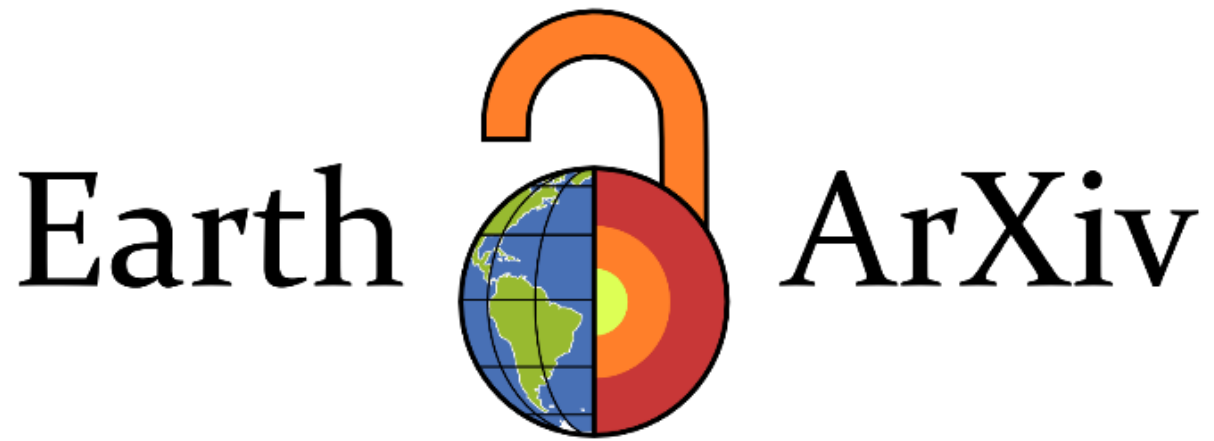




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Submesoscale frontal dynamics modulate ocean heat transport beneath Antarctic sea ice

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Antarctic sea ice has experienced unprecedented declines in recent years, yet the processes governing the transfer of subsurface ocean heat toward the ice remain poorly understood. Here, sub-kilometer hydrographic observations collected by an instrumented seal reveal intense, intermittent, and bidirectional vertical heat transport beneath sea ice in the East Antarctic marginal ice zone. During a late-summer melt regime, submesoscale frontal dynamics intermittently connect the underlying modified Circumpolar Deep Water (mCDW) reservoir to the mixed layer, producing localized km-scale heat transport events exceeding $1,000 \text{ W m}^{-2}$. A subsequent autumn ice-growth regime exhibits weaker frontal activity and substantially reduced vertical heat transport. These observations suggest that warm mCDW may contribute to Antarctic sea-ice loss not only by delaying winter ice growth, but also by enhancing basal melt during the melt season through km-scale ocean processes.

Introduction

Antarctic sea-ice extent has experienced unprecedented declines in recent years [1, 2, 3], motivating renewed efforts to understand the processes governing ocean–ice–atmosphere interactions. In the Antarctic marginal ice zone (MIZ), sea-ice evolution emerges from interactions between atmospheric forcing, ocean circulation, and upper-ocean stratification [4, 5]. Atmospheric cooling and turbulent heat loss promote ice formation, whereas oceanic heat supplied by modified Circumpolar Deep Water (mCDW) can enhance basal melt and delay winter ice growth [6, 7, 8]. However, the physical processes regulating the exchange of heat between subsurface ocean reservoirs and the sea-ice–ocean boundary remain poorly understood [9, 10, 11].

Recent studies suggest that mesoscale and submesoscale dynamics (spatial scales of $O(1\text{--}50\text{ km})$ and timescales of $O(\text{hours--weeks})$) may play a central role in mediating this exchange. Submesoscale fronts can drive intense vertical motions capable of transporting heat between the ocean interior and the surface ocean [12, 13]. Around Antarctica, sea-ice melt, brine rejection, and polynyas generate strong lateral buoyancy gradients and shallow stratified mixed layers that are conducive to energetic frontal activity [9, 10, 14]. At the same time, the strong stratification associated with meltwater layers may inhibit vertical exchange and isolate the mixed layer from subsurface heat reservoirs [11]. The net impact of sea-ice-induced stratification and frontal activity on ocean heat transport beneath Antarctic sea ice therefore remains poorly understood, owing in part to the scarcity of *in situ* observations. Recent studies have shown that under different hydrographic conditions, km-scale frontal motions enhance upward heat delivery toward the ice base and redistribute heat downward into the ocean interior [15, 16].

Animal-borne sensors provide a unique opportunity to observe the ocean beneath sea ice at high spatial resolution. Instrumented marine mammals have been widely used to sample oth-

erwise inaccessible regions of the Southern Ocean, providing detailed measurements of water-mass structure and variability [17, 18]. These observations are particularly valuable for studying fine-scale variability in the Antarctic MIZ [19, 15, 20].

Here, we combine novel sub-kilometer hydrographic observations collected by an instrumented seal with satellite and atmospheric reanalysis products to investigate ocean–ice–atmosphere interactions in the East Antarctic MIZ (Fig. 1a). We identify two contrasting seasonal regimes: a thick ice-front regime during late-summer melt conditions and a thin-ice regime associated with autumn sea-ice growth downstream of a coastal polynya (Fig. 1b). Using a diagnostic framework to estimate vertical velocities and heat transport, we show that submesoscale frontal dynamics generate intense, intermittent, and bidirectional vertical heat exchange beneath sea ice. The strongest transport events occur in the thick ice-front regime, where frontal circulations directly access the underlying mCDW heat reservoir and provide a pathway for subsurface heat to reach the ice–ocean boundary. These observations suggest that the influence of mCDW on Antarctic sea ice depends not only on the availability of subsurface heat, but also on stratification, which is dominated by salinity, and the fine-scale ocean processes that regulate heat exchange exchange with the upper ocean.

Regional setting and sea-ice conditions

In the East Antarctic marginal ice zone, a $\sim 4,000$ km trajectory sampled by an instrumented Southern Elephant Seal over more than 10 weeks captures a wide range of ocean-ice-atmosphere conditions during the transition from the late-summer sea-ice melt season to the onset of autumn sea-ice growth (Fig. 1a; [21]). Throughout most of the record, the seal follows the 1,000 m isobath, sampling both the continental shelf and adjacent continental slope waters (white line in Fig. 1). The trajectory crosses the MIZ between Queen Maud Land and Enderby Land, including Lützow-Holm Bay and the Shirase Glacier region [22]. This region is known to host

strong shelf-slope exchange that allows mCDW to access the continental shelf and interact with cold shelf waters in the vicinity of Lützow-Holm Bay and the Shirase Glacier system [23, 24]. Mean sea-ice concentration (SIC) ranges from > 0.9 to near open water conditions, revealing a transition from high-concentration ice ($34\text{--}52^\circ\text{E}$) to a weaker SIC toward lower longitudes ($<34^\circ\text{E}$), suggesting that the seal track spans both a compact ice-front system near the coast in mid January - late February 2024 and a more fragmented thin ice regime later in March 2024 (Fig. 1a).

High-resolution synthetic aperture radar (SAR, spatial resolution of tens of meters, see Methods) imagery further resolves the fine-scale structure of the ice cover along the trajectory (Fig. 1b). The SAR imagery highlights the spatial heterogeneity of sea-ice conditions at sub-kilometer scales, including the mélange near the Shirase glacier terminus, sea-ice leads, the footprint of mesoscale eddies and submesoscale filaments, icebergs of various sizes, and fragmented ice structures that are not captured by coarse-resolution SIC products (Fig. 1a). In particular, two distinct regimes emerge. In Lützow-Holm Bay, the seal encounters a thick ice front characterized by dense, compact ice with sharp boundaries, marked by the footprint of underlying meso- and submesoscale turbulence; a cyclone with a diameter of ~ 40 km and numerous submesoscale fronts with a width of 1–10 km are clearly visible (Figure 1b). In contrast, the western sector is marked by a more fragmented ice field composed of icebergs and thin ice, likely advected in part from the nearby coastal polynya visible on the SAR image at $\sim 69^\circ\text{S}, 32.2^\circ\text{S}$ (Fig. 1b) and on a MODIS visible image (ED Fig. 1).

These observations provide a unique view of the East Antarctic MIZ, spanning the transition from the late-summer melt season to the onset of autumn sea-ice growth while resolving fine-scale variability across the continental shelf and slope. We use these observations to investigate sea-ice-ocean coupling by quantifying submesoscale vertical velocities and heat transport beneath sea ice, and by examining how these processes evolve across contrasting melt-season

and ice-forming conditions.

Water-mass structure across contrasting ice regimes

Temperature–salinity (TS) diagrams reveal a broad range of water masses characteristic of the East Antarctic MIZ, while highlighting marked hydrographic contrasts between the late-summer ice-front and autumn ice-forming regimes (Fig. 2). Near the surface, low-salinity waters define a meltwater layer close to the freezing point (white dots), reflecting freshwater input from sea-ice melt. Beneath this layer, a temperature and salinity maximum identifies mCDW and Circumpolar Deep Water (CDW), which together form a subsurface reservoir of ocean heat [7]. Winter Water, a colder, fresher and relatively homogeneous water mass that forms through surface cooling and brine rejection during sea-ice growth, occupies the region between the deep CDW and surface meltwater layers [25, 26, 27]. The freezing line delineates the boundary between supercooled and above-freezing conditions, with some near-surface waters falling below the local freezing point.

During the thick ice-front period, the hydrographic observations span a broad range of temperature and salinity values and include a larger fraction of relatively warm and saline waters associated with mCDW (Fig. 2b). These water masses reveal the coexistence of surface meltwater, Winter Water, and a substantial reservoir of ocean heat associated with mCDW beneath the sea ice.

In contrast, the thin-ice period is characterized by a stronger imprint of Winter Water and supercooled waters, with observations occupying a narrower range of temperature–salinity properties and exhibiting a reduced contribution from mCDW (Fig. 2c). The prevalence of supercooled waters is consistent with recent observations of ice-forming conditions in the Southern Ocean [28]. Overall, the hydrographic properties are more homogeneous and occupy a colder region of temperature–salinity space than during the thick ice-front period.

Together, these TS characteristics reveal a transition from a late-summer regime influenced by surface meltwater, Winter Water, and mCDW to an autumn ice-forming regime characterized by Winter Water and supercooled surface waters. Some of the observed variability likely also reflects spatial differences along the seal trajectory. The upstream region is associated with more mCDW over the shelf break (Fig. 1), and therefore will be a region of stronger ice-shelf melt, implying that the upper part of the water column may be warmer than the downstream thin ice regime due to the influence of glacially-modified mCDW. These hydrographic contrasts provide the background for the differences in sea-ice–ocean coupling, submesoscale dynamics and vertical heat transport explored below.

Ocean-ice-atmosphere coupling and submesoscale variability

Along-track sections reveal pronounced contrasts in ocean–ice–atmosphere conditions between the two regimes and strong km-scale variability (Fig. 3). During the thick ice-front period, sea-ice concentration fluctuates between ~ 0.3 and 0.85 , while relatively mild atmospheric conditions prevail, with 2 m air temperatures ranging from $+1$ to -3°C (Fig. 3a). Temperature and salinity sections show a stratified upper ocean, with cold Winter Water overlying warmer and saltier subsurface mCDW (Fig. 3b,c). The mixed layer is shallow, with a mean depth of ~ 20 m. Strikingly, relatively warm waters (above 1°C) are observed within and immediately below the mixed layer, often coinciding with localized intrusions of saltier waters, for example near 1,600 km (Fig. 3c). These features are associated with pronounced upward tilting of isopycnals and upward excursions of the warm, saline mCDW layer toward the mixed layer.

Lateral buoyancy gradients (b_x) reveal intense coherent structures indicative of submesoscale fronts (Fig. 3d), with typical magnitudes of $\sim 4 \times 10^{-7} \text{ s}^{-2}$, near the upper range of reported values in the Southern Ocean [29, 30, 19, 12, 31, 15]. These fronts extend well below the mixed layer, reaching depths of up to 500 m and often exhibiting a slanted structure indica-

tive of energetic (sub)mesoscale dynamics at the ice–ocean boundary [9]. Exceedance statistics further reveal a prevalence of strong buoyancy gradients during the thick ice-front period both within and below the mixed layer compared to the full trajectory and to the thin ice period (Fig. 4). Within the mixed layer, where approximately 10% of observations exceed 10^{-6} s^{-2} (Fig. 4a), strong buoyancy gradients coincide with low-salinity surface waters, suggesting that freshwater variability associated with sea-ice melt contributes to the observed frontal structure. Positive net surface heat fluxes ($Q_{\text{net}} > 0$ from ERA5, see Methods; Fig. 5a and ED Fig. 5a) further indicate atmospheric conditions favorable for sea-ice melt during this period. Together, these observations point to a close coupling between meltwater fronts, atmospheric forcing, and submesoscale frontal dynamics both within and below the mixed layer.

In contrast, the thin ice regime exhibits a colder and more homogeneous upper-ocean structure, with Winter Water and supercooled waters occupying much of the mixed layer (cyan bars, Fig. 3f,g). These hydrographic characteristics occur alongside markedly colder atmospheric conditions, with 2 m air temperatures ranging from -3 to below -10°C , lower sea-ice concentrations (0.2–0.6; Fig. 3e), and negative net surface heat fluxes ($Q_{\text{net}} < 0$; Fig. 5a and ED Fig. 8a), consistent with autumn sea-ice growth conditions. The mixed layer also deepens to ~ 50 m, substantially deeper than in the thick ice front regime. This deeper mixed layer is accompanied by a weaker stratification at its base (ED Figs. 5f and 8f), resulting in a less distinct separation between the mixed layer and underlying Winter Water. Although lateral buoyancy gradients penetrate below the mixed layer to at least 300 m and reach magnitudes up to $\pm 4 \times 10^{-7} \text{ s}^{-2}$, the frontal structures are generally thinner and shallower than those observed during the thick ice front period. Strong mixed-layer fronts are also less prevalent as revealed by the exceedance statistics showing a substantially lower occurrence of intense buoyancy gradients than during the thick ice-front period (Fig. 4). Within the mixed layer less than 1% of observations exceed 10^{-6} s^{-2} compared to 10% in the thick ice period (Fig. 4a). A similar contrast persists below

the mixed layer, where strong buoyancy gradients (i.e., $> 10^{-6} \text{ s}^{-2}$) are roughly four times less frequent during the thin ice period than during the thick ice front period (Fig. 4b).

These contrasts point to potentially important differences in submesoscale dynamics between the two regimes. The early thick ice-front period occurs closer to the shelf break, where exposure to energetic mesoscale variability may enhance frontal activity [32, 20], whereas the later thin-ice period occurs predominantly over the continental shelf, where eddies are expected to be less energetic [33]. In the following section, we investigate how these differences translate into vertical velocities and associated vertical heat transport beneath sea ice.

Submesoscale vertical velocities and heat transport

We now diagnose vertical velocities (w) using the two-dimensional quasi-geostrophic omega equation and the associated vertical heat transport ($\text{VHT} \propto wT'$, where T' is the temperature anomaly) along the seal trajectory (see Methods; Fig. 3e,f,k,l). The assumptions, limitations, and uncertainties associated with these diagnostics are discussed in the Methods and quantified through a Monte Carlo uncertainty analysis presented in the Methods and in the Supplementary Information. Clear contrasts emerge between the two regimes.

In the thick ice front regime, vertical velocities are intermittent (kurtosis of 4.37) and organized into alternating bands of upwelling and downwelling that extend from the mixed layer to depths of several hundred meters (Fig. 3e). These structures of a few km-wide are closely aligned with regions of enhanced lateral buoyancy gradients (Fig. 3d), consistent with submesoscale-driven vertical motions. Vertical velocities often exceed $\pm 20 \text{ m day}^{-1}$ and locally approach $\pm 50 \text{ m day}^{-1}$. The associated vertical heat transport is also highly intermittent (kurtosis of 10.11), with localized upward and downward heat fluxes exceeding $\pm 1,000 \text{ W m}^{-2}$ (Fig. 3f, saturated colorbar). Episodes of strong upward heat transport frequently coincide with upward displacements of the warm, saline mCDW layer toward the mixed layer, highlighting

the potential for submesoscale motions to connect the subsurface mCDW heat reservoir with the sea-ice–ocean interface. Conversely, strong downward heat transport events are also common, particularly below ~ 200 m depth, where frontal overturning can upwell colder Winter Water and/or deepen the underlying warm mCDW layer. This behavior highlights the coexistence of strong upward and downward heat transport events within the same frontal system. Since buoyancy is primarily set by freshwater content, temperature behaves approximately as a passive tracer, indicating that the sign of the heat transport depends strongly on the vertical arrangement of warm and cold water masses relative to the frontal circulation.

In contrast, the thin ice regime exhibits weaker vertical velocities of at most ± 15 m day $^{-1}$ (Fig. 3j), consistent with the weaker buoyancy gradients described above (Fig. 3h). The resulting vertical heat transport is substantially reduced (Fig. 3l), with most of the water column characterized by weak heat fluxes of a few tens of W m $^{-2}$ punctuated by occasional isolated events of stronger VHT. Despite their smaller amplitude, both the vertical velocities and associated heat transport remain highly intermittent, with kurtosis of 5.16 and 23.17, respectively. Unlike the thick ice front regime, vertical heat transport does not penetrate the mixed layer.

Despite the strongly intermittent and bidirectional nature of the diagnosed heat transport, the mean VHT profiles reveal clear differences between the two regimes (Fig. 5d). The thick ice front regime exhibits a pronounced positive VHT maximum of approximately 50 W m $^{-2}$ centered near 200 m depth, coincident with the upper portion of the mCDW layer. Below this depth, mean VHT becomes negative, reaching values of approximately -50 W m $^{-2}$ near 300 m, consistent with the coexistence of upward and downward heat-transport events described above. However, uncertainty estimates from the Monte Carlo analysis indicate substantially lower confidence at these depths, with both positive and negative values remaining plausible (see Supplementary Information). As such, the negative mean VHT at depth is less robust than the positive upper-ocean maximum. In contrast, the thin ice regime is characterized by weak

mean heat transport throughout the water column, generally remaining within $\pm 10 \text{ W m}^{-2}$. The full trajectory mean lies between these two end members. Together, these profiles indicate that although instantaneous heat transport is highly variable, the thick ice-front regime is associated with substantially stronger vertical heat exchange and a net upward redistribution of heat from the mCDW layer toward the upper ocean.

Evolution of submesoscale ocean heat transport beneath Antarctic sea ice

To place these localized heat transport events in a broader context, we examine their evolution along the full seal trajectory and their relationship to topography (off-shelf/on-shelf, Fig. 1), mixed-layer structure, sea-ice conditions and atmospheric forcing from late-summer melt to autumn ice growth (Fig. 5). All atmospheric variables are shown as 24 h rolling means, which filter out the diurnal cycle.

During the late-summer portion of the trajectory (20 January–11 February), when the seal samples beneath extensive summer sea ice and crosses the thick ice front over the shelf break, net surface heat fluxes are positive, consistent with atmospheric conditions favorable for sea-ice melt (Fig. 5a). Air temperatures remain relatively mild, fluctuating around 0°C (Fig. 5b), while the mixed layer is shallow, typically less than 20 m deep (Fig. 5b), consistent with the strong near-surface stratification (ED Fig. 5f) and meltwater fronts described above. During this period, the strongest vertical heat transport events are observed, coinciding with enhanced frontal activity and elevated mean VHT (Fig. 5d). The magnitude of these events (e.g., on 02/03) is comparable to that of the net surface heat flux, suggesting that this intermittent ocean heat transport may contribute substantially to the local heat balance alongside atmospheric forcing.

A marked transition occurs at the onset of the autumn sea-ice growth during late February (dashed vertical line in Fig. 5a–c). Air temperatures decrease from approximately $0\text{--}1^{\circ}\text{C}$ to below -8°C , net surface heat fluxes become predominantly negative, and turbulent heat loss

increases substantially (Fig. 5a,b). Concurrently, the mixed layer deepens to 40–80 m and persistent easterly winds become more frequent (Fig. 5c). These easterlies, likely associated with katabatic outflows from the Antarctic continent, are consistent with conditions favorable for sea-ice formation in coastal polynyas such as the Shirase Glacier–Lützow-Holm Bay system [34, 35]. The subsequent thin ice period is characterized by weak mean VHT and a near absence of strong upward or downward heat transport events, highlighting a reduced role for submesoscale ocean heat transport in regulating sea-ice conditions during the autumn ice-growth period on the continental shelf.

Discussion

We identify a regime in which submesoscale frontal dynamics intermittently connect Antarctic sea ice to the underlying mCDW reservoir. The strongest heat-transport events occur during late-summer conditions characterized by shallow mixed layers, strong meltwater fronts, and active frontal variability, whereas autumn ice-growth conditions are associated with substantially weaker heat transport. These observations suggest that the transport of ocean heat to sea ice depends not only on the presence of mCDW but also on the fine-scale processes that regulate access to this heat reservoir.

A key implication of these observations is that ocean heat exchange beneath sea ice is not a steady upward flux, but an intermittent and bidirectional process. Intense upward heat-transport events can penetrate the mixed layer, providing a pathway by which subsurface mCDW heat can be transferred toward the sea-ice–ocean boundary, with the potential to locally enhance basal melt [9]. Equally energetic downward transport events indicate that the same frontal dynamics can also redistribute heat away from the ice–ocean boundary, highlighting that submesoscale motions that act to reduce potential energy stored in density fronts [13] are also likely to be associated with downward heat transport since warmer waters are associated with saltier and

denser waters at the Antarctic margins [36]. Ocean-driven melt may therefore be spatially and temporally heterogeneous, concentrated near transient frontal events rather than distributed uniformly beneath the ice [32]. Although upward and downward transport events coexist, the positive mean VHT near the top of the mCDW layer indicates that upward transport events dominate there on average, increasing the likelihood of contact between sea ice and mCDW-derived heat and therefore the potential for basal melt.

Previous observational studies based on mixed-layer eddy parameterizations have shown that submesoscale heat transport beneath Antarctic sea ice can be either upward or downward depending on the orientation of temperature gradients across fronts and the regional hydrographic setting [19, 16, 15]. In these frameworks, the sign of the heat flux is strongly influenced by the hydrographic structure within the mixed layer, while the overturning circulation itself remains fundamentally restratifying [37]. In the observations analyzed here, frontal motions extend well below the mixed layer and directly intersect the mCDW reservoir. Consequently, the sign of the heat transport is controlled not only by the frontal circulation itself but also by the vertical arrangement of Winter Water and mCDW. Similar overturning circulations can therefore produce opposite heat transports depending on the local hydrographic structure.

Several caveats remain. Our observations are limited to a single trajectory in East Antarctica during late summer and early autumn, and the vertical velocities and heat transports are inferred quantities that are subject to a range of assumptions and uncertainties (discussed in the SI). In addition, we do not explicitly account for either horizontal sea-ice advection or horizontal ocean heat advection, which prevents us from closing the local sea-ice and upper-ocean heat budgets. Consequently, while the diagnosed upward heat transport events are consistent with enhanced basal melt, we cannot fully quantify the relative contributions of vertical heat transport, horizontal advection, atmospheric forcing, and ice dynamics to the observed sea-ice evolution.

Although available observations do not allow this mechanism to be demonstrated directly, the close correspondence between meltwater fronts, strong buoyancy gradients, and elevated heat transport is consistent with the possibility of a positive feedback between atmospheric melt, frontal instability, and ocean heat transport. In this scenario, deep-reaching fronts established by mesoscale stirring support submesoscale frontal circulations that intermittently transfer subsurface heat toward the ice. Surface melt may sharpen the near-surface expression of these fronts, potentially reinforcing the initial melt. This interpretation is consistent with recent studies highlighting coupled ice–ocean dynamics in the marginal ice zone [9, 38, 11, 39, 33]. While warm CDW is increasingly recognized as a mechanism limiting sea-ice growth and delaying winter ice formation [7, 40, 41, 42, 43], our results suggest an additional role during the melt season: once atmospheric forcing initiates melt, submesoscale frontal dynamics may enhance the delivery of mCDW-derived heat to the ice base and locally accelerate basal melt.

The mechanisms identified here are likely relevant across broader sectors of the Antarctic marginal ice zone. Strong lateral buoyancy gradients associated with meltwater fronts and ice edges are widespread around Antarctica, providing favorable conditions for intermittent vertical exchange between surface waters and subsurface heat reservoirs [10, 11, 9, 36]. The strength of this coupling likely varies seasonally as upper-ocean stratification, mixed-layer depth, sea-ice conditions, and atmospheric forcing evolve throughout the year [44, 45, 46]. It should also vary regionally depending on the availability of subsurface heat reservoirs such as mCDW [43, 8]. Although direct observations beneath Antarctic sea ice remain sparse, recent altimetry observations suggest that energetic mesoscale eddies may persist beneath the ice cover [14].

Future increases in meltwater stratification and ice-edge buoyancy gradients associated with Antarctic ice loss, together with warming subsurface waters and changes in upper-ocean circulation, may enhance the prevalence and impact of submesoscale frontal activity beneath sea ice [47, 48, 8]. More broadly, these observations suggest that the influence of CDW on Antarc-

tic sea ice depends not only on the amount of heat stored beneath the mixed layer, but also on the efficiency with which kilometer-scale frontal dynamics intermittently connect that heat reservoir to the ice base, with implications for understanding and predicting future changes in Antarctic sea-ice cover [49].

Materials and Methods

Seal data

We use high-resolution (sub-km) data collected by an instrumented male Southern Elephant Seal, who departed from the Kerguelen Islands (49.35°S, 70.22°E, Indian sector of the Southern Ocean) and reached East Antarctica during the course of his voyage. We analyze the portion of its trajectory south of 64.5°S (64.5-69°S, 30-60°E, Fig. 1) from 01/12/2024 04:09:21 UTC to 03/22/2024 05:19:52 UTC, i.e. more than 10 weeks during the austral mid-summer to early autumn. The total cumulative distance analyzed is 3 879 km, the median horizontal distance between consecutive dives is 669 m, the maximal depth is 807 m and the median depth is 322 m. In addition, ED Figure 1 shows the number of profiles as a function of depth and ED Figures 4-8 show the full 10+ week vertical transects. The seal is localized through the Argos satellite system and is equipped with sensors that record conductivity, temperature and pressure (CTD-Satellite Relay Data Logger) at a continuous frequency of 0.5Hz [18, 50]. Only the ascending phase of a dive is used because it is more uniform in speed and direction compared to the descent when the seal dives sinuously to forage [51]. Final accuracies are estimated to be of $\pm 0.02^{\circ}\text{C}$ for temperature and $\pm 0.03 \text{ g kg}^{-1}$ for salinity after post-processing and calibration [52, 51]. The animal was handled in accordance with the ethical and Polar Environment Committee guidelines of the Institut polaire français Paul-Émile Victor (IPEV) as part of the SNO-MEMO and IPEV program 109, and all experimental protocols were approved by the corresponding IPEV Ethics and Polar Environment Committees.

Mixed layer depth The mixed layer depth is defined as the level of a 0.03 kg m^{-3} density increase from 15m depth [53].

Buoyancy and its horizontal gradients We compute horizontal gradients of buoyancy, i.e., a proxy for submesoscale fronts [54, 55, 56]. To do so, we follow the same methodology as in refs [12, 57, 58]: first we compute along-track time series of buoyancy, $b = g(1 - \frac{\rho}{\rho_0})$, where g is gravity, ρ is potential density and $\rho_0 = 1025 \text{ kg m}^{-3}$ is a reference seawater density. For the analysis, in particular the calculation of horizontal buoyancy gradients $b_x = \frac{\partial b}{\partial x}$, with x the along-track distance, buoyancy b was first linearly interpolated along the seal’s path onto a regular grid of 100 m resolution. A moving average with a 1 km window was then applied such that the final dataset had a horizontal resolution of 1 km and a vertical resolution of 1 m, yielding a total of 3880 profiles. Horizontal gradients of buoyancy are resolved by multiple vertical profiles, such that the structures are not related to aliasing of the along-track data. Potential density used to compute b is calculated from the corrected conservative temperature and absolute salinity with the TEOS10 equation [59] using the Gibbs Seawater oceanographic toolbox [60].

Vertical velocities Vertical velocities (w) are derived from the two-dimensional (x,z) quasi-geostrophic version of the omega equation, following refs [12, 57]. This equation reads

$$N^2 w_{xx} + f^2 w_{zz} = -2 (u_x b_x)_x, \quad (1)$$

where subscripts indicate derivatives and x denotes the along seal-track direction. u_x is estimated from SO-SLICE ([61], see section “Altimetry data” below) and assumed constant with depth, f the Coriolis parameter is computed as a function of latitude. $b_x(x, z)$ and $N(x, z)^2$ are estimated from the seal’s measurements. Equation (1) is solved using the `xinvert` Python

package, which provides a flexible framework for solving elliptic equations in geophysical flows [62]. A detailed discussion of the assumptions, implications and limitations of the two-dimensional omega equation is provided in Siegelman et al. (2020)[12]. To further assess the uncertainties associated with this framework, we perform a sensitivity analysis to the along-track strain and seal orientation, see ED Fig. 3a,b and section “Monte Carlo uncertainty propagation” below.

Vertical heat transport Vertical heat transport (VHT) is computed as

$$\text{VHT} = \rho_0 c_p \overline{w' \text{CT}'}, \quad (2)$$

where w' and CT' are the anomalies of vertical velocity and Conservative Temperature, respectively, obtained by removing the along-track mean profile at each depth. Here, $\rho_0 = 1.025 \text{ kg m}^{-3}$ is a reference seawater density and $c_p = 3.985 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of seawater at constant pressure. The overbar denotes the averaging operator, and the resulting flux represents the vertical transport of heat associated with submesoscale motions. A sensitivity analysis is presented in ED Fig. 3c,d (see also section “Monte Carlo uncertainty propagation” below).

Altimetry data

Along-track strain is estimated from sea level anomaly (SLA) fields from the Southern Ocean Sea Level in Ice and Open-ocean Conditions Estimate (SO-SLICE, [61]). SO-SLICE is an SLA reconstruction that combines several altimetry missions in both ice-free and ice-covered regions using lead-based retracking techniques and objective mapping, spanning 2003-2024. The version used here relies on correlation length scales of 50 km for the mapping, which better resolves the mesoscale. Geostrophic velocity anomalies are computed from SLA gradients, and horizontal strain-rate components are obtained from spatial derivatives of the geostrophic

velocity field. The strain tensor is then projected onto the local seal-track orientation to obtain the along-track strain rate,

$$\epsilon_s = \frac{1}{2} (s_n \cos(2\theta) + s_s \sin(2\theta)) , \quad (3)$$

where $s_n = u_x - v_y$ is the normal strain component and $s_s = u_y + v_x$ is the shear strain component [63]. Here, u and v denote the zonal and meridional geostrophic velocities, respectively. The angle θ represents the orientation of the seal trajectory relative to the eastward direction and is computed from the local trajectory tangent. To reduce small-scale variability associated with positioning uncertainty and short-scale meanders of the trajectory, θ is smoothed along the trajectory using a one-dimensional Gaussian filter with a standard deviation of 3 km prior to the strain projection. This scale is comparable to the estimated positional uncertainty of the seal track [51]. Sensitivity tests using smoothing scales between 2 and 5 km produced negligible differences in the resulting along-track strain estimates. The final along-track strain time series is subsequently used as forcing in the omega-equation calculations.

Monte Carlo uncertainty propagation

To assess the sensitivity of the omega-equation solutions, and subsequent VHT calculations, to uncertainty in both the projected strain forcing and the orientation of the seal trajectory, we perform a Monte Carlo ensemble experiment. For each realization, the along-track strain is recomputed from the SO-SLICE-derived normal and shear strain components after perturbing the local seal-track orientation. The perturbed along-track strain is computed as

$$\epsilon_s^{(i)} = \frac{1}{2} (s_n \cos(2\theta^{(i)}) + s_s \sin(2\theta^{(i)})) , \quad (4)$$

where s_n and s_s are the normal and shear strain components, respectively, and $\theta^{(i)}$ is the perturbed trajectory angle for realization i . The trajectory-angle perturbations are drawn from

a Gaussian distribution with zero mean and a standard deviation of 10° , then smoothed along track using a one-dimensional Gaussian filter with a standard deviation of 5 km to impose spatial coherence.

For each realization, the final strain forcing is defined as $\tilde{\epsilon}_s^{(i)}(x) = A^{(i)}\epsilon_s^{(i)}(x) + \eta^{(i)}(x)$, where $A^{(i)}$ is a realization-dependent amplitude factor drawn from $A^{(i)} \sim \mathcal{U}(0.5, 2)$, and $\eta^{(i)}(x)$ is spatially correlated Gaussian noise. The noise is generated as $\eta^{(i)}(x) = \mathcal{G}_\sigma [\eta_0^{(i)}(x)]$, $\eta_0^{(i)} \sim \mathcal{N}(0, \sigma_\epsilon^2)$, where $\mathcal{G}_\sigma$ denotes a one-dimensional Gaussian smoothing operator along the trajectory with $\sigma = 5$ km, and σ_ϵ is the standard deviation of the reference along-track strain time series. The resulting strain realizations are clipped to remain within $\pm 5 \times 10^{-5} \text{ s}^{-1}$.

An ensemble of $N = 50$ strain realizations is generated and used as forcing for independent omega-equation inversions. For each realization, vertical velocity and vertical heat transport are recomputed. The uncertainty associated with the strain forcing is quantified from the resulting ensemble using the median and the 5th–95th percentile range of the diagnosed w and VHT fields as a function of depth for both the full trajectory and selected dynamical regimes.

Sea-ice data

Daily sea ice concentration (SIC) data were obtained from the NOAA/NSIDC Climate Data Record (CDR) (G02202, Version 6), available at the National Snow and Ice Data Center. This dataset provides quality-controlled gridded, passive microwave–derived SIC estimates at 25 km spatial resolution [64]. In Figure 1a, the mean SIC (0–1) is computed from daily fields between 2024-01-12 and 2024-03-22, retaining at each time step only pixels within 100 km of the seal’s nearest position, then averaged over time (NaNs ignored).

ERA5 atmospheric variables

Atmospheric and surface forcing variables were obtained from the ERA5 reanalysis produced by the European Centre for Medium-Range Weather Forecasts. ERA5 provides hourly estimates on a ($\sim 0.25^\circ$) latitude–longitude grid by combining numerical weather prediction model output with a wide range of observations through data assimilation [65]. Variables extracted along the seal trajectory included 2 m air temperature (t_{2m}), 10 m wind speed and direction, sensible heat flux (SHF), latent heat flux (LHF), net shortwave radiation, and net longwave radiation. Turbulent heat flux (THF) was calculated as the sum of sensible and latent heat fluxes ($THF = SHF + LHF$), while the net radiative flux (R_{net}) was computed as the sum of net shortwave and net longwave radiation. The net surface heat flux was then defined as ($Q_{net} = R_{net} + THF$), with positive values indicating heat flux into the ocean.

SAR imagery

Synthetic Aperture Radar (SAR) imagery from the Sentinel-1 mission was used to characterize sea ice surface properties at high spatial resolution varying between 40 m and 420 m. Sentinel-1 provides all-weather dual-polarization backscatter measurements (σ^o) that are sensitive to surface roughness and dielectric properties, including variations in brine content, allowing discrimination between open water and different sea ice types, such as newly formed ice, thus enabling detailed analysis of fine-scale sea ice structures and dynamics [66].

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Author contributions

L.S. designed and implemented the analysis. C.M.D. processed the altimetry product used in this study. All authors contributed to the interpretation of the results and editing of the manuscript.

Competing interests

The authors declare no competing interests.

Data and materials availability

The seal data is available online at <https://meop.net/database/>, SO-SLICE data is available online on Zenodo at <https://doi.org/10.5281/zenodo.17467407>, SIC data is available online at <https://nsidc.org/data/g02202/versions/6>, SAR images are available online at <https://dataspace.copernicus.eu/data-collections/copernicus-sentinel-missions/sentinel-1>. All the codes for the analysis and figure generation are available at <https://github.com/liaSiegelman>.

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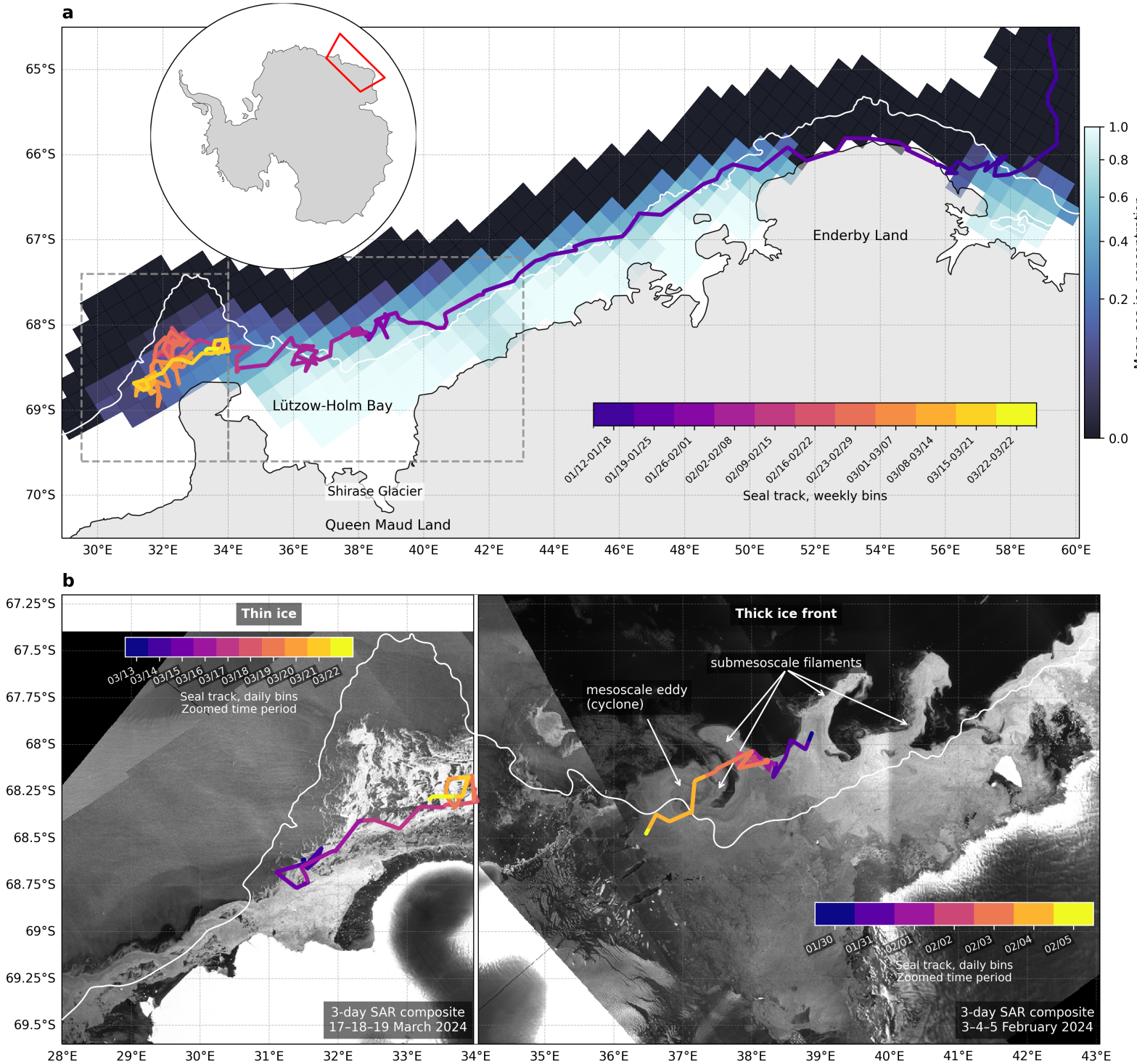


Figure 1: **Regional map** (a) Mean sea-ice concentration during the seal's voyage (NSIDC v6), overlaid with the weekly binned seal track. The inset shows the East Antarctic study region. (b) Three-day SAR composite images (Sentinel-1; processed by Ocean Data Lab), corresponding to the dashed rectangles in (a), with the seal track over the selected periods. The 1,000 m isobath from RTOPO-2 is shown by the white line [67].

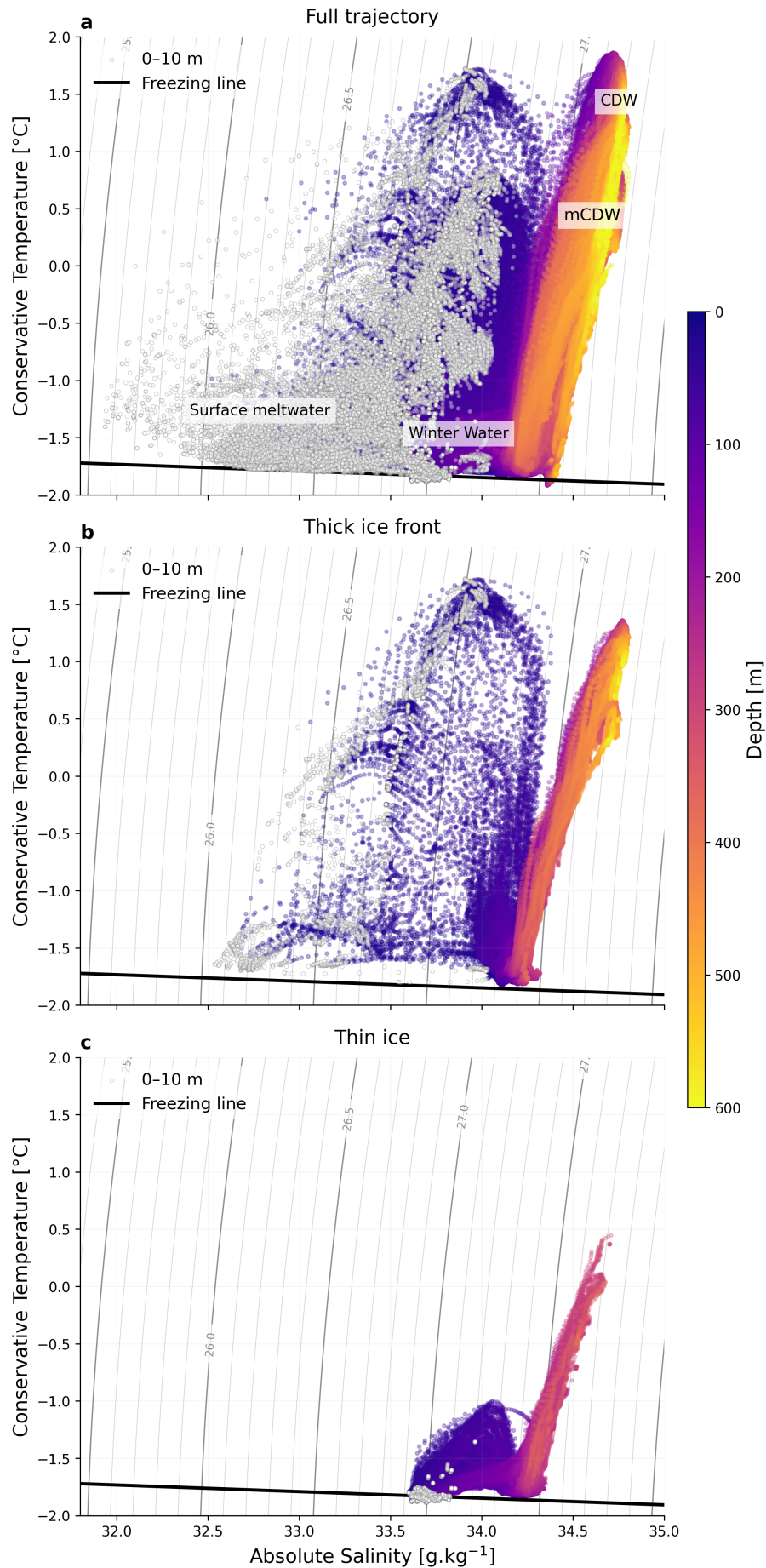


Figure 2: **TS diagrams** colored by depth for the (a) full trajectory with labeled main water masses, (b) thick ice front period (Fig. 1b, right), (c) thin ice period (Fig. 1b, left). Dots correspond to the upper 10 m. The black line denotes the freezing line.

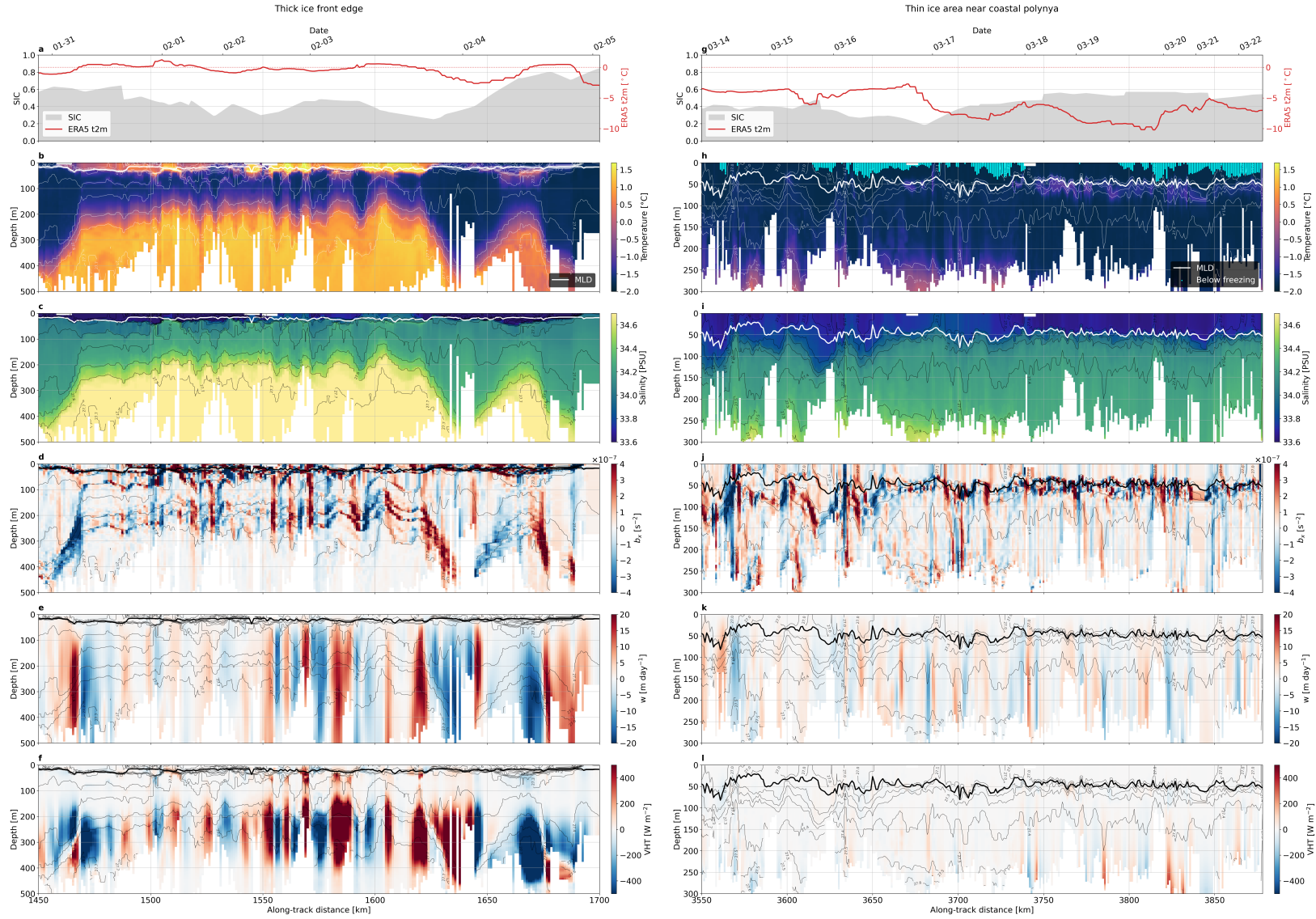


Figure 3: Along-track ocean-ice-atmosphere conditions during two contrasted regimes. Left column: thick ice-front regime (1–5 February); right column: thin ice near a coastal polynya (14–22 March). (a,g) Sea-ice concentration (SIC; gray shading) and ERA5 2 m air temperature (red). (b,h) seal-sampled temperature sections, (c,i) salinity sections. (d,j) lateral buoyancy gradient, b_x . (e,k) vertical velocity, w , (f,l) vertical heat transport (VHT). In panels (b–l), the thick black line indicates mixed-layer depth (MLD), while isopycnals (every 0.1 kg m^{-3}) are overlaid as black and white contours. Distance is along-track, increasing with time; dates are shown above the upper panels. Positive values of w and VHT correspond to upward motion and upward oceanic heat transport, respectively.

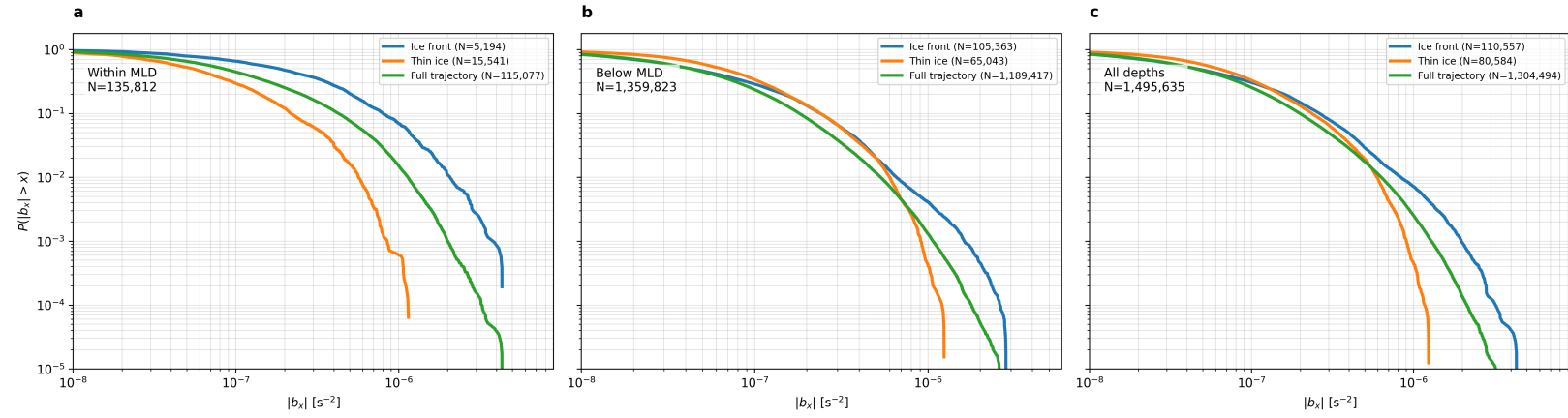


Figure 4: **Exceedance distributions of $|b_x|$** for observations (a) within the mixed layer, (b) below the mixed layer, and (c) over the full water column. Curves show $P(|b_x| > x)$ for the thick-ice-front period (blue), thin-ice period (orange), and full trajectory (green). Sample sizes are indicated in each panel and legend.

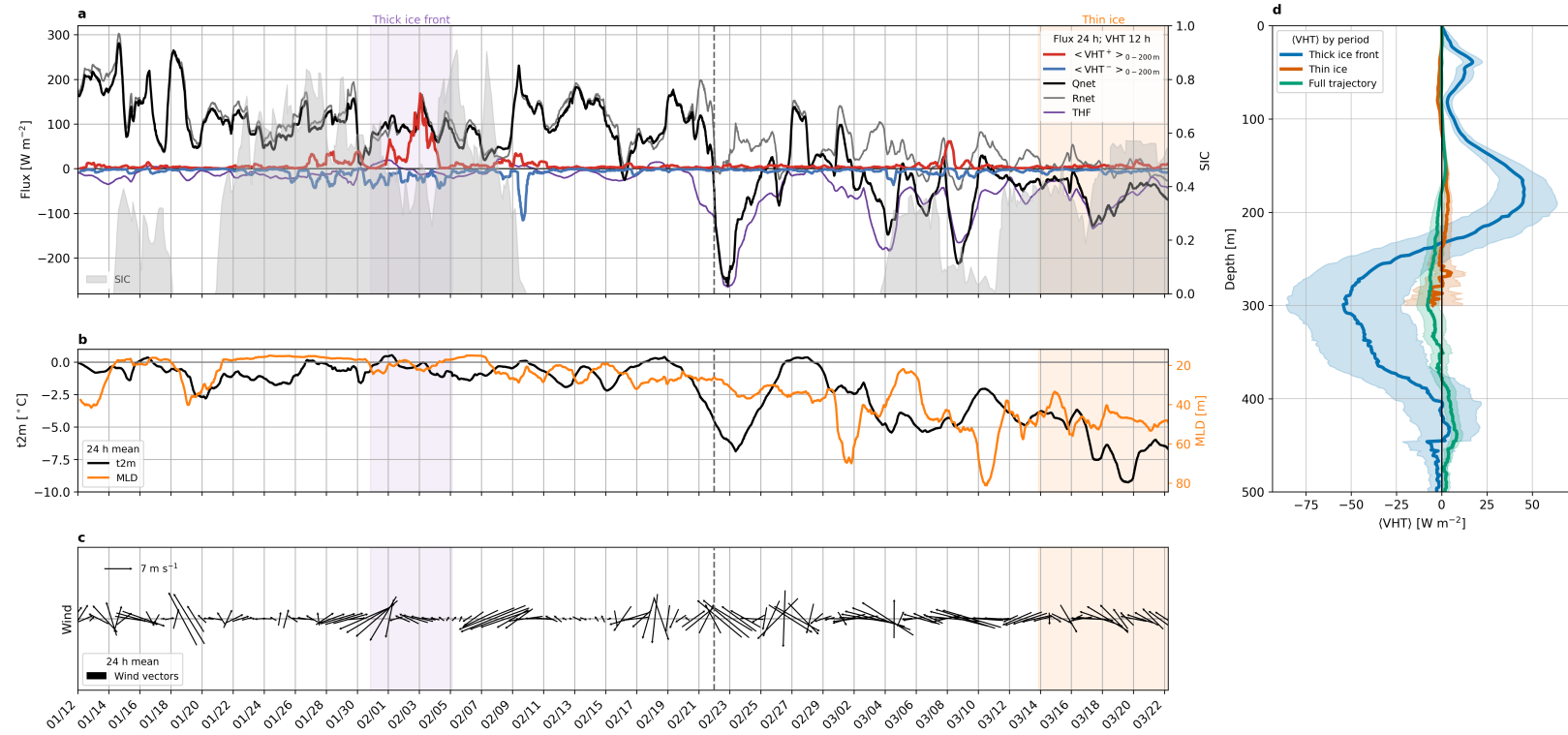


Figure 5: Surface forcing, upper-ocean heat transport, and atmospheric conditions along the full seal trajectory. (a) Surface heat fluxes from ERA5: net heat flux (Q_{net} , black), net radiative flux (R_{net} , gray), and turbulent heat flux (THF, purple). Red (blue) curves show positive (negative) vertical heat transport integrated over the upper 200 m, $\langle \text{VHT}^+ \rangle_{0-200\text{m}}$ ($\langle \text{VHT}^- \rangle_{0-200\text{m}}$). Sea-ice concentration (SIC; gray shading) is shown on the right axis. (b) ERA5 2 m air temperature ($t_{2\text{m}}$, black) and mixed-layer depth (MLD, orange; right axis). (c) 10 m wind vectors. The purple and orange shaded regions indicate the thick ice-front and thin-ice periods, respectively. All variables are shown as 24 h rolling means, except VHT, which is shown as a 12 h rolling mean. (d) Mean vertical heat transport profiles ($\langle \text{VHT} \rangle$) for the thick ice-front period (blue), thin-ice period (orange), and full trajectory (green). Shaded envelopes denote ± 1 standard error of the mean. The vertical dashed line indicated the onset of the ice-growth season.

Supplementary Material for
“Submesoscale frontal dynamics modulate ocean heat
transport beneath Antarctic sea ice”

Siegelman et al.

This PDF file includes:

Supplementary text on uncertainty propagation.

Extended Data Figures 1 to 8.

Sensitivity of diagnosed vertical velocities and heat transport to strain and seal-orientation uncertainty

To quantify the uncertainty associated with the mesoscale strain field and seal orientation, we performed a Monte Carlo ensemble analysis using the strain estimates described in the Methods sections (“Altimetry Data” and “Monte Carlo uncertainty propagation”). For each realization, strain values were randomly sampled from the observed strain distribution and combined with perturbations to the seal orientation used to project the along-track strain. The resulting strain time series are shown in Extended Data Fig. 3a, where the black line denotes the ensemble median, the gray envelope indicates the 5–95% range, and the blue line shows the original strain estimate used in the main analysis. The ensemble reproduces the temporal variability of the original strain record, while capturing the uncertainty associated with both strain magnitude and orientation. Ensemble spread generally increases during periods of elevated strain magnitude and variability, and decreases during quiescent periods when strain remains close to zero.

For each realization, the two-dimensional QG omega equation was solved to obtain vertical velocity and vertical heat transport (VHT) fields. Median profiles of root-mean-square vertical velocity (w_{rms}), root-mean-square vertical heat transport (VHT_{rms}), and mean vertical heat transport ($\langle \text{VHT} \rangle$) across the Monte Carlo ensemble are shown for the thick ice, thin ice, and full trajectory periods in Extended Data Fig. 3b–d. Shaded envelopes denote the 5–95% range of the ensemble, while dashed lines indicate the reference solution presented in the main analysis.

The thick ice regime exhibits the strongest vertical exchange. Root-mean-square vertical velocities increase from near-surface values of a few m day^{-1} to approximately 10 m day^{-1} between 200 and 350 m depth, substantially exceeding values observed during the thin-ice period (Extended Data Fig. 3b). The associated VHT_{rms} reaches $500\text{--}700 \text{ W m}^{-2}$ near 250–300 m depth (Extended Data Fig. 3c), with the 5–95% ensemble range extending to nearly

1000 W m^{-2} . In contrast, the thin-ice regime is characterized by substantially weaker vertical motions ($< 5 \text{ m day}^{-1}$) and heat transport ($< 100 \text{ W m}^{-2}$) throughout most of the water column.

Mean VHT profiles further highlight the contrast between regimes (Extended Data Fig. 3d). During the thick-ice period, positive heat transport peaks near 180–220 m depth, reaching $\sim 50\text{--}70 \text{ W m}^{-2}$, while strong negative transport occurs near 300 m depth, with ensemble-mean values approaching -50 W m^{-2} . These extrema coincide with the depth range of modified Circumpolar Deep Water and the strongest lateral buoyancy gradients. By comparison, the thin-ice period exhibits weak mean heat transport throughout the water column, generally remaining within $\pm 10 \text{ W m}^{-2}$. Profiles computed over the full trajectory fall between these two end-member regimes.

Overall, the Monte Carlo analysis demonstrates that uncertainties in the strain field and seal orientation primarily affect the magnitude of the diagnosed circulation and heat transport, while leaving the overall vertical structure and regime contrast unchanged. Across the ensemble, the thick-ice regime consistently exhibits stronger vertical velocities and substantially enhanced vertical heat transport relative to the thin-ice regime, with the largest heat-transport anomalies occurring between approximately 200 and 350 m depth. The intermittent nature of the diagnosed vertical exchange is also preserved across realizations, indicating that episodic bursts of strong vertical motion and heat transport are a robust feature of the flow rather than an artifact of a particular strain estimate or orientation assumption.

While the enhanced variability and elevated heat transport associated with the thick-ice regime are robust, the sign and magnitude of the mean VHT at depth are less certain. In particular, the negative (downward) mean VHT observed near 300 m depth during the thick-ice period may be sensitive to sampling limitations and reduced profile availability at depth. Consistent with this interpretation, the Monte Carlo 5–95% envelope spans both negative and positive val-

ues over portions of the deep water column, indicating that the sign of the mean transport is not as robustly constrained as the intermittency and enhanced magnitude of the vertical exchange. These uncertainties likely reflect a combination of limited sampling, the spatial heterogeneity of the underlying flow, and the reduced effective resolution of the seal observations at depth (ED Fig. 1).

Figure S1. Costal polynya from MODIS visible image

Figure S2. Number of profiles as a function of depth

Figure S3. Monte Carlo uncertainty analysis of strain- and orientation-driven vertical velocities and heat transport

Figure S4. 1-2nd week transect

Figure S5. 3rd-4th week transect

Figure S6. 5-6th week transect

Figure S7. 7-8th week transect

Figure S8. 9-10th week transect

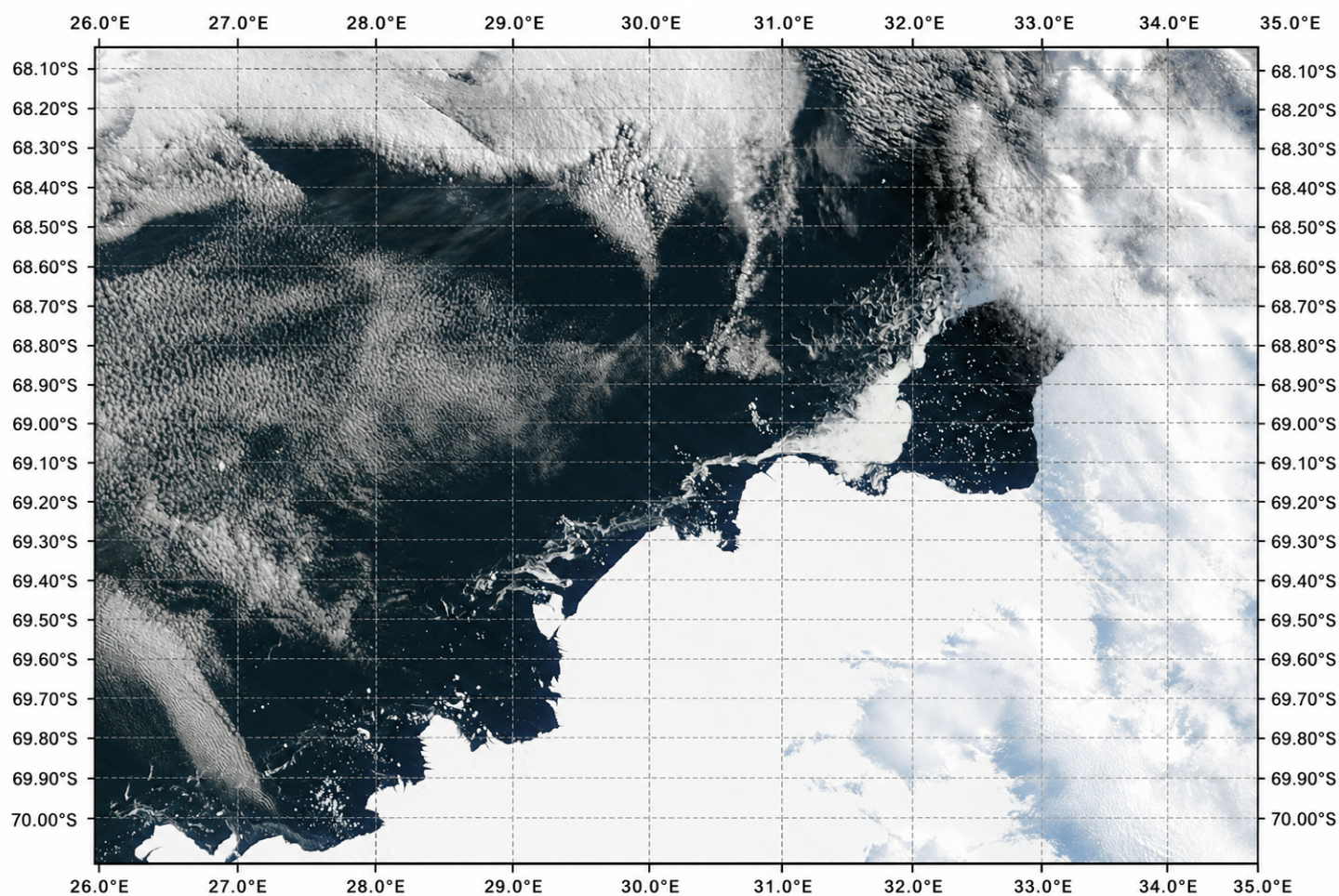


Figure 1: Visible MODIS Terra image taken on March, 4, 2024 (processed by Ocean Data Lab) highlighting the presence a coastal polynya at $\sim 68.8 - 69.2^{\circ}\text{S} - 32 - 33^{\circ}\text{E}$, just south of the seal's trajectory as seen in Figure 1.

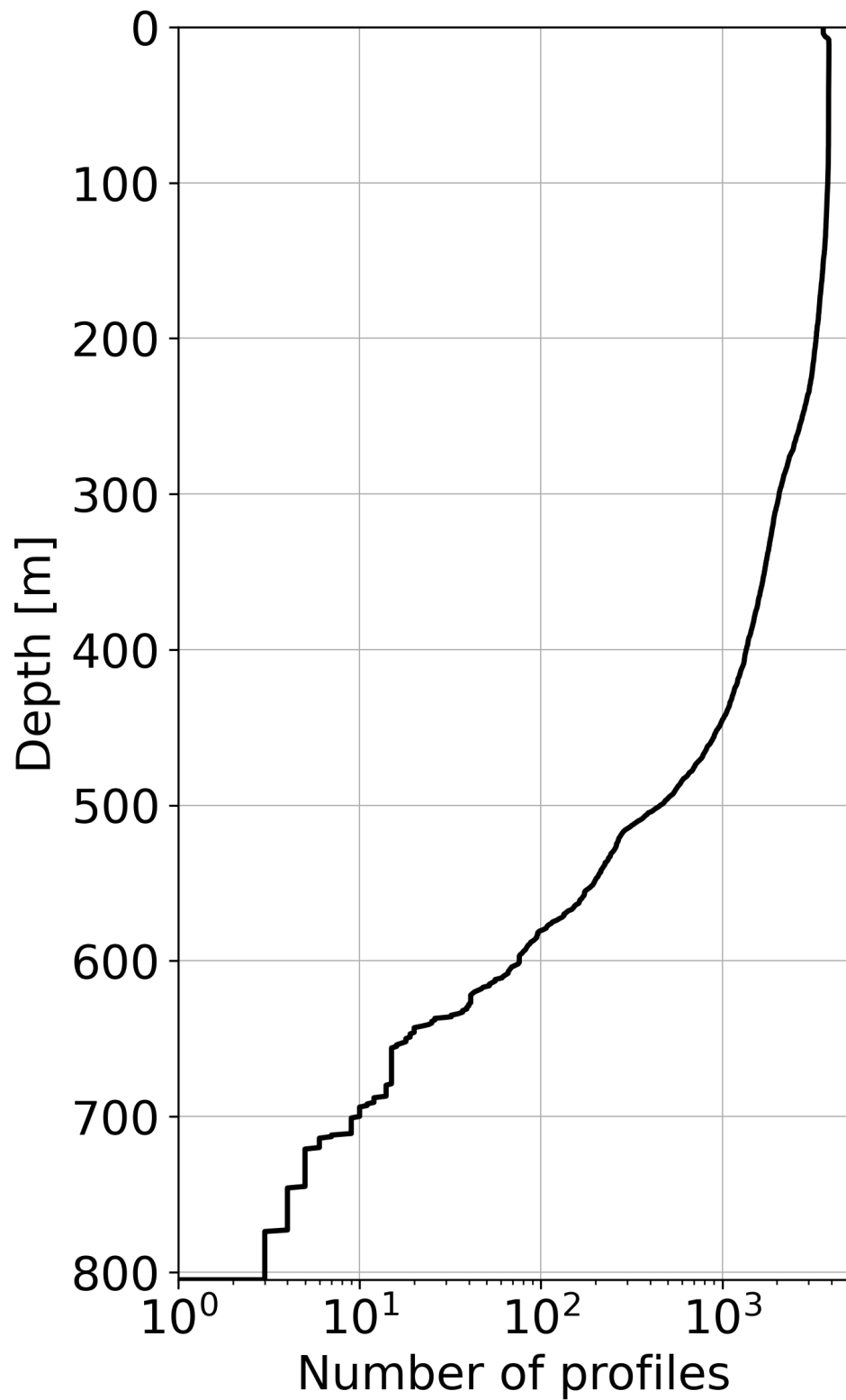


Figure 2: Number of available seal hydrographic profiles as a function of depth along the analyzed trajectory. The number of profiles decreases rapidly below ~ 500 m depth, reflecting the limited deep sampling coverage of the dataset.

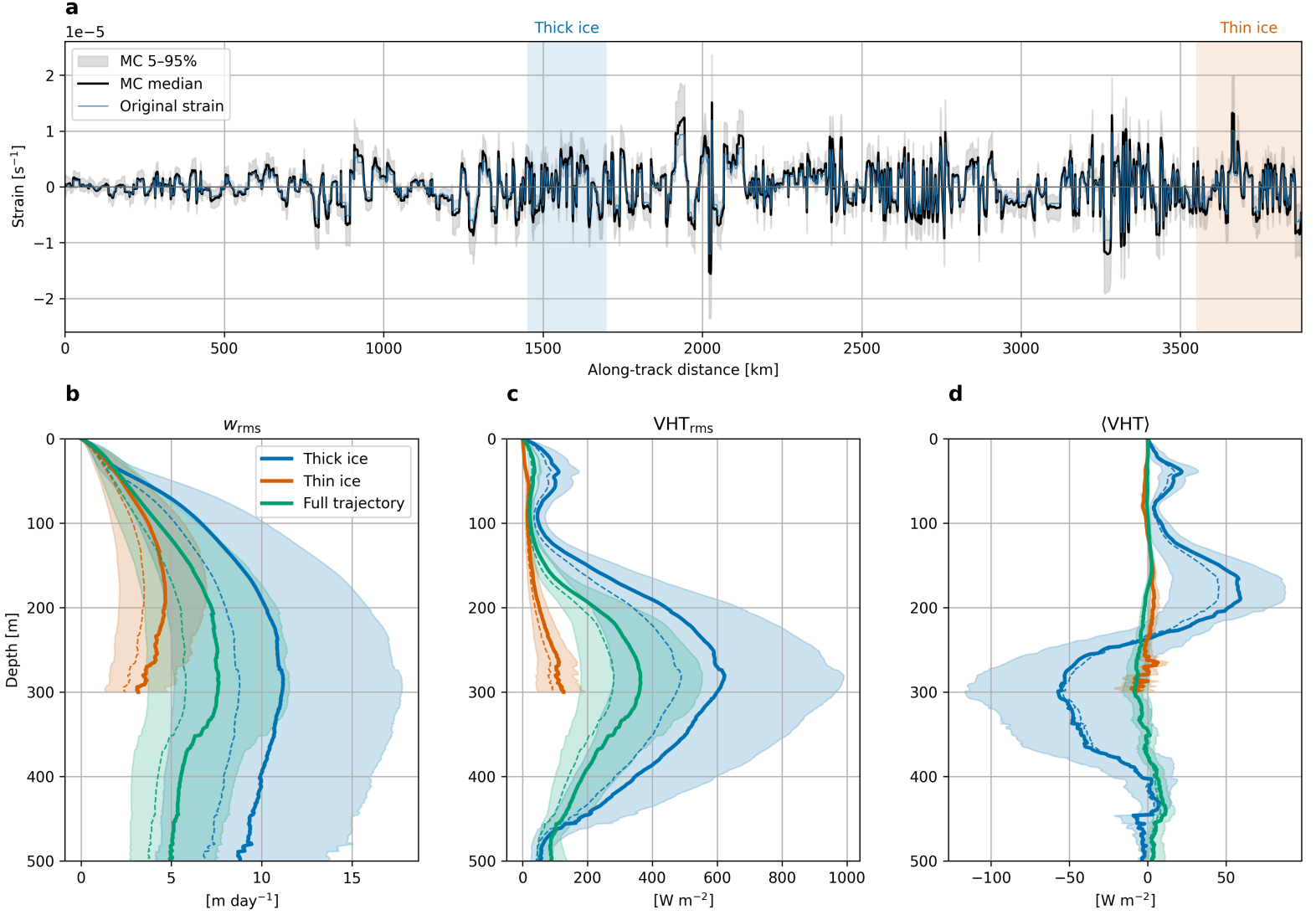


Figure 3: (a) Monte Carlo realizations of the along-track strain field used to force the QG omega equation. The black line shows the ensemble median, the gray shading denotes the 5–95% range, and the blue line indicates the reference strain estimate used in the main analysis. Blue and orange shaded regions mark the thick-ice and thin-ice periods, respectively. (b) Root-mean-square vertical velocity (w_{rms}), (c) root-mean-square vertical heat transport (VHT_{rms}), and (d) mean vertical heat transport ($\langle VHT \rangle$) as a function of depth for the thick-ice period (blue), thin-ice period (orange), and full trajectory (green). Solid lines show the median across all Monte Carlo realizations, shaded envelopes denote the 5–95% range, and dashed lines indicate the reference solution presented in the main text.

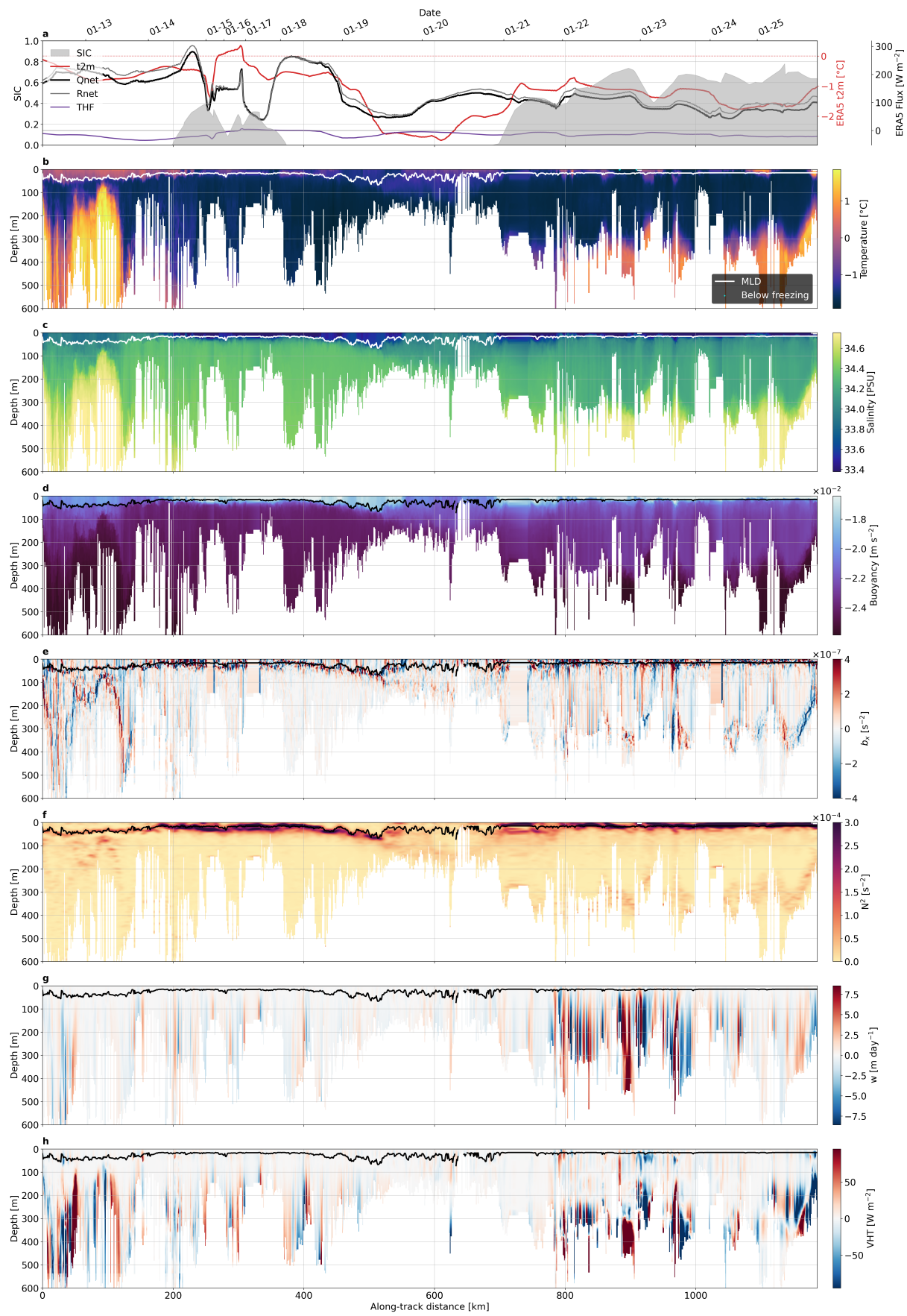


Figure 4: Along-track evolution of atmospheric, hydrographic, and dynamical conditions during the seal trajectory from 12–25 January 2024. (a) sea-ice concentration (SIC), ERA5 2-m air temperature, ERA5 surface heat fluxes (Q_{net} , R_{net} and THF). Vertical sections of (b–d) temperature, salinity, and buoyancy, respectively, with the mixed layer depth indicated by the black contour, (e–f) horizontal buoyancy gradients (b_x) and Brunt–Väisälä frequency squared (N^2), (g–h) vertical velocity (w) and associated vertical heat transport (VHT).

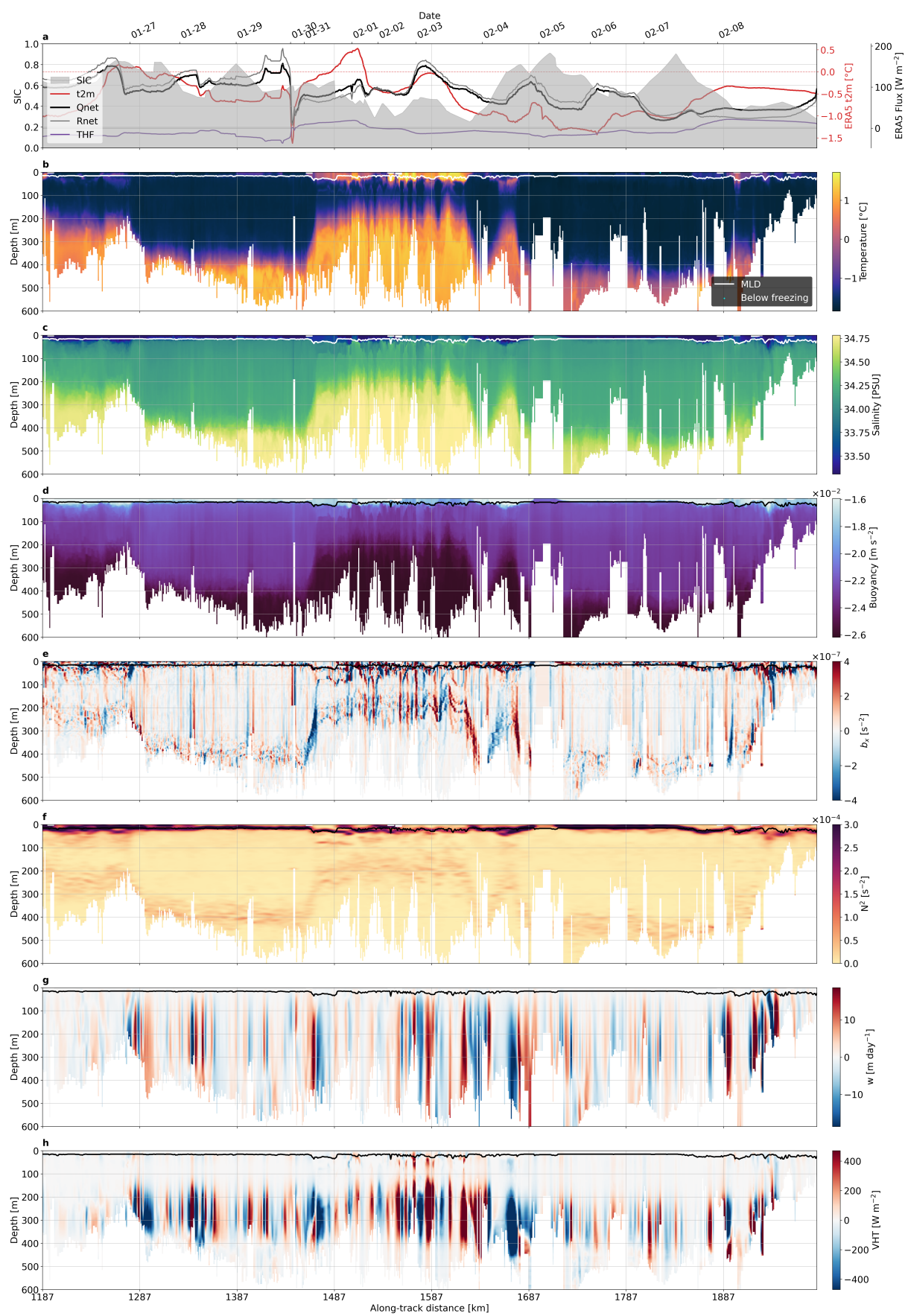


Figure 5: Same as ED Fig. 4 during the sea ice trajectory from 25 January – 9 February 2024.

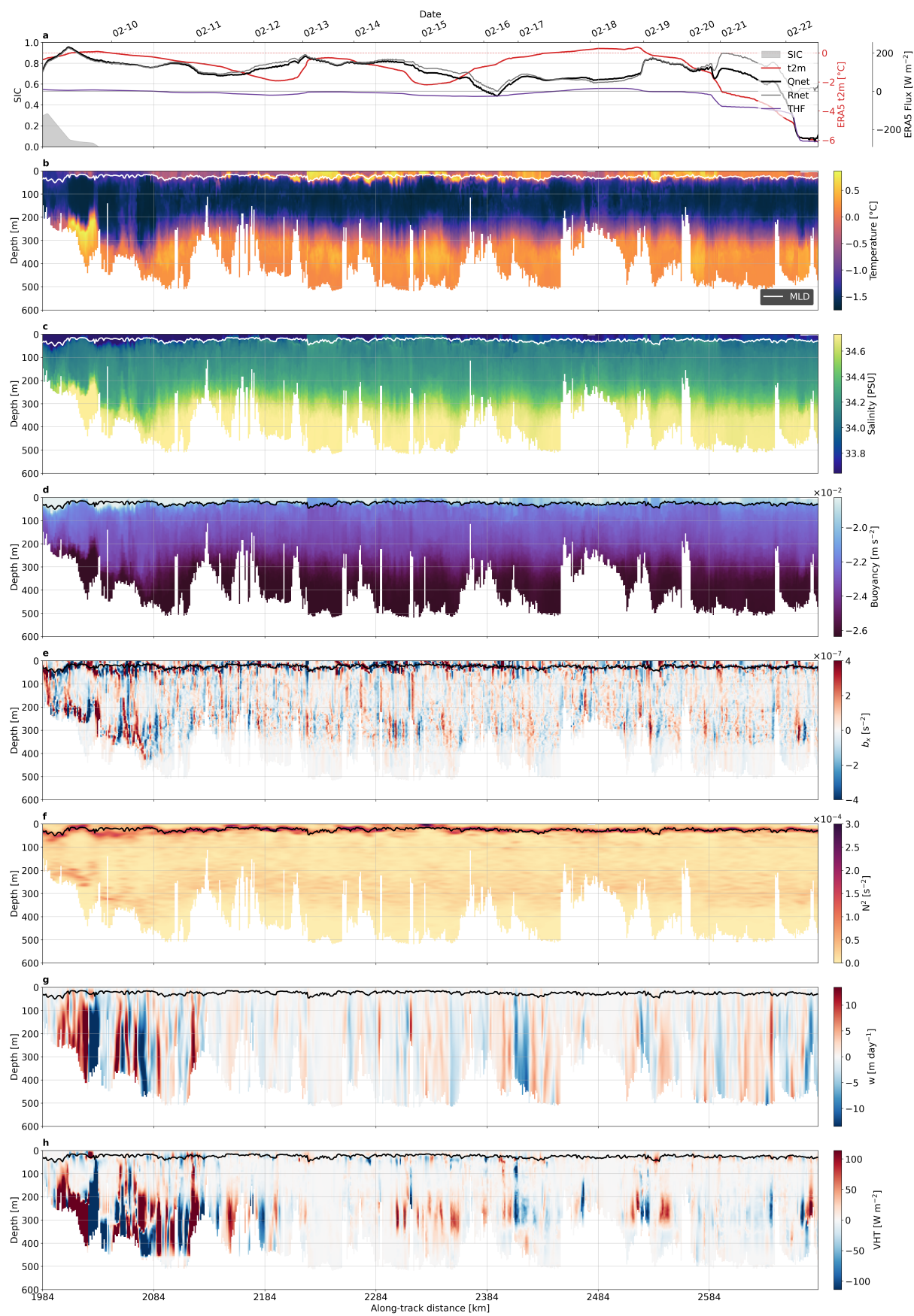


Figure 6: Same as ED Fig. 4 during the seal trajectory from 9–23 February 2024.

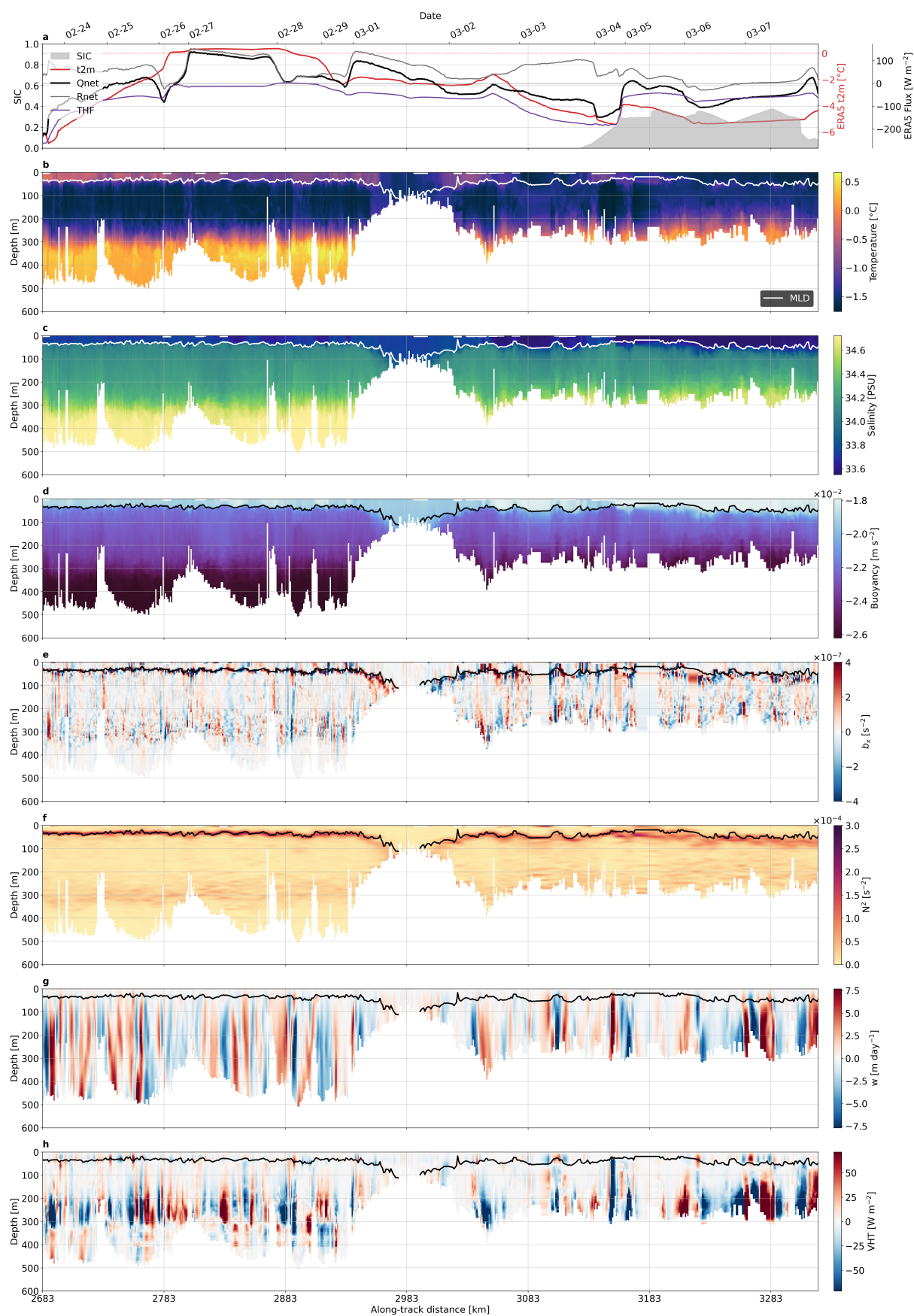


Figure 7: Same as ED Fig. 4 during the sea trajectory from 23 February – 8 March 2024.

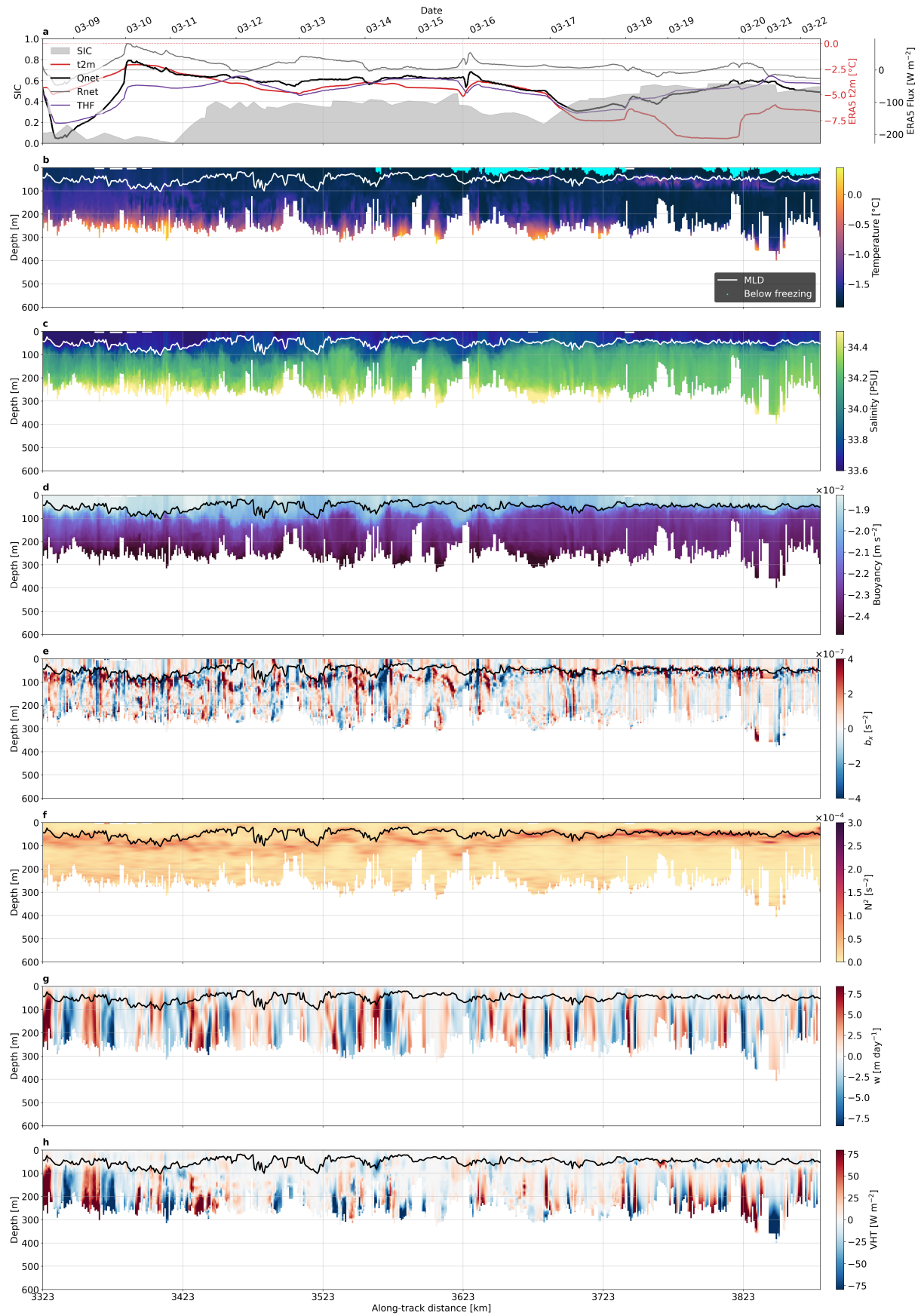


Figure 8: Same as ED Fig. 4 during the seal trajectory from 8–22 March 2024.