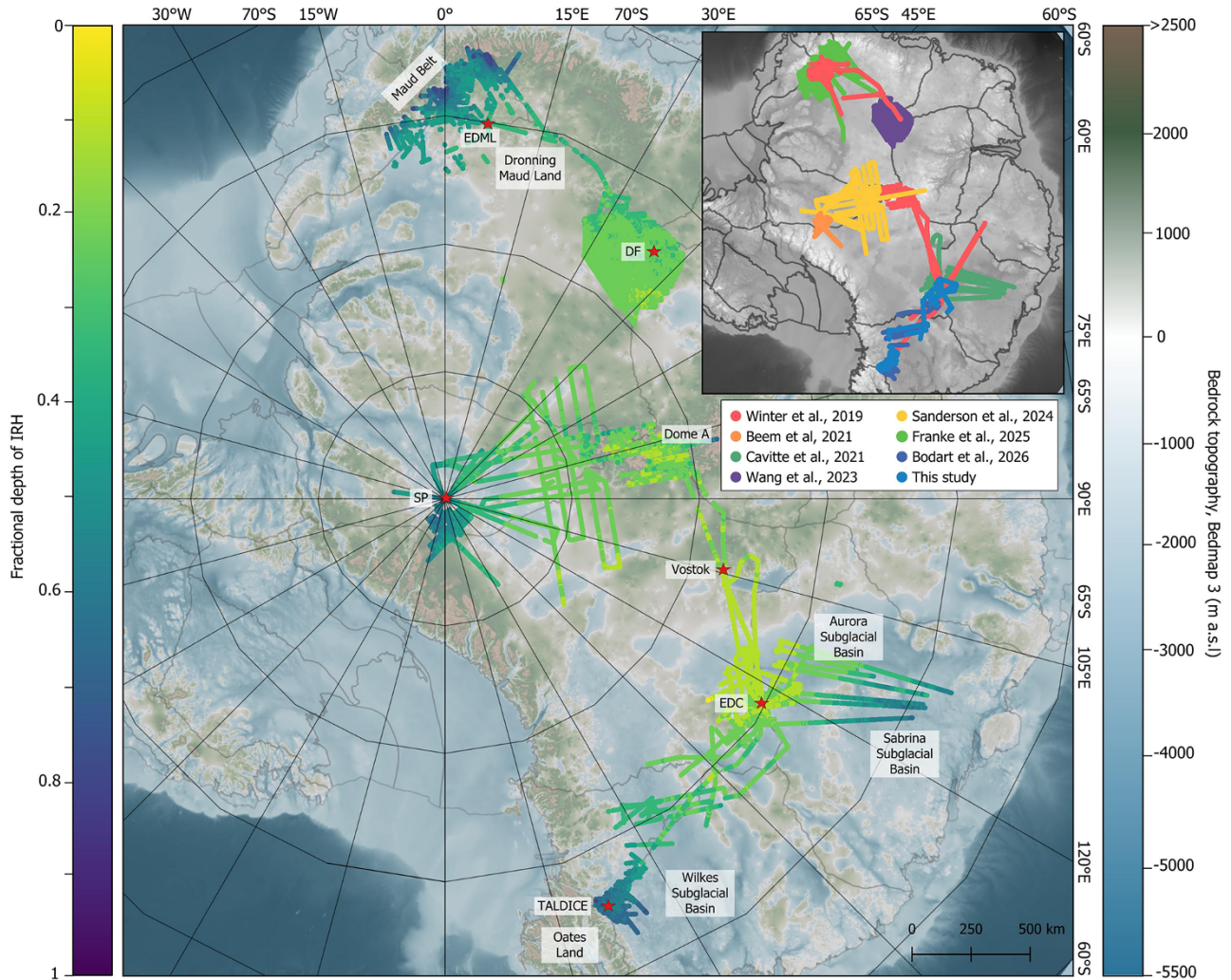


Figure 9. Fractional depth of the ~ 38 -ka IRH traced in multiple studies across East Antarctica. Upper right-hand inset shows the IRH tracing coverage from each study. In total, the ~ 38 -ka IRH has been traced over eight studies (including this study, connecting 7 ice cores). Background is Bedmap3 (Pritchard and others, 2025), with drainage basin outlines from Zwally and others (2012).

traced in the WSB-area to a large volcanic eruption, originating from Mt Berlin, West Antarctica, that was first reported and dated by Narcisi and others (2006) to 92.5 ± 2.0 ka in an EDC tephra layer; this event falls within the age bounds of our 90.3 ± 2.3 ka IRHs.

It is, however, possible that other debris sources, such as wind-blown dust, may also contribute to IRH distinctiveness and could be the reason for the ubiquitousness of certain isochrones within our study area (Fig. 5). Sugden and others (2009) noted a correspondence between dust peaks in EDC ice-core records and stages in Patagonian glacial cycles where the glaciers discharged onto large outwash plains, leading to high quantities of sediment available for wind-driven erosion. They found dust peaks in EDC to match these stages at ~ 20 , ~ 24 and ~ 65 ka. In our study, we dated one of the IRHs we traced out of TALDICE to 63.2 ± 4.8 ka (Fig. 4, Table 1), corresponding with the last such age. We therefore suggest that it need not only be volcanic fallout events that are represented by some of the most immediately traceable IRHs in Antarctica's ice. As the community further expands the tracing of IRHs across Antarctica, we will increasingly be able to compare IRH chemistry with ice cores to determine better the causes of IRH distinctiveness

in each case, while the overall geographical distribution of each IRH may present further clues as to its source as volcanic fallout, wind-blown dust or potential other sources of high IRH reflectivity.

Lessons for IRH tracing and AntArchitecture

This study, along with others in the SCAR AntArchitecture initiative (e.g. Ashmore and others 2020; Bodart and others 2021; 2023; Wang and others 2023; Sanderson and others 2024; Franke and others 2025), has demonstrated that there are some particularly distinctive IRHs that are pervasive and, in principle, more readily traceable across the Antarctic Ice Sheet. However, as exemplified in this study, not every IRH can be traced continuously between the available ice-core sites. For example, here we could not trace most of our IRHs directly all the way between EDC and TALDICE, and we did not identify in our data from WSB a particularly clear IRH corresponding to the ~ 73 ka IRH found elsewhere in East Antarctica (Winter and others, 2019; Franke and others, 2025), attributed to fallout from an eruption of Indonesia's Mount Toba (~ 74 ka) (Svensson and others, 2013). Furthermore, while the IRH we attribute to 90.3 ± 2.3 ka was particularly bright when initially

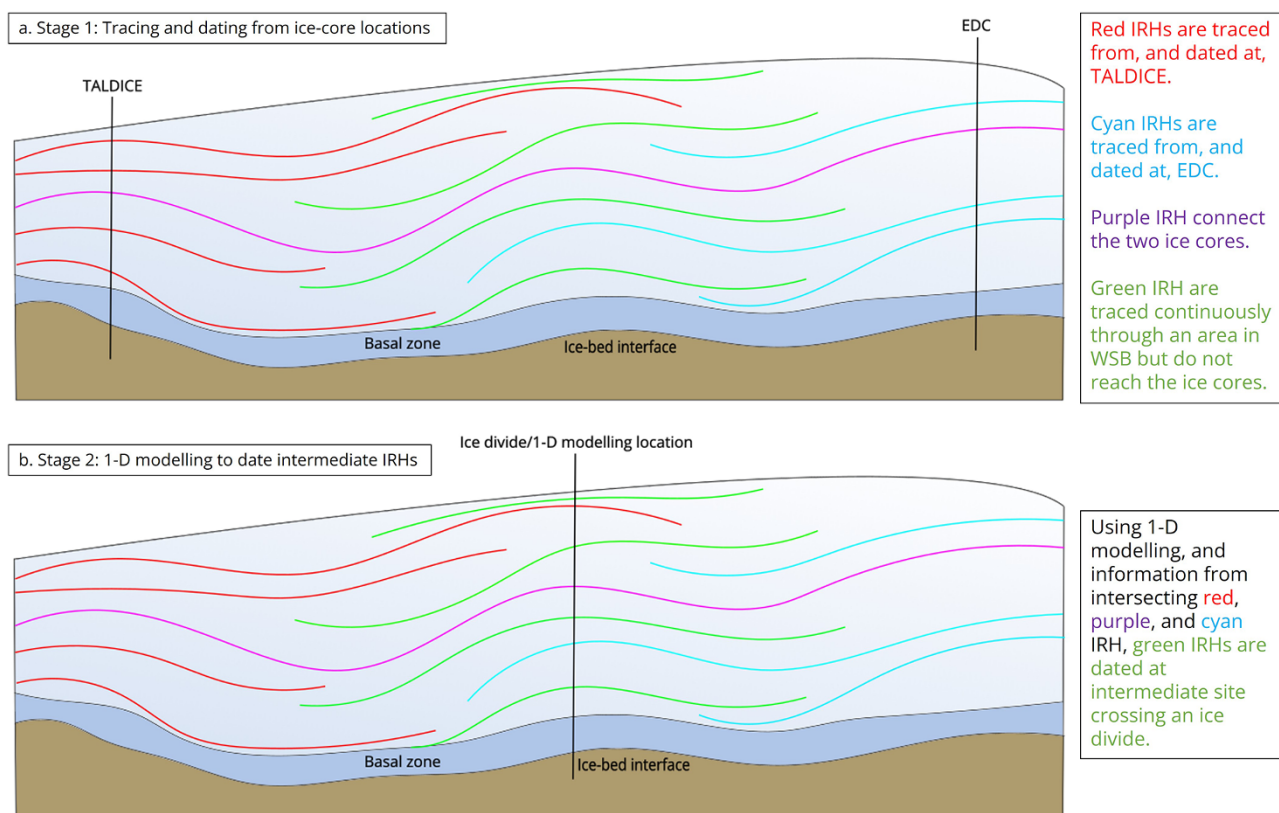


Figure 10. Schematic illustrating the idea of using 1-D modelling along ice divides to extend opportunities to trace and date IRHs. IRHs are traced through an ice-core location and dated using ice-core data (a), and with the help of these IRHs crossing an ice divide, 1-D modelling is used to date IRHs which have been traced extensively through the area, but are not traceable back to an ice core location (b).

traced crossing from Dome C Plateau into WSB, it became steadily deeper and untraceable closer towards TALDICE. This is despite this IRH being ubiquitous elsewhere across East Antarctica.

The reasons why even generally ubiquitous, particularly distinctive and potentially debris-rich IRHs such as the 90.3-ka IRH become untraceable across some areas of the ice sheet are multiple, arise from ice conditions or RES specifications depending on the individual site and survey, and may be due to, for instance, significant buckling or disruption from fast or changing ice flow, or flow over steep or complex bed topography (Bingham and others, 2025). These reasons are discussed at length elsewhere (see Bingham and others 2025, and e.g. Holschuh and others 2017).

The fact that not every IRH can be directly followed uniformly across Antarctica is becoming increasingly apparent as a growing number of IRHs are traced regionally in different studies across Antarctica, and patterns of certain brighter and more easily traceable IRHs are emerging. This means that new strategies are needed to enable the combination of IRHs directly traced and dated at ice-core intersections, with age-depth constraints acquired along RES tracks from other methods. We illustrate a potential concept for the use of one-dimensional modelling to enable dating and further tracing of IRHs that otherwise do not intersect with an ice-core location (Fig. 10). This concept is broadly based on the tracing methodology we have employed across WSB, between EDC and TALDICE. We propose that the pathway ahead is to incrementally build up an age-depth model for Antarctica's ice by using interlocking suites of traced IRHs, marrying direct age attribution of

the ice at ice-core intersections with one-dimensional modelling of age-depth at intermediate locations close to ice divides.

The approach suggested in Fig. 10 complements that taken by MacGregor and colleagues (2015) for developing an age stratigraphy for the Greenland Ice Sheet, which made use of a 'quasi-Nye' model to vertically interpolate ages from IRHs which are not traced near an ice core. In their method, ages for undated IRHs are derived where they are vertically overlain and underlain by pairs of ice-core-dated IRHs using a best-fit vertical strain from a locally applied strain-rate age-depth model (MacGregor and others, 2015; 2025). This technique has worked effectively in developing an age-depth stratigraphy for the Greenland Ice Sheet, although doing the same for the Antarctic Ice Sheet presents additional challenges in a few forms. Firstly, for IRHs to be dated using this 1D modelling approach, the undated IRHs must be either sandwiched between pairs of ice-core-dated IRHs with limited vertical spacing between the pairs, or vertically close to them. The larger the distance, the more challenging it becomes to assume linearly related conditions in vertical strain rate and accumulation rates between the dated and undated IRHs (MacGregor and others, 2015). In Antarctica, most studies have traced at most 8 isochrones across the ice column (excluding ice-core sites; see Fig. 7 of (Bingham and others, 2025)) compared with > 30 isochrones spread spatially across Greenland (MacGregor and others, 2025), making the vertical spacing between dated pairs rather large in Antarctica. Secondly, as opposed to Antarctica, open-access Greenland radar data have largely been acquired by one

data provider, using primarily different versions of one radar system and similar processing techniques. For this reason, the radar data from which Greenland IRHs have been traced have relatively similar resolution and characteristics, resulting in relatively comparable IRHs regardless of when and with what system the data were acquired and how they have been processed. In Antarctica, however, the diversity in data acquisition methods and processing (i.e. radar systems, processing, surveying objectives, bandwidth and system resolution, amongst others) makes this approach more difficult, as IRHs can appear very different across radar datasets, in turn affecting the tracing and also the assumptions made in the 1-D model to date the IRHs.

In the approach we suggest, the 1-D modelling method could make use of, for instance, the quasi-Nye approach (MacGregor and others, 2015), or other 1-D modelling methods that have been applied in other studies, such as the Dansgaard and Johnsen 1-D vertical ice-strain rate model (Dansgaard and Johnsen 1969, as used by e.g. Karlsson and others 2012; Ashmore and others 2020; Bodart and others 2021), the Nye model (Nye 1957, e.g. Bodart and others 2023), or the Lliboutry approximation (Lliboutry 1979, e.g. Davis and others 2026), to create point-based age-depth measurements in new areas where not all IRHs are traceable back to an ice core. These would act as partial ‘virtual ice cores’, to be used as progressive starting points for further tracing, granted that these ‘virtual ice cores’ are located at ice divides due to constraints in 1-D ice-flow modelling methods. This would enable both the dating of previously traced but undated IRHs that otherwise do not intersect with ice cores, and generate new starting points for the tracing of certain targeted IRHs that have been uniformly traced elsewhere, such as the 38, 73, or 90 ka IRH. We acknowledge that, as with MacGregor and colleagues (2025), most dates derived from modelling at virtual ice cores will include greater uncertainties than those derived from direct ice-core intersections. Nevertheless, this methodology comprises a solution for extending the englacial stratigraphy to regions devoid of ice cores (e.g. Aurora Subglacial Basin, Enderby Land). Furthermore, in some contexts, where no ice flow can confidently be assumed, the use of 1-D modelling could provide an opportunity to address assumptions about, e.g. climate forcing and accumulation, and compare this against the observed isochrones (e.g. Bodart and others 2023; Sanderson and others 2023; Davis and others 2026; Napoleoni and others 2026). In WSB, there is an ice divide running diagonally between the Dome C Plateau and Talos Dome (see Fig. 1), and 1-D modelling should therefore be possible along this divide. It would be interesting to see if a ‘virtual ice core’ generated from already-traced IRHs here would enable further tracing of, for instance, the 73-ka IRH, which, despite being extensive elsewhere in East Antarctica (e.g. Franke and others 2025), we did not find to be uniformly traceable across WSB in this study.

Conclusions

We have traced englacial stratigraphy across Wilkes Subglacial Basin, East Antarctica and ascribed ages to eight isochrones using their intersections with one or both of the EPICA Dome C and Talos Dome ice cores. The isochrone ages span from 13.1 ± 0.5 to 122.5 ± 2.0 ka, covering the period since the Last Interglacial until just before the start of the Holocene. One of these isochrones, dated to 38.7 ± 2.8 ka, was traced directly between EPICA Dome C and Talos Dome Ice Cores, linking with previously made tracing connections to Vostok and the South Pole. The dataset thereby made here can be used as a three-dimensional observational record

of glacial history in Wilkes Subglacial Basin, to be further studied alongside other observational data, or used as observational constraints in three-dimensional ice-sheet models.

The englacial stratigraphy shows that the age-depth profile of the ice across Wilkes Subglacial Basin is, as expected, significantly affected by the contemporary and long-standing ice-flow regime that draws ice downwards at the onset zones of Cook and David Glaciers. We identify five zones of interest where observations show how this regime interacts with the isochrones. This includes observations of ice drawing downwards at the onset zones of Cook and David Glaciers, areas of particular interest for investigating potential geothermal heat flow interactions with the ice sheet, and an indication of the areas where some of the oldest ice lies within WSB, with some regions of the upper basin, ~ 300 km from the grounding line, containing 20% of the ice column below 90.3 ± 2.3 ka. While we could not trace isochrones older than 90 ka into this region, and therefore cannot definitively conclude whether ice was continuously present here during the Last Interglacial, we postulate from the depth of ice older than 90 ka that it is highly likely that upper WSB was last deglaciated much earlier than the Last Interglacial.

This study adds to a growing network of mapped englacial stratigraphy across East Antarctica, and we have shown that a particularly bright IRH of ~ 38 ka age, already mapped across most of East Antarctica, is also pervasive across WSB. These radar-imaged isochrone links act as important validation for each age-depth record determined at the local sites by ice-core acidity peak-matching (Bouchet and others, 2023), and contribute towards the development of a pan-Antarctic 3-D age-depth model.

Data availability statement. The dataset of newly traced and dated IRHs across the Wilkes Subglacial Basin and Dome C Plateau, generated in this study, is available on Zenodo (<https://doi.org/10.5281/zenodo.21535143>).

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