

1 **Thwaites Eastern Ice Shelf Cavity Observations Reveal Multi-year**
2 **Sea Ice Dynamics and Deep-Water Warming in Pine Island Bay,**
3 **West Antarctica**

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17 **Abstract.**

18 Pine Island Bay (PIB), situated in the Amundsen Sea, is renowned for its retreating ice shelves and highly variable sea ice.
19 While brine rejection from sea ice formation and glacial meltwater influence seawater properties, the downstream impacts
20 beneath the region’s floating ice shelves remain poorly understood. Here, we exploit an unprecedented, multi-year (2020–
21 2023) oceanographic time series from instruments deployed through boreholes beneath the Thwaites Eastern Ice Shelf
22 (TEIS), immediately downstream of PIB, offering new insight into how ice-ocean-atmosphere interactions in PIB shape
23 oceanographic conditions within the subshelf cavity. Our observations reveal a sustained warming and thickening of the
24 modified Circumpolar Deep Water (mCDW) layer near the seabed since January 2020, critical in a region where mCDW
25 drives basal melting beneath West Antarctica’s most vulnerable outlet glaciers. Concurrently, the retreat of the multi-year
26 sea ice edge by over 150 km across most of PIB has enhanced the advection of Winter Water, contributing to a cooling of
27 more than 1°C in the upper 250 m beneath TEIS between July 2021 and January 2023. Superimposed on these trends are
28 episodic temperature and salinity anomalies lasting several weeks, originating in PIB and advecting past the moorings. These
29 events link mobile sea ice cover to subshelf hydrography, as mid-depth waters temporarily warm and increase in salinity,
30 leading to an increase in density, while deeper mCDW simultaneously cools and freshens, reducing its density. Overall,
31 these changes are associated with reduced stratification in the cavity. As sea ice continues to decline in a warming Antarctic
32 climate, our results offer a glimpse into how ocean circulation and basal melting may evolve across the Amundsen Sea
33 Embayment. This dataset provides a critical benchmark for refining process-based models and improving melt-rate
34 parametrizations in coupled ice-ocean simulations.

35 **1 Introduction**

36 Ice shelves encircle much of Antarctica, acting as critical buffers that slow the flow of continental ice into the ocean (Fürst et
37 al., 2016). However, many ice shelves have thinned or even collapsed in recent decades (Doake and Vaughan, 1991; Rack
38 and Rott, 2004; Scambos et al., 2004; Lhermitte et al., 2023), triggering rapid acceleration of grounded ice (Rignot et al.,
39 2004; Scambos et al., 2014). This process is particularly concerning in the Amundsen Sea Embayment, where Pine Island
40 and Thwaites glaciers could together contribute 1.16 m to global sea-level rise if marine ice-sheet instability takes hold
41 (Schoof, 2007; Joughin et al., 2014; Rignot et al., 2019; Gudmundsson et al., 2023; Morlighem et al., 2024). Thwaites
42 Glacier has become a focal point in climate research (Scambos et al., 2017) due to its rapid retreat (Rignot et al., 2019;
43 Milillo et al., 2019; Wild et al., 2022; Rignot et al., 2024) and the on-going deterioration of its last remaining ice shelf (Alley
44 et al., 2021; Wild et al., 2024), largely driven by the intrusion of modified Circumpolar Deep Water (mCDW; Dutrieux et al.,
45 2014; Christianson et al., 2016; Jenkins et al., 2018; Nakayama et al., 2019). However, sub ice-shelf cavities remain among
46 Earth's least explored regions, and limited observational data hinder our ability to model the intricate interplay between
47 oceanic warming, ice-shelf stability, grounding-zone processes, and the fate of Thwaites Glacier (Seroussi et al., 2017; Yu et
48 al., 2018; Holland et al., 2023).

49 Circumpolar Deep Water accesses the continental shelf through deep glacially carved troughs (Heywood et al., 2016).
50 It gradually cools and freshens as it moves southward, following narrow bathymetric pathways (10–20 km wide) and mixing
51 with on-shelf water masses before intruding into the deeper cavities beneath ice shelves and glacier fronts (Nakayama et al.,
52 2019). By the time it reaches Pine Island Bay (PIB), mCDW ($>0\text{ }^{\circ}\text{C}$, $>34.7\text{ g kg}^{-1}$) remains 2–4 $^{\circ}\text{C}$ above the in-situ freezing
53 point, supplying substantial thermal energy for basal melting. The Thwaites Trough extends from the north, reaching depths
54 of ~1300 m and splitting into three narrower branches west of the pinning point buttressing Thwaites Eastern Ice Shelf
55 (TEIS), while the adjacent Pine Island Bay Trough, slightly deeper (~1400 m), extends beneath TEIS from the east but is
56 thought to be constrained by a bathymetric sill (Fig. 1b). Autonomous underwater vehicle (AUV) surveys indicate that
57 mCDW enters the TEIS cavity predominantly through the easternmost branch near its pinning point (T3), with meltwater-
58 enriched waters exiting through the westernmost branch (T2, Fig. 1b; Wåhlin et al., 2021). Notably, hydrographic signatures
59 from PIB have been detected near the pinning point (Wåhlin et al., 2021), suggesting mixing between these two competing
60 water masses at depth and an extensive westward influence of PIB circulation (Seroussi et al., 2017; Nakayama et al., 2019).

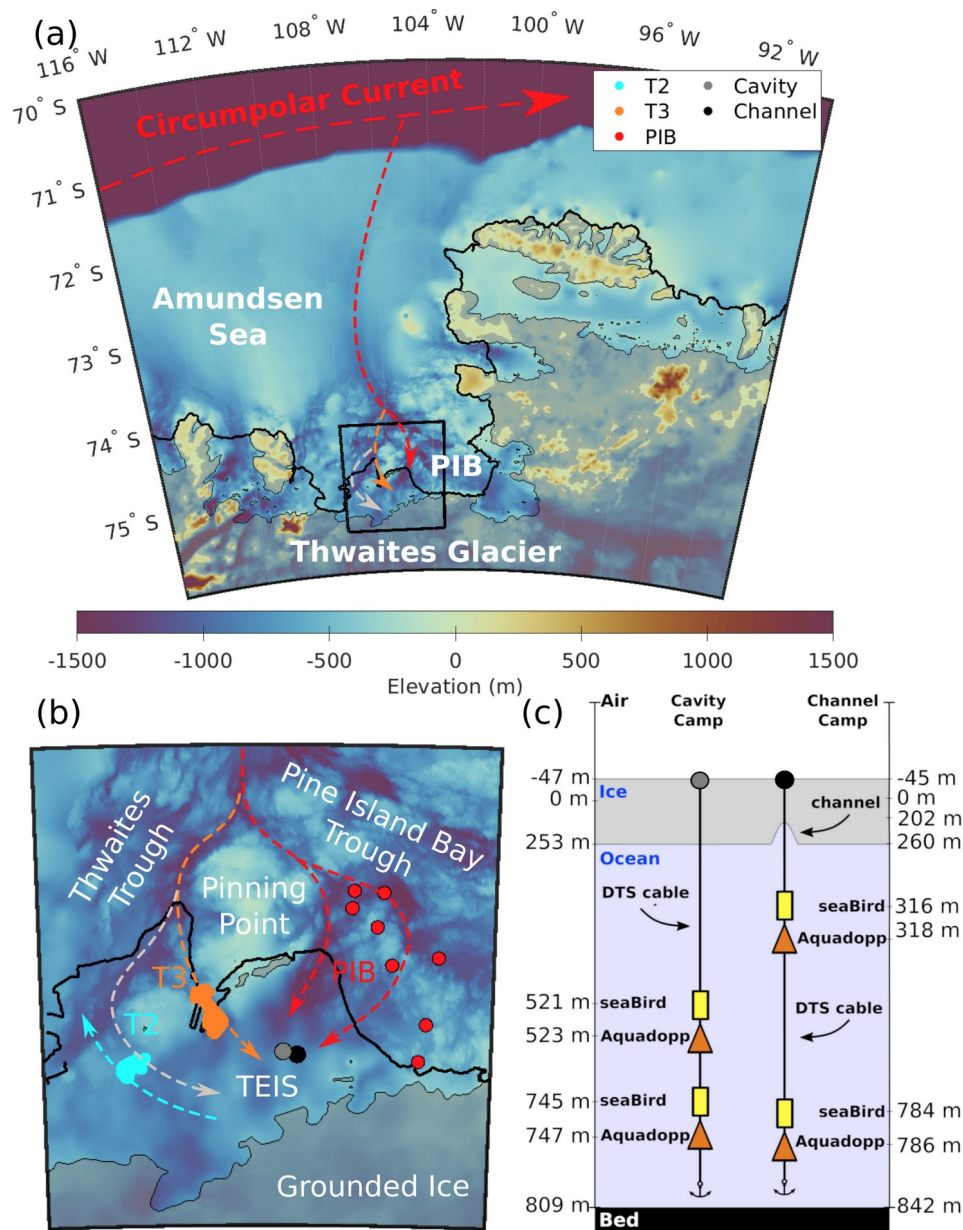


Figure 1: (a) Bathymetric map showing water pathways into Pine Island Bay (PIB). (b) Location of Cavity Camp and Channel Camp on Thwaites Eastern Ice Shelf (TEIS) and the location of its pinning point. Red dots indicate locations of ship-based Conductivity-Temperature-Depth (CTD) measurements in February 2019 capturing PIB water masses, while light blue and orange dots represent AUV measurements in the bathymetric troughs T2 and T3, respectively, which branch from the Thwaites Trough (Wählin et al., 2021). (c) Illustration presenting a cross-sectional view of an idealized ice-shelf featuring a basal channel, showing the positions of Cavity Camp and Channel Camp, the two Distributed Temperature Sensing (DTS) cables, MicroCATs, and Aquadopp instrument pairs deployed in the subshelf ocean cavity.

70 Observational studies have demonstrated that subshelf oceanography is strongly influenced by neighbouring ocean
71 conditions (Webber et al., 2017; Davis et al., 2018, Zheng et al., 2022, Dotto et al., 2022, Davis et al., 2023). AUV and ship-
72 based Conductivity-Temperature-Depth (CTD) surveys have revealed competing mCDW sources beneath TEIS, originating
73 from both PIB and Thwaites Trough (Wåhlin et al., 2021). In PIB, surface circulation is dominated by a gyre system—a
74 rotating ocean circulation shaped by regional wind forcing, bathymetry, and glacial meltwater fluxes (Thurnherr et al., 2014;
75 Heywood et al., 2016; Yoon et al., 2022). Its strength and sense of rotation can be altered by the concentration and mobility
76 of landfast sea ice—stationary, often multi-year sea ice anchored to the coastline (hereafter, ‘fast ice’) that eventually forms
77 a stable, immobile platform that isolates the ocean from atmospheric wind stress (Zheng et al., 2022). Extended periods of
78 fast ice coverage promote weakening of the PIB gyre leading to an accumulation of glacial meltwater (i.e., a relatively
79 warmer water derived from mCDW melting the ice base) near the surface, which leads to shallower isopycnals beneath the
80 neighbouring TEIS and thus to warmer conditions at the TEIS base (Dotto et al., 2022). In contrast, fast ice breakouts
81 combined with a cyclonic PIB gyre enhance the intrusion of cooler surface waters into the subshelf cavity (Dotto et al.,
82 2022), potentially explaining the suppressed basal melt beneath the ice shelf (Wild et al., 2024).

83 Previous studies have provided valuable insights into the relationship between sea ice and ocean conditions, but
84 they have been limited in their spatial and temporal scope, restricting our understanding of multi-year variability. In
85 particular, while different sea ice types are known to modulate ocean surface stress and gyre dynamics (Zheng et al., 2022),
86 the implications for heat transport toward ice shelves remain poorly constrained (St-Laurent et al., 2015). The vertical extent
87 of warmer water within subshelf cavities under prolonged fast ice cover, as suggested by Dotto et al. (2022), also remains
88 unknown. Here we build on the ideas of Dotto et al. (2022) by extending the observational record from January 2020–March
89 2021 to January 2020–January 2023, allowing us to capture interannual changes in ocean conditions beneath the Thwaites
90 Eastern Ice Shelf (TEIS). We specifically investigate how transitions between thin, mobile first-year sea ice and thick,
91 immobile multi-year fast ice influence the water column beneath TEIS. Additionally, we assess how the competing water
92 masses from PIB and Thwaites Trough respond to the persistence and extent of multi-year fast ice.

93 The paper is organized as follows: First, we present the dataset and analyze the temporal variability of hydrographic
94 properties at shallow, mid-depth, and deep water layers. Next, we compare our measurements beneath TEIS with published
95 datasets from nearby ship-based surveys. We then examine the temporal co-variability of our expanded dataset, revealing a
96 progressive warming of the mCDW layer at depth, periodically disrupted by distinct events lasting a few weeks in which the
97 mCDW temporarily cools and freshens, while mid-depth waters become denser. Using Distributed Temperature Sensing
98 (DTS) profiles, we assess the vertical extent of these events throughout the water column. Finally, we analyze remotely
99 sensed sea ice cover in PIB, identifying that events align with first-year sea ice formation that persist until May 2021. After
100 this period, the upper water column undergoes substantial cooling, likely driven by the gradual retreat of multi-year fast ice

101 in PIB. This retreat enhances Winter Water (WW) formation through air-sea fluxes (Webber et al., 2017), promoting the
102 intrusion of WW beneath the adjacent TEIS.

103 **2 Data and methods**

104 **2.1 Observations and processing**

105 In December 2019, we established two hot water drilling camps on TEIS to access its underlying ice-shelf cavity: Cavity
106 Camp, situated centrally above the ocean cavity beneath the ice, and about 4 km eastward Channel Camp, positioned at the
107 apex of an ice-shelf basal channel (Fig. 1; Dotto et al., 2022; Scambos et al., 2025). We present atmospheric and
108 hydrographic measurements of both sites collected between January 2020 and January 2023 by two automated stations
109 (Automated Meteorology-Ice-Geophysics Observing Systems - 3, or AMIGOS-3; Scambos et al., 2025). These on-ice
110 mooring systems incorporated instruments on the ice-shelf surface (e.g., air temperature, wind, and pressure sensors), and
111 DTS fibre optic systems drilled through the ice shelf and the entire water column beneath to capture ice and ocean
112 temperature profiles. Each AMIGOS-3 station also included an under-ice mooring with a suite of ocean instruments attached
113 (described in detail below), including a set of MicroCAT instruments for measuring ocean conductivity, temperature, and
114 pressure, each paired with Aquadopp current meter instruments (Fig. 1c).

115 **2.1.1 Borehole CTD cast**

116 On January 12, 2020, hot water drilling activities were conducted at Channel Camp, followed by the collection of an initial
117 CTD profile down to the seabed at a depth of 842 m. This initial CTD cast was used to establish the relationship between
118 temperature, salinity, and ambient pressure within the ocean cavity (Appendix A). To focus on long-term averages we
119 excluded the depth range of the thermocline, between 270 m and 425 m, and fitted a second-order polynomial function to the
120 remaining CTD measurements.

121 **2.1.2 MicroCAT CTDs**

122 Four Sea-Bird MicroCAT SBE 37-IMP instruments were employed at fixed depths to monitor temporal variability of
123 conductivity, temperature, and ambient pressure in three distinct water layers. One was positioned at an initial depth of 316
124 m (referred to as the “shallow” MicroCAT), while a second one was positioned at 521 m (“mid-depth” MicroCAT), and two
125 other sensors were positioned at 745 m and 784 m depth (“deep” MicroCATs) beneath the ocean surface (Fig. 1c). We
126 conducted cross-calibration of these instruments in the circulating seawater tanks at McMurdo Station.

127 Following two years of uninterrupted recording at a temporal resolution of 10 minutes, the shallow MicroCAT
128 instrument stopped functioning in January 2022. The mid-depth and both deep MicroCAT instruments remained operational
129 for an additional year until January 2023, when the dataset was retrieved from the instruments. Conservative temperature (θ ;

130 °C), absolute salinity (S_A ; g kg^{-1}) and potential density referenced to zero pressure from each instrument were computed
131 using the Thermodynamic Equations of Seawater-10 (McDougall et al., 2011). We then used a Chebyshev low-pass filter
132 with a 1 hour cutoff frequency to filter these records for outliers and calculated depth below the ocean surface from the
133 filtered in-situ pressure measurements.

134 2.1.3 DTS thermal profiling

135 DTS temperature profiles through the ocean column were used as a proxy for hydrographic variability at different depths and
136 over varying time scales. A DTS laser interrogator system (Silixa XT, Silixa LTD, Hertfordshire UK) was attached to an
137 armored multi-strand, fibre-optic cable (FIMT) connected to the primary steel cable holding the ocean instruments (Scambos
138 et al., 2025). This setup enabled the collection of temperature profiles with a vertical sampling of 25 centimeters resulting in
139 an approximate spatial resolution of 50 cm (Tyler et al., 2009). DTS measurements were integrated over 1 minute with
140 estimated temperature resolution of 0.033 °C and 0.038 °C at the deepest measurement for Cavity and Channel Camp
141 mooring, respectively. The temperature resolution is estimated by calculating the variance of DTS-derived temperatures
142 within a 2.5 m section near the bottom of each mooring. The 2.5 m sections were centered at 730 m for Cavity Camp and
143 750 m for Channel Camp, deep in the profile where no vertical gradients would be measurable over the 2.5 m section.

144 DTS measurements at both stations were generally captured every 4 hours during the austral spring to early-autumn
145 (October-April), but were extended to 24-hour intervals from mid-autumn through winter (May-September) to conserve
146 power. At Channel Camp, DTS data were acquired from January 2020 to August 2021. In January 2023, we gathered
147 additional DTS data at Channel Camp, with recordings every ~90 seconds over a duration of 2 hours and 45 minutes (UTC
148 Start: January 8, 2023 21:31:37; End: January 9, 2023 00:15:53). Subsequently, these 154 individual DTS profiles from that
149 short period were averaged to create a consolidated DTS profile for January 2023. At Cavity Camp, the DTS data record
150 spans January 2020 until October 2021.

151 We calibrated the DTS data using the MicroCAT instruments, which sampled the water column during the DTS
152 measurements. For most of the record, we applied a straightforward two-point calibration (slope and offset) to each DTS
153 trace. In 2023, when only the deep MicroCAT instrument was operational at Channel Camp, we performed a three-point
154 calibration using an assumed constant minimum ice temperature from the middle of the ice shelf layer and the pressure
155 melting point at the ice shelf-ocean interface. In both cases, we used a single-ended calibration method. Calibrating the DTS
156 with MicroCATs effectively corrects for temporal drift in the system, improving temperature estimates across the full depth
157 of the water column. The calibrated DTS data were then binned into daily bins.

158 After calibration, we used the relationship between in-situ temperature and salinity from the initial CTD cast to
159 calculate Θ profiles based on the 'proxy salinity profiles'. This approach assumes that the proxy salinity profile derived on
160 January 12, 2020, remains representative throughout the three years of DTS data collection. To validate this assumption, we
161 compared it against a time series of in-situ temperature, salinity, and pressure from two MicroCATs. We calculated Θ in two
162 ways: (1) using the salinity time series and (2) using a constant salinity from the initial measurement. The differences

163 between these two methods were negligible (RMSE of 0.0002 °C for the shallow MicroCAT and 0.001 °C for the deep
164 MicroCAT at Channel Camp, compared to mean values of -0.88 °C and 1.05 °C, respectively). Based on these results and
165 the lack of other measurements, we assume a constant salinity profile to derive seawater density profiles, allowing us to
166 assess the net effect of in-situ temperature changes on mean water-column density (Appendix A). A caveat of this
167 assumption is that this approach primarily captures warm/salty and cold/fresh water masses, and does not account for the
168 warm/fresh combination typical for glacial meltwater in this region.

169 **2.1.4 Aquadopp current meters**

170 Nortek Aquadopp current meters were installed two meters below each MicroCAT, capturing current velocities to determine
171 the ocean circulation patterns related to the water characteristics captured by the CTD and DTS systems (Fig. 1c). Ocean
172 current data were acquired hourly with a data gap between August 10-28, 2020, for the Channel mooring, and May 29 to
173 August 28, 2020, for the Cavity mooring, owing to low station power. The velocity components measured by the Aquadopps
174 were corrected for the magnetic declination, 50.07°E. The Aquadopp records were later binned into daily data chunks for
175 visualization to show temporal variability of ocean current speed and direction.

176 **2.1.5 Atmospheric dataset**

177 We used wind speed and direction measurements to determine the prevailing atmospheric circulation that may impact ice
178 and ocean processes near TEIS. The AMIGOS-3 were equipped with a multi-parameter Vaisala 530 series weather sensor,
179 which acquired hourly air temperature, wind speed and direction at 7 to 3 m above the surface of the ice shelf (as
180 accumulation slowly buried the AMIGOS-3 tower). Here we focussed on the atmospheric data record from Channel Camp as
181 the difference in atmospheric variability from Cavity Camp is negligible within the context of this study, and the Channel
182 Camp data record is slightly longer (Scambos et al., 2025). Given the potential influence of atmospheric winds on upper
183 ocean circulation patterns, we compared the wind data with the variability observed in ocean sensors measuring current
184 speed and direction. For this comparison we relied on ERA5 reanalysis on single levels (Hersbach et al., 2020) because of
185 temporal gaps in our wind record (April 19–May 19, 2020; June 30–July 23, 2020; and August 8–September 11, 2020).
186 From ERA5's $0.25^\circ \times 0.25^\circ$ spatial resolution, we selected and averaged three grid points (Latitude: -75°, Longitudes: -
187 105.76°, -105.51°, and -105.26°) to obtain a representative dataset for the TEIS region. We used ERA5's native hourly
188 resolution for wind speed, wind direction, and 2 m temperature. The validity of ERA5 was assessed by comparing it with our
189 wind measurements during periods when observations were available (Appendix B).

190 To calculate daily mean wind and current directions and speeds, we first converted the directional data into
191 eastward and northward vector components. These components were then averaged by day to avoid errors associated with
192 circular averaging (e.g., averaging 1° and 359°). The daily mean direction was reconstructed from the averaged components
193 using the arctangent of the northward and eastward means, and the mean speed was calculated from their Euclidean norm.

194 2.2 Monitoring sea ice variability remotely

195 2.2.1 Satellite SAR data from Sentinel-1A

196 Since water circulation beneath TEIS is likely to be impacted by regional sea ice coverage (Dotto et al., 2022), we used
197 publicly available satellite radar imagery from the Sentinel-1A operating at C-band (5.4 GHz/5.6 cm) to monitor sea ice
198 variability in PIB. This active microwave sensor has captured synthetic aperture radar (SAR) images every 12 days over PIB
199 since 2014, having the advantage of being able to continuously observe the surface in polar night and through cloud cover,
200 unlike optical imaging systems. We used the extra wide swath mode product with single HH (i.e., horizontally transmitted
201 and horizontally received radar signals) polarization, covering a broad 400 km area at a medium ground resolution of 20 m
202 by 40 m. Using these images, we compiled a video illustrating the regional evolution of sea ice in PIB (Supplementary
203 Video).

204 2.2.2 Sea ice concentration time series

205 We complement the SAR data snapshots with a more complete, but lower spatial resolution, time series of daily sea ice
206 concentration provided by the University of Bremen's sea ice data center (Spreen et al., 2008). We used the Antarctic daily
207 product (asi_daygrid_swath) with no land mask applied and processed to 3.125 km grid spacing
208 (Antarctic3125NoLandMask). We apply the Norwegian Polar Institute Quantarctica 3 Basemap
209 (ADD_Coastline_high_res_polygon_Sliced) land and ice shelf masks around PIB to retrieve only concentrations over open
210 ocean and calculate the daily mean sea ice concentration (%) across the PIB sea ice sampling box (102° - 106° W, 74.5° -
211 75.0° S) from January 2020 to January 2023.

212 2.3 Wavelet analysis

213 We employed wavelet transforms on the hydrographic records to uncover any systematic patterns in their temporal
214 variability at different depths and to differentiate scales of forcing. This was carried out using the MATLAB package
215 developed by Grinsted et al. (2004) using the Morlet wavelet. Unlike traditional harmonic analysis integrating signals over
216 time, wavelet analysis has the advantage of identifying changes in power over time for a specific period.

217 The continuous wavelet transform of a single time series decomposes the signal into time-frequency space, allowing
218 the identification of localized oscillatory behavior at different periods. The wavelet coefficients retain the units of the
219 original signal, and the wavelet power, computed as the squared magnitude of the coefficients, represents the localized
220 variance. To aid interpretation, we normalize the power by the total variance of the time series, producing a dimensionless
221 quantity, and visualize the logarithm (base 2) of this normalized power. This highlights both dominant periodicities and their
222 temporal evolution, with colour representing the relative strength of variability at each period. Thus, we resolve intermittent
223 signals from hourly periods to longer-period ones, spanning up to several months.

224 Furthermore, we applied the cross-wavelet transform to examine the common power and phase relationships
225 between pairs of time series, for example density variations and environmental drivers such as wind and ocean currents. The
226 cross-wavelet transform highlights regions in time and frequency space where the two time series exhibit high covariance,
227 allowing for the identification of temporally-localized, period-dependent coupling. The resulting cross-wavelet power is
228 dimensionless and plotted on a logarithmic (base 2) scale, with arrows indicating relative phase between the two time series,
229 where the arrow direction indicates if one time series leads the other at that specific period or if they occur harmonically in
230 phase. This provides insight into both the strength and timing of shared variability between the signals.

231 The statistical significance of the identified periodicities in covariance for both the continuous- and the cross-
232 wavelet transform was determined using standard Monte-Carlo methods against red noise background (see Grinsted et al.,
233 2004). Before computing wavelet transforms, we linearly interpolated the data onto evenly-spaced temporal resolution
234 increments of 10 minutes, applied a Chebyshev low-pass filter to eliminate any outliers and detrended the time series. The
235 cut-off period of the Chebyshev filter consequently sets the minimum signal that can be resolved with the wavelet transform.
236 Given that the Amundsen Sea exhibits a diurnal tidal regime, we applied a cut-off period of 0.125 days (or 3 hours) for the
237 Chebyshev filtering to resolve the tidal variability in our datasets. Throughout the paper, uncertainties represent the
238 variability in the time series of that variable and are calculated as plus/minus one standard deviation.

239 **3 Results**

240 **3.1 Ocean variability beneath TEIS**

241 Hydrographic properties observed by the MicroCATs show variability across a wide range of timescales (Fig. 2). Θ and S_A
242 increase with depth, with mean Θ of -0.88 ± 0.24 °C at 316 m, 0.34 ± 0.09 °C at 521 m, and 1.04 ± 0.04 °C and 1.05 ± 0.03
243 °C near the seafloor at depths of 745 m and 784 m, respectively (Fig. 2). We observe a warming trend with time at all depths,
244 following the sensor deployment in January 2020 at the shallow and mid-depth layers, and around April 2020 at the deeper
245 layers. This warming persisted until July 2021. After this, warming stalled at depth, while mid-depth and shallow layers
246 cooled until January 2022. Thereafter, warming resumed at mid-depth and both deeper layers, continuing through to January
247 2023.

248 From January 2020 to July 2021, the shallow MicroCAT recorded a 1°C increase in Θ at a rate of 0.4 °C yr⁻¹, followed
249 by a 1 °C decrease at an accelerated rate of -1.8 °C yr⁻¹ until the instrument ceased operation in January 2022 (Fig. 2a). After
250 July 2021, fluctuations in S_A became more pronounced, consistently exceeding the overall mean of 34.23 g kg⁻¹ and
251 exhibiting a declining trend from July 2021 to January 2022. The Pearson correlation coefficient between Θ and S_A at the
252 shallow instrument was 0.4 before July 2021, increasing to 0.7 afterwards.

253 The mid-depth MicroCAT recorded a 0.1 °C increase in Θ over the entire record, although in a stepped fashion (Fig.
254 2b). The warming trend was 0.2 °C yr⁻¹ until July 2021, steepening notably between March and July 2021, when Θ and S_A

255 increased in tandem. This was followed by a gradual decline beyond their initial values at a rate of $-0.5\text{ }^{\circ}\text{C yr}^{-1}$ until January
256 2022, after which warming resumed at $0.2\text{ }^{\circ}\text{C yr}^{-1}$ until January 2023.

257 Both deep MicroCATs recorded a $0.1\text{ }^{\circ}\text{C}$ warming from April 2020 to January 2023, accompanied by a 0.02 g kg^{-1}
258 increase in S_A (Fig. 2c,d). Θ and S_A fluctuations were generally synchronous at both deep MicroCATs. Near the seabed at
259 Cavity Camp, warming occurred at a rate of $0.04\text{ }^{\circ}\text{C yr}^{-1}$ until July 2021, then plateaued until January 2022, after which it
260 resumed warming at a rate of $0.01\text{ }^{\circ}\text{C yr}^{-1}$ until January 2023. At Channel Camp, the warming trend near the seabed was also
261 $0.04\text{ }^{\circ}\text{C yr}^{-1}$ until July 2021, then plateaued before increasing to $0.02\text{ }^{\circ}\text{C yr}^{-1}$ after January 2022. This suggests that between
262 January 2022 and January 2023, the warming trend re-emerged in both mid-depth and deep layers.

263 Superimposed on the long-term variability, we observe several distinct events, characterized by rapid Θ and S_A
264 excursions over several weeks, notably in April and July 2020, as well as in February and April 2021. During these events,
265 concurrent decreases in Θ and S_A of more than $0.05\text{ }^{\circ}\text{C}$ and 0.03 g kg^{-1} , respectively, were recorded at the deep sites. The
266 mid-depth and shallow instruments simultaneously displayed opposite signals to the deep sites, with rising Θ and S_A
267 anomalies of more than $0.3\text{ }^{\circ}\text{C}$ and 0.04 g kg^{-1} , and $0.2\text{ }^{\circ}\text{C}$ and 0.03 g kg^{-1} , respectively. Simultaneous current velocity
268 measurements revealed accelerated current speeds at all depths during those events (grey-shaded time spans in Fig. 2).

269 Between January 2020 and January 2022, both shallow (315 m) and deep (782 m) sensors at Channel Camp sank at
270 rates of 2.21 m yr^{-1} and 2.17 m yr^{-1} , respectively (Appendix C). A background sinking rate of approximately 1.86 m yr^{-1} is
271 derived from the AMIGOS-3 elevation record. The shallow MicroCAT stopped recording on January 11, 2022, at 319 m
272 depth, while the deep sensor continued operating until January 2, 2023, reaching 788 m (Fig. C1). Notably, the sinking rate
273 of the deep sensor decreased to 1.62 m yr^{-1} during 2022, indicating a possible water-mass change in the overlying water
274 column and a concurrent decline in firn compaction, a non-linear process that occurs rapidly at first but slows over time as
275 the underlying firn becomes denser. At Cavity Camp, the mid-depth MicroCAT was initially deployed at 520 m and the deep
276 MicroCAT at 744 m. Both began recording on January 2, 2020, and continued until December 26, 2022, reaching depths of
277 523 m and 747 m, respectively (Fig. C2). The consistent sinking trends observed at each site, along with the strong
278 agreement between pressure records from sensors at the same site, rule out the possibility that the mooring cables grounded
279 on the seafloor.

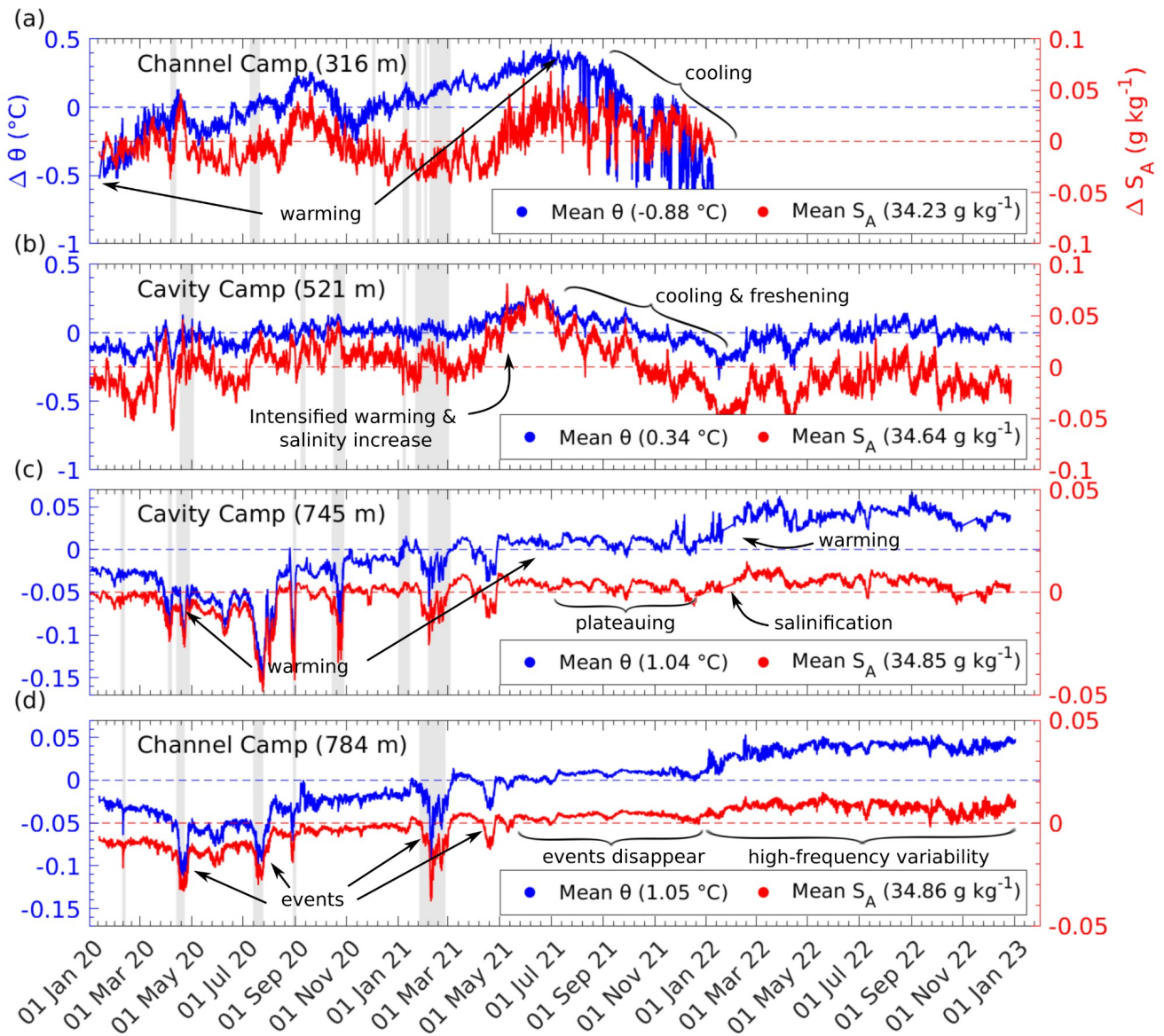


Figure 2: Time series of anomalies in conservative temperature (Θ) and absolute salinity (S_A) at (a) 316 m at Channel Camp, (b) 521 m at Cavity Camp, (c) 745 m at Cavity Camp and (d) 784 m at Channel Camp. Mean values of Θ and S_A are indicated in the respective legends. Grey bars indicate periods when the measured current speeds were elevated. No additional Aquadopp current meter data are available after March 2021.

Our dataset exhibits variability across multiple time scales, with certain signals emerging or fading throughout the duration of the record. The continuous wavelet transforms visualize periods of pronounced density variability (Fig. 3). Clusters of relatively high wavelet power, enclosed by contours indicating statistical significance, highlight how density anomalies at different depths evolve over time. Statistical significance declines sharply at all depths for periods shorter than 0.5 days. At shallow depths, statistically significant covariance with a Morlet wavelet at periods longer than 8 days is only identified in

290 April 2020 (Fig. 3a). However, from July 2021 until October 2021, we see increases in both power and statistical
291 significance for periods between 0.5 and 8 days, indicating a change in water masses. At mid-depth, similar covariance with
292 periods up to 24 days emerges in April 2020, with occasional occurrences of significant covariance lasting more than a week
293 observed in September 2020 (Fig. 3b). Following this, multi-day covariance shifts primarily to sub-daily covariance for most
294 of the remaining record. At greater depths, statistically significant covariance with periods lasting several months is
295 observed, especially at Cavity Camp (Fig. 3c). This longer-term covariance diminishes after July 2021, shifting toward
296 shorter periods of around one day by January 2022.

297 Notably, the long-term signal at depth is overlaid by significant diurnal and semi-diurnal fluctuations, which are also
298 more prominent at Cavity Camp than Channel Camp (Fig. 3c,d). This shorter-term variability is closely tied to the prevailing
299 tidal regime, which is predominantly diurnal with some semi-diurnal components. Significant tidal periods exhibit enhanced
300 power with a fortnightly modulation, indicating influence from the 14-day spring-neap tidal cycle. We observe, however,
301 only little covariance at tidal periods in most of the shallow record and throughout the mid-depth record, whereas tidal
302 covariance is evident at both deep sites (Fig. 3c,d).

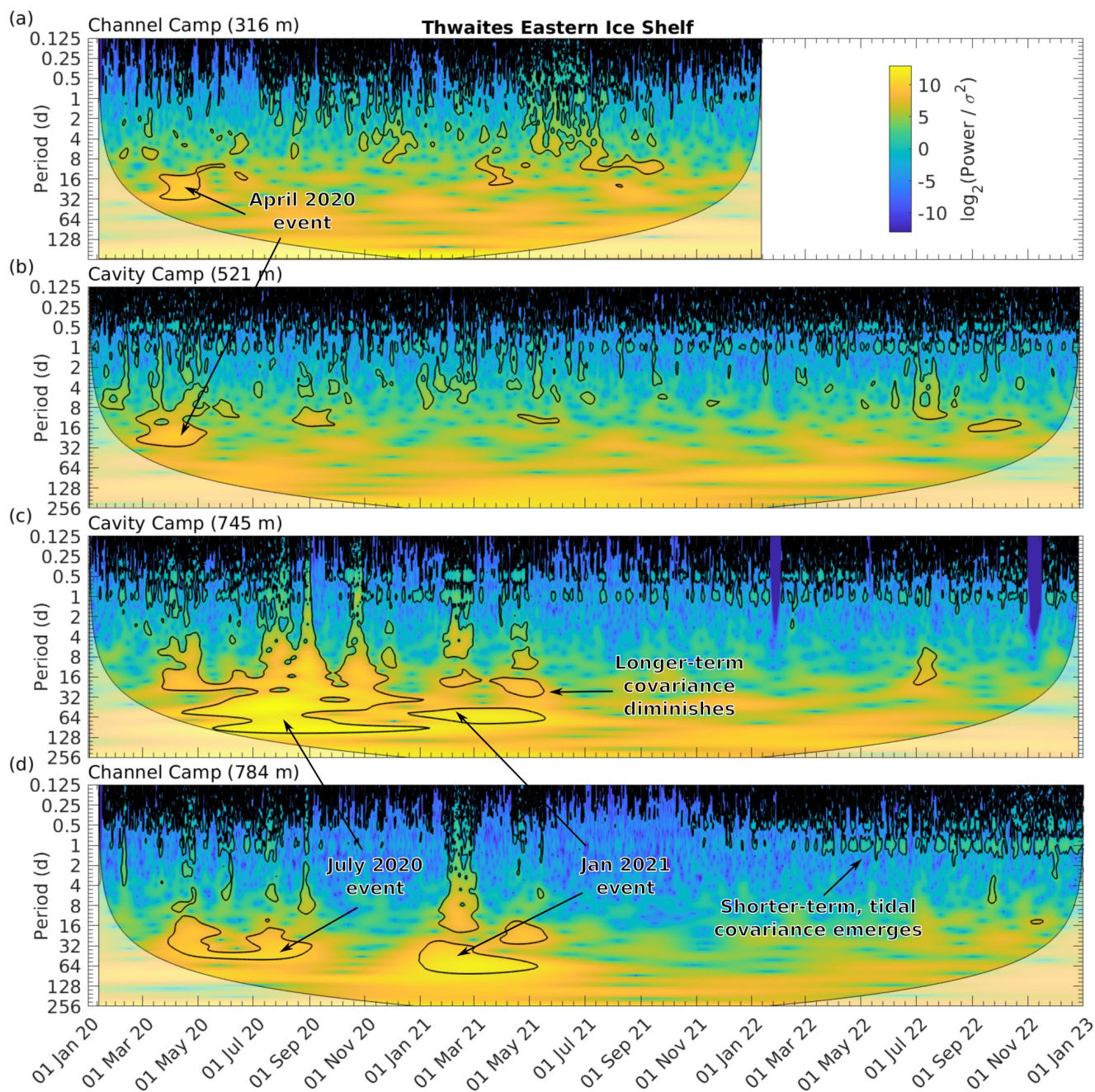
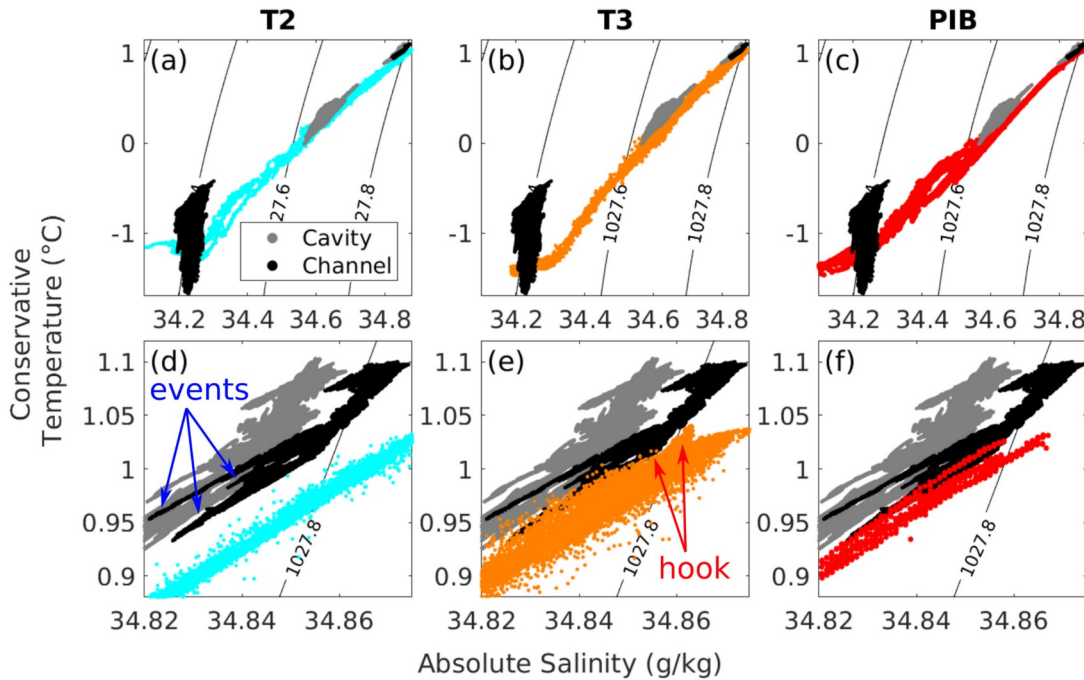


Figure 3: Continuous-wavelet transform of potential density time series at (a) shallow, (b) mid-depth, and deep sensors (c) at Cavity Camp and (d) Channel Camp. Warm colours show high power at the corresponding period. Black contours depict statistical significance. Greyed out is the cone of influence where edge effects might obscure the cross-wavelet transform.

3.2 Linking subshelf cavity observations to PIB-sourced waters

Different water masses have characteristic combinations of conservative temperature (Θ ; °C) and absolute salinity (S_A ; g kg⁻¹), which can be used to trace their origin beneath TEIS. To identify the water sources advecting past our sensors at shallow, mid-depth, and deep layers, we compare our MicroCAT CTD data recorded from January 2020 to January 2023 with two AUV datasets collected at T2 and T3 in February and March 2019 (Wåhlin et al., 2021) as well as a set of ship-based CTD measurements from PIB collected during the same cruise (see Fig. 1 for locations). This comparison is visualized in Θ - S_A diagrams to illustrate the distinct water masses and their interactions. PIB-sourced water is generally the warmest throughout the water column, followed by T3 and T2 (Wåhlin et al., 2021). At depth, our measurements from both sites align most closely with those from PIB (Fig. 4d–f). The observed events at depth are characterized by cold and fresh water types (blue arrows in Fig. 4d). Notably, a distinct hook in our deep-layer data, observed at both Cavity Camp and Channel Camp, follows the 1027.8 kg m⁻³ isopycnal (red arrows in Fig. 4e). This characteristic, also present in the AUV data from T3, was previously traced to PIB by Wåhlin et al. (2021) using both Θ - S_A as well as dissolved oxygen and results from isopycnal mixing between PIB and Thwaites Trough water, indicating the far western extent of PIB influence. The slope of this hook is also represented in our hydrographic data, even more prominently than in the T3 AUV dataset, though with a slight offset in S_A (Fig. 4e). Overall, our analysis shows that the water masses beneath TEIS originate from PIB. Additionally, none of our measurements overlap with the coldest water masses observed at T2 in Θ - S_A space, reinforcing the hypothesis of Wåhlin et al. (2021) that cooled, meltwater-enriched water exits the subshelf cavity via T2.

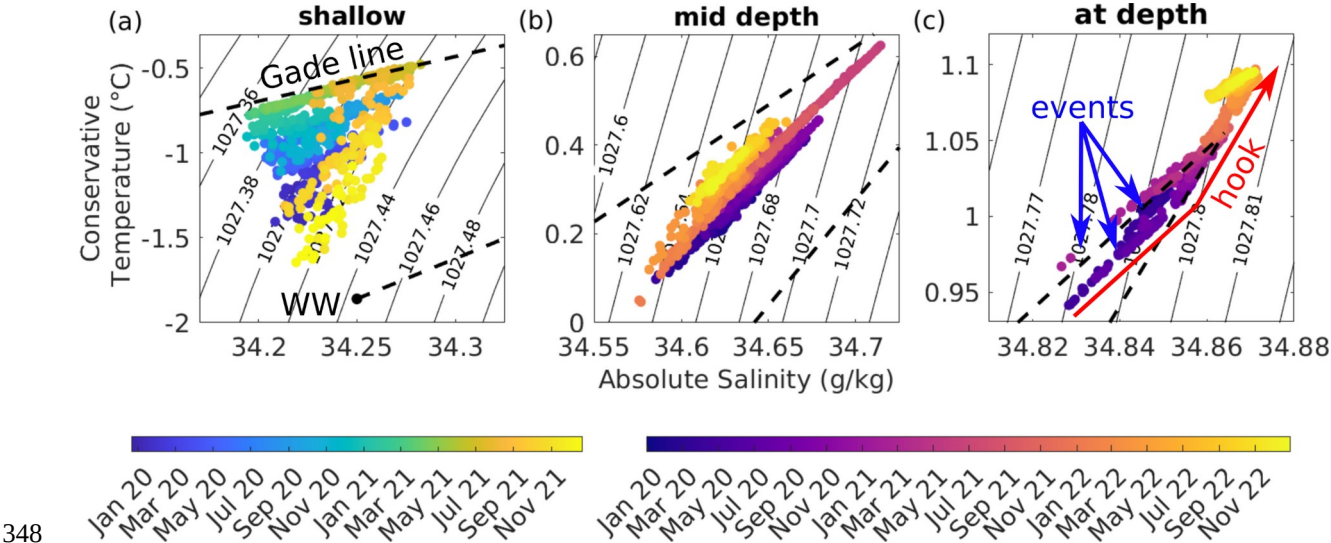


326 **Figure 4:** Θ - S_A diagrams from MicroCATs at Cavity and Channel Camp (grey and black), compared with AUV measurements at (a) T2
 327 (blue), (b) T3 (orange), and (c) ship-based CTD (red) in PIB. See Figure 1 for a map of these locations. (d-f) Close-up views of the mCDW
 328 layer at depth, with potential density isopycnals (kg m^{-3}). Labels refer to features discussed in the text.

329 3.3 Tracing glacial meltwater and Winter Water mixing beneath TEIS

330 We observe temporal changes in the hydrographic properties of water masses at three surveyed depths. In a Θ - S_A diagram,
 331 mixing between two water masses results in intermediate properties that lie along a straight line connecting their respective
 332 endmembers. The Gade line represents the mixing between glacial meltwater and mCDW, where small salinity changes
 333 correspond to significant temperature variations due to heat and salt exchange during ice melting (Gade, 1979). The mCDW-
 334 Winter Water (WW) mixing line, on the other hand, reflects the dilution of WW with mCDW. WW is characterized by a
 335 subsurface temperature minimum and represents the remnant of the winter surface mixed layer, which becomes capped in
 336 summer by fresher and warmer water due to sea ice melt and air-sea heat fluxes. At the shallow MicroCAT, water masses
 337 gradually shift toward the Gade line from January 2020 to January 2021 and closely follow it until July 2021 (Fig. 5a).
 338 Thereafter, they align with the 1027.42 g kg^{-1} isopycnal, indicating reduced glacial meltwater influence due to increased WW
 339 advection into the TEIS subshelf cavity. At mid-depth, data cluster along a linear trend between the Gade and WW mixing
 340 lines, suggesting a stable water mass structure with a gradual warming and freshening trend (Fig. 5b). At depth, waters
 341 follow a narrow mixing path between these two lines, with long-term warming and salinification. The highlighted events,
 342 where Θ and S_A exhibit low values for several weeks, align with the Gade line (blue arrows in Fig. 5c), while the long-term
 343 evolution of the densest waters follows an extension of the WW mixing line. This characteristic "hook" shape (red arrow in
 344 Fig. 5c), previously identified by Wåhlin et al. (2021), is indicative of PIB-sourced mCDW mixing with T3 waters (Fig. 4e).
 345 In summary, this indicates that PIB-sourced mCDW mixed with glacial meltwater between January 2020 and July 2021,
 346 after which WW became the dominant water mass advected beneath TEIS.

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349 **Figure 5:** Θ - S_A diagrams from MicroCATs at Cavity Camp and Channel Camp, showing changes in water mass composition and mixing
350 over time. (a) The shallow record (316 m) covers only the period from January 2020 to January 2022, while (b) the mid-depth (521 m) and
351 (c) deep records (784 m) extend from January 2020 to January 2023. In all panels, the upper dashed line represents the Gade line,
352 indicating water mass modification through ice-shelf melting, while the lower dashed line is the WW mixing line, showing the influence of
353 cold surface water mixing. Solid black lines represent isopycnals of potential density (kg m^{-3}). Labels refer to features discussed in the text.

354 3.4 Wind and ocean current dynamics

355 To provide context to the events we observed in our hydrographic data, we analyze the temporal variability of wind forcing
356 at the surface and ocean currents beneath TEIS, which both influence the transport of water masses. These environmental
357 conditions are visualized using feather plots, where vector length represents the magnitude of wind and current speeds. The
358 orientation of the wind vectors shows the direction from which the wind is blowing, while the current vectors indicate the
359 direction to which the ocean currents are flowing, with true north pointing upward.

360 Winds sweeping across the ice-shelf surface predominantly originate from the ESE (Fig. 6a). The average wind
361 speed at Channel Camp was 10 m s^{-1} , with occasional spikes surpassing 60 m s^{-1} during winter or early spring. In-situ and
362 ERA5 air temperature and wind speed showed strong agreement, whereas wind direction data agreed to a much lesser extent
363 (Appendix B).

364 The ocean currents beneath TEIS are usually slow ($< 4 \text{ cm s}^{-1}$) and toward the SSW with one exception. Aquadopp
365 records from both sites agreed that the slowest mean currents occurred deepest within the water column (Cavity, 745 m
366 depth: $0.9 \pm 0.7 \text{ cm s}^{-1}$, Channel at 784 m: $0.8 \pm 0.8 \text{ cm s}^{-1}$). Shallow (Channel; 316 m) and mid-depth (Cavity; 521 m) mean
367 current speeds were progressively faster at $2.2 \pm 1.8 \text{ cm s}^{-1}$ and $3.7 \pm 2.2 \text{ cm s}^{-1}$, respectively. The shallow, mid-depth and
368 deep Cavity currents had similar mean current directions, predominantly flowing to the SSW ($211^\circ \pm 71^\circ$, $221^\circ \pm 58^\circ$, and
369 $227^\circ \pm 64^\circ$, respectively), while the deep Channel site was anomalous, flowing towards the North with higher temporal
370 variability ($8^\circ \pm 136^\circ$).

371 Current velocities deviated significantly from the mean during the hydrographic events noted in section 3.1. During the
372 April 2020 event, currents at the shallow Aquadopp intensified, reaching speeds exceeding 7 cm s^{-1} and flowing toward
373 NNW (Fig. 6b). At mid-depth, currents accelerated to a similar magnitude but flowed toward the SW (Fig. 6c). In the deep
374 layer, currents also flowed toward SW, with a maximum recorded speed of 4.6 cm s^{-1} on April 18, 2020 (Fig. 6d,e). Another
375 event occurred in July 2020, when the shallow Aquadopp at Channel Camp recorded an accelerated current of 9 cm s^{-1} ,
376 flowing toward the SSW. However, this event was not clearly observed at the deep Aquadopp at Channel Camp, and data
377 gaps from both Aquadopps at Cavity Camp prevent further investigation. The most widespread event occurred in February
378 2021, when all four Aquadopps recorded elevated current speeds. The shallow Aquadopp measured persistent currents of ~ 9
379 cm s^{-1} toward the SSW, while the mid-depth Aquadopp recorded even higher speeds of $\sim 11 \text{ cm s}^{-1}$ directed SE. At Cavity
380 Camp, the deep Aquadopp peaked at 4 cm s^{-1} toward the SW on February 7, 2021, whereas the deep Aquadopp at Channel
381 Camp exhibited a contrasting current direction of 5 cm s^{-1} toward the NW. The Aquadopps ceased operation before the
382 fourth temperature and salinity excursion in April 2021, preventing the determination of dominant current directions for this
383 event.

384 At all depths, multi-weekly temperature and salinity anomalies, likely accompanied by enhanced current speeds, ended
385 after May 2021 and were replaced by increased shorter-period covariance (0.5 to 16 days; Fig. 3).

386 **3.5 Linking environmental drivers and density variations across depths**

387 To determine if the changes in hydrography during the events are driven by ocean currents, we performed the cross-wavelet
388 transform between water density and current speed. For the shallow and mid-depth sensors, increasing current speeds are
389 associated with increasing density, while at depth increasing current speed is associated with decreasing density. In a cross-
390 wavelet transform, periods of strong similarity between the two time series are shown as clusters of high common power. A
391 surrounding black contour indicates the statistical significance of these clusters. We identify significant long-period
392 covariance between one to four weeks in April 2020 and in July 2020. All sensors show covariance from sub-daily to multi-
393 weekly time periods in February 2021 (Fig. 6g-j). These covariances confirm that ocean currents mainly drive the observed
394 hydrographic variability during the events. We also find significant multi-week covariance between ERA5 wind speed and
395 density variations at the shallow ocean sensor in April and July 2020 (Fig. 6f).

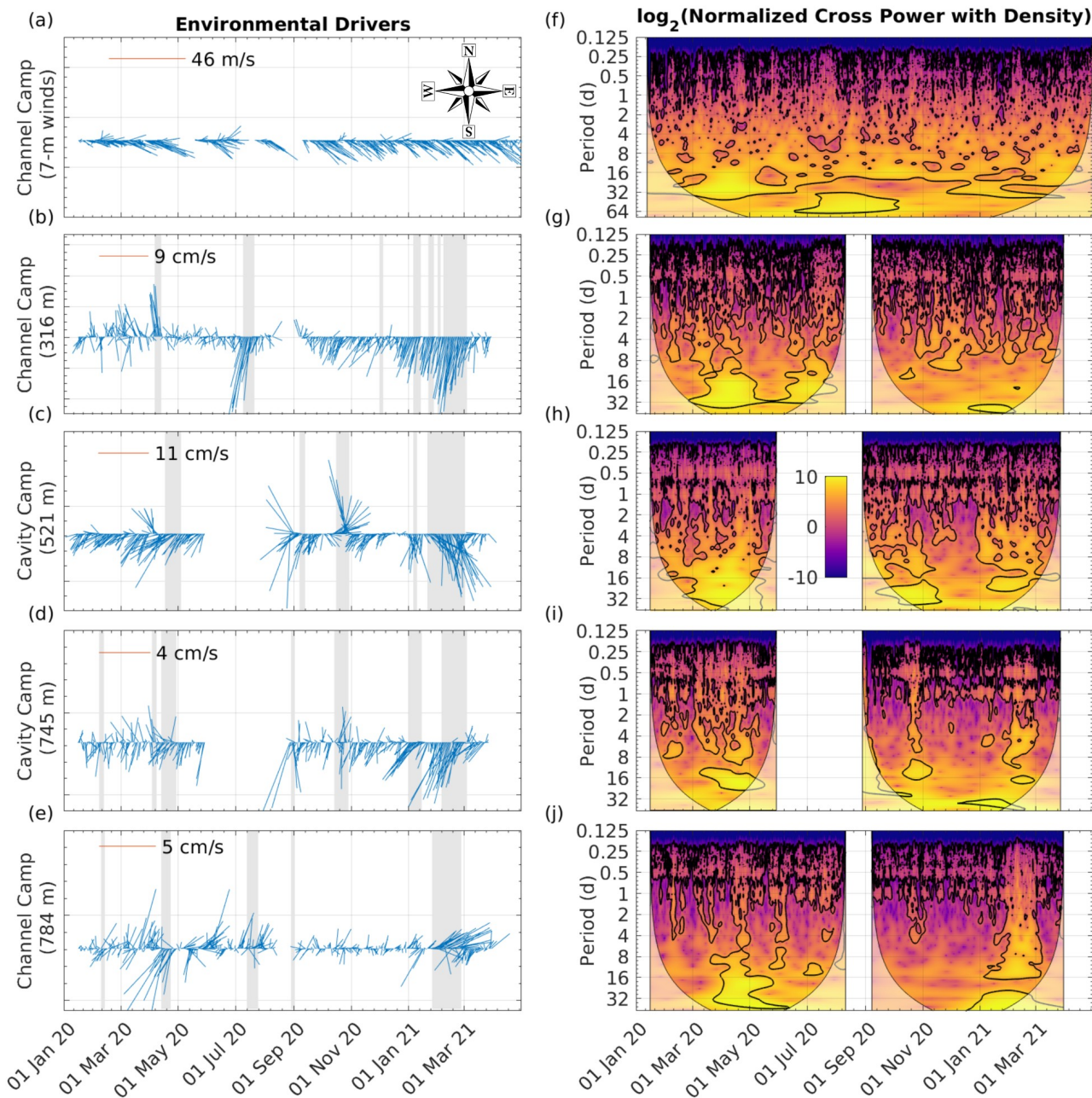


Figure 6: Feather plots of average daily (a) in-situ wind and (b-e) current speed and direction from January 2020 to March 2021. The line orientation represents wind and current direction (with the top of the graph indicating North or 360°), while line length corresponds to speed. Wind direction follows the meteorological convention, indicating the direction from which the wind originates, whereas currents are shown flowing toward their respective directions. The grey shaded areas denote periods of elevated current speeds as discussed in the text. (f) Cross-wavelet transform between shallow density and ERA5 windspeed covariance. (g-j) Cross-wavelet transforms between density and current speed time series for each depth.

3.6 Consistent thermal patterns observed during events

We now use the DTS profiles to assess the vertical structure and extent of temperature changes during the events, complementing the more sparsely-spaced MicroCAT time series. The DTS temperature profiles at Channel Camp during the four highlighted events reveal a consistent pattern of temperature changes within the water column (Fig. 7). Throughout all events, water masses between 400 and 600 m depth exhibit anomalous warm temperatures, with the most pronounced temperature increase occurring around 450 to 500 m depth. Conversely, the deeper water between 600 and 800 m experiences anomalous cool temperatures, which is strongest at 700 m depth. Additionally, a near-isothermal layer forms between 300 and 400 m, suggesting vertical mixing over this depth range. The temperature profiles show a progressive shift in thermal structure, with warming and cooling trends developing simultaneously in distinct layers. Notably, the 600 m depth emerges as a clear transition point, marking the boundary between the warming upper layers and the cooling deeper waters.

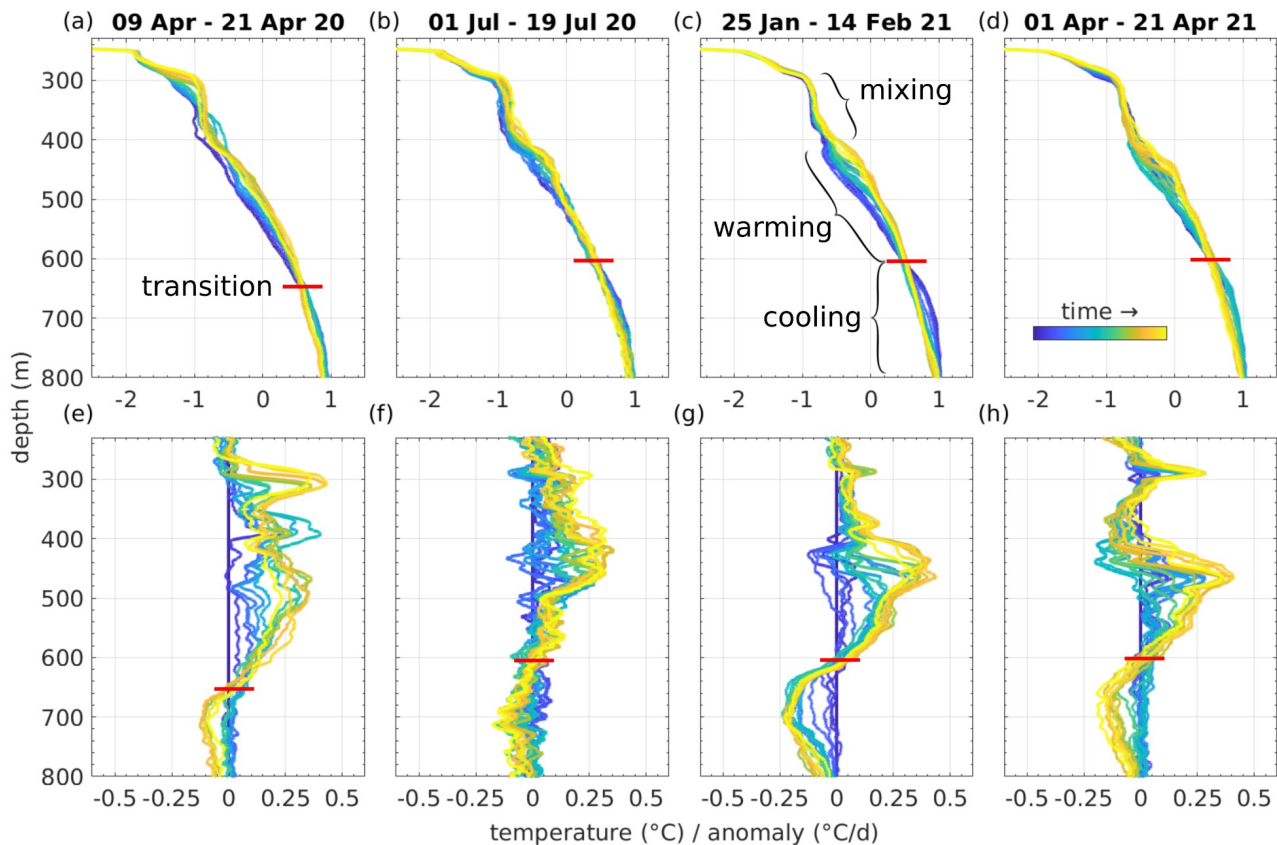
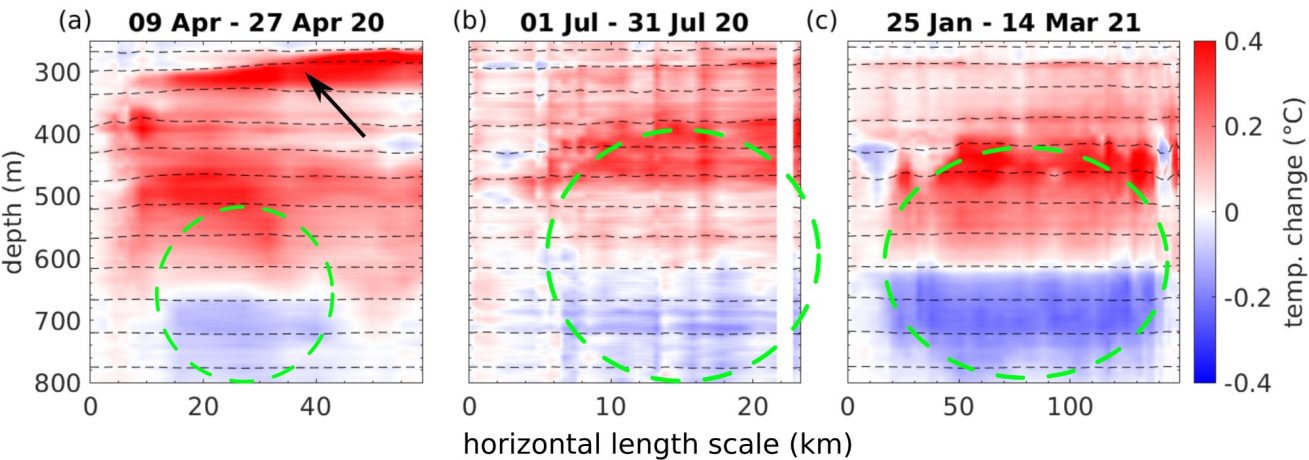


Figure 7: Daily mean DTS temperature from Channel Camp profiles for the specified events. The plots reveal a warming trend in the upper two-thirds of the water column, accompanied by cooling in the lower third. The profiles are colour-coded, transitioning from cool to warm colours, to represent the progression of time.

419

420 To estimate the horizontal length scale of the advecting features, we combined DTS temperature anomalies with current
421 speed measurements from Aquadopps instruments (Fig. 8). Specifically, we used mid-depth currents (521 m) at Cavity Camp
422 and near-bottom currents (784 m) at Channel Camp to calculate daily mean speeds, which we assumed represent the entire
423 water column. Using these speeds (Fig. 6), we converted the duration of temperature anomalies into horizontal length scales.
424 The April 2020 event corresponds to a feature ~30 km long (Fig. 8a), while the July 2020 event is ~20 km (Fig. 8b), though
425 a data gap during the austral winter limited its full characterization. The February 2021 event is the largest and clearest, with
426 an estimated length scale of ~100 km (Fig. 8c). Malfunctioning Aquadopps in March 2021 prevented assessment of the April
427 2021 event.

428 Isopycnals, estimated by combining the DTS temperature profiles with salinity from CTD profiling on January 12,
429 2020, indicate that the warming observed between 400 m and ~600 m depth is associated with minimal upward displacement
430 of isopycnals, while cooling between 600 and 800 m depth results in negligible downward displacement (Fig. 8). At depth,
431 changes in density are driven primarily by changes in salinity, which do not show a large vertical gradient (Appendix A),
432 explaining the relatively smaller isopycnal shifts in the deeper layer compared to mid-depth (Fig. 8).



433

434 **Figure 8:** Temperature anomalies over time for three distinct periods. Each panel shows the deviation from the first profile in the
435 respective period. The colour scale represents the magnitude of temperature change, with negative values indicating cooler temperatures at
436 depth and positive values indicating warmer temperatures above ~600 m depth. The x-axis reflects the distance travelled by features
437 advecting through the water column, based on Aquadopps current speed measurements, available during the first three events. Dashed black
438 lines show isopycnals. The black arrow in panel (a) shows warming in the shallowest layer discussed in the text. Dashed green circles
439 show the identified features.

440

441 3.7 Thermodynamics in the water column

442 The DTS data provide a continuous vertical record of ocean temperatures. Both mooring sites feature an approximately 100
443 m thick layer of mCDW near the bottom that exhibits temperatures exceeding 1.1 °C. This bottommost layer is not only
444 warming with time (Fig. 2c/d), but also thickening by about 50 m in its vertical extent throughout the record (Fig. 9).
445 Situated above this warmest layer, a 200-m-thick zone demonstrates a sharp thermocline between 500 and 700 m depth, with
446 temperatures generally above 0 °C. Further up the water column lies another 200-m-thick layer (300 to 500 m deep),
447 characterized by temperatures between -1 and 0 °C. At the Channel Camp site (Fig. 9a-c), within a narrow band spanning the
448 next 40 m, a thin layer approaches -1.5 °C, nearing the in-situ freezing point at approximately -2 °C. This cold layer thins
449 between January 2020 and July 2021 at this site. In the immediate vicinity of the ice-shelf base, a 2-3 m thin layer at the
450 pressure melting point (-2 °C at about 250 m depth) is observed. This insulating layer, which has also been documented in
451 proximity to the ice-shelf grounding zone at greater depth, effectively suppresses basal melting through strong stratification
452 (Davis et al., 2023). The ice base with a draft of 260 m lies above the depth of the mCDW which is greater than 600 m. Even
453 without the insulating layer, the thermal forcing is low and insufficient to sustain significant basal melt rates.

454 The DTS record at the Channel Camp site suffers a substantial data gap from August 2021 to January 2023 (Fig. 9a),
455 but reveals a significant cooling trend of more than 1.2 °C in the upper half of the water column across that gap (Fig. 9c).
456 This cooling in the 250 m directly beneath the floating ice contrasts with the continuous DTS record prior to the data gap,
457 suggesting considerable changes in the subshelf hydrographic properties. Notably, the 40-m-thick cold layer, nearing the in-
458 situ freezing point that is observed in the August 2021 profile, expanded to a 150-m-thick layer (250-400 m depth) in the
459 January 2023 profile (Fig. 9c). Between 400 and 500 m depth, a sharp temperature gradient of 0.013 °C m⁻¹ is observed.
460 However, the lower half of the water column exhibits temperatures similar to those observed in August 2021, suggesting that
461 the water masses in the lower half of the water column persisted, while the upper half experienced a considerable change in
462 hydrographic properties. This decrease in temperature corresponds to a change in mean water column density from 1029.3 to
463 1029.1 kg m⁻³, assuming no change in salinity between 250 to 500 m depth, and is therefore negligible when inverting
464 remotely-sensed ice-shelf freeboard to ice thickness (Appendix A).

465 The DTS record at Cavity Camp is similar to the record at Channel Camp but provides additional data from August to
466 the end of October 2021, after which no further DTS measurements were taken at this site. Notably, the Cavity Camp DTS
467 recorded the onset of the cooling of the upper water column (Fig. 9d). By analyzing the last 100 DTS profiles dating back to
468 June 2021, we determined that the cooling occurred rapidly in late July 2021, reaching a depth of approximately 450 m
469 before the DTS record ended by early October 2021 (Fig. 9e,f).

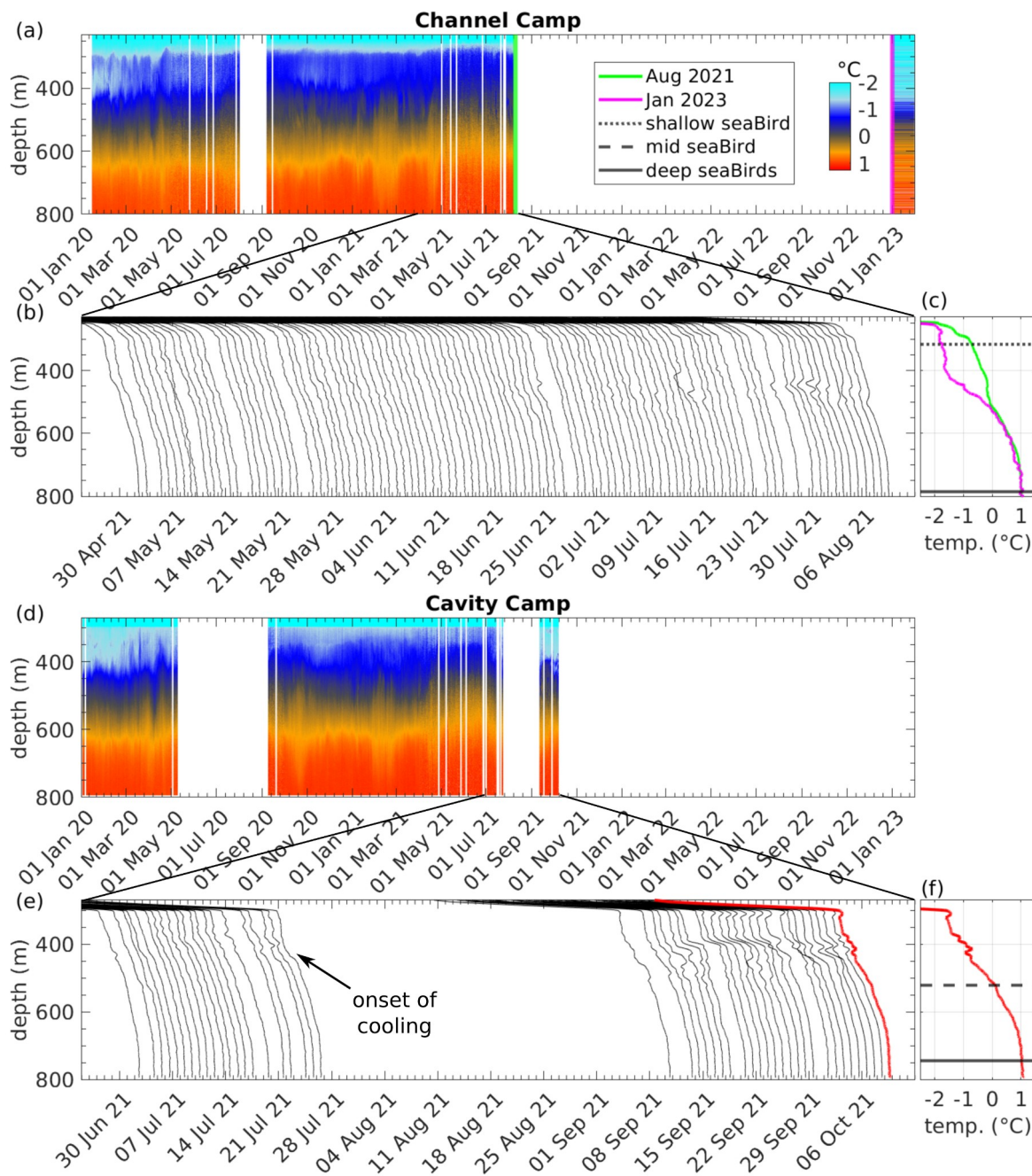
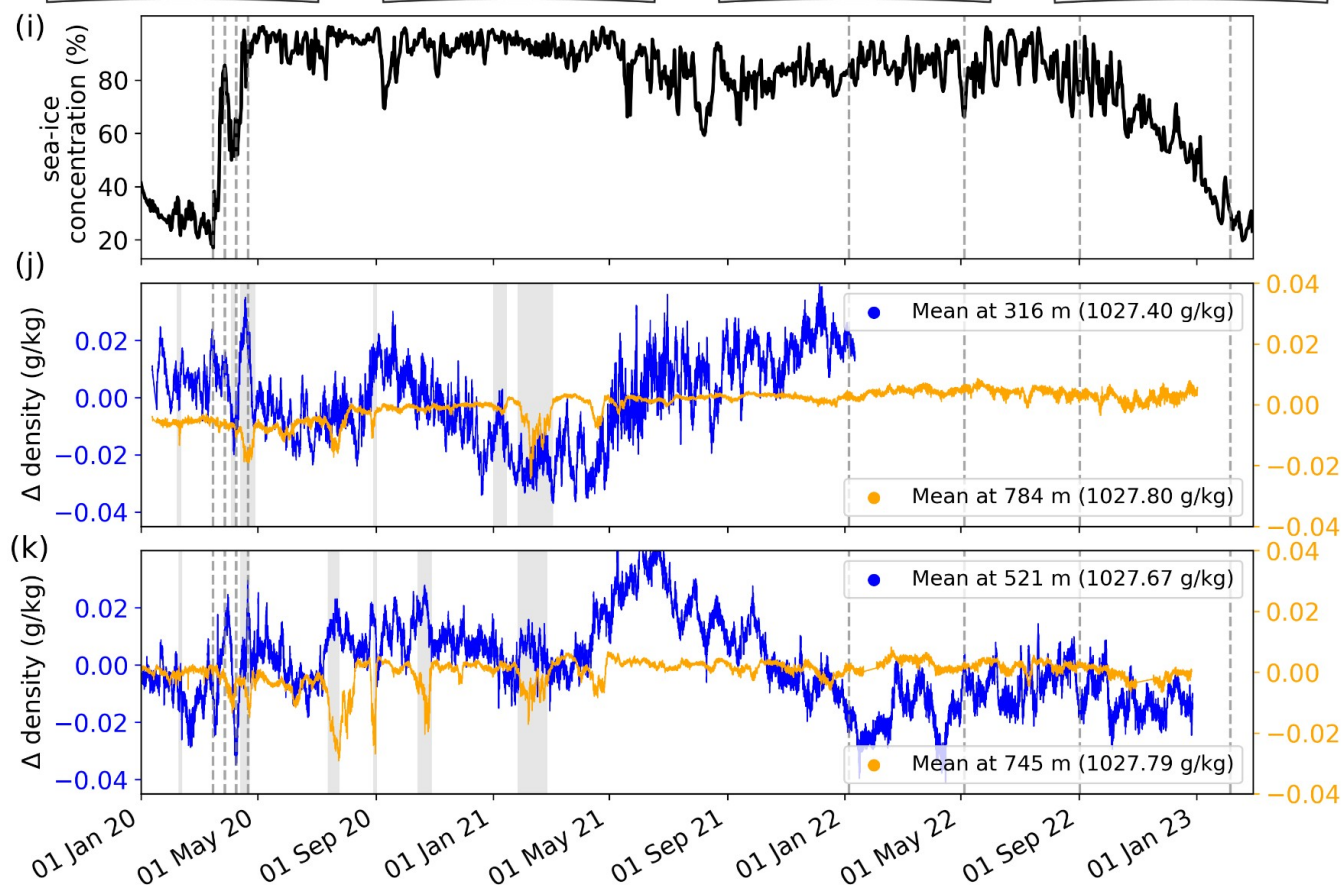
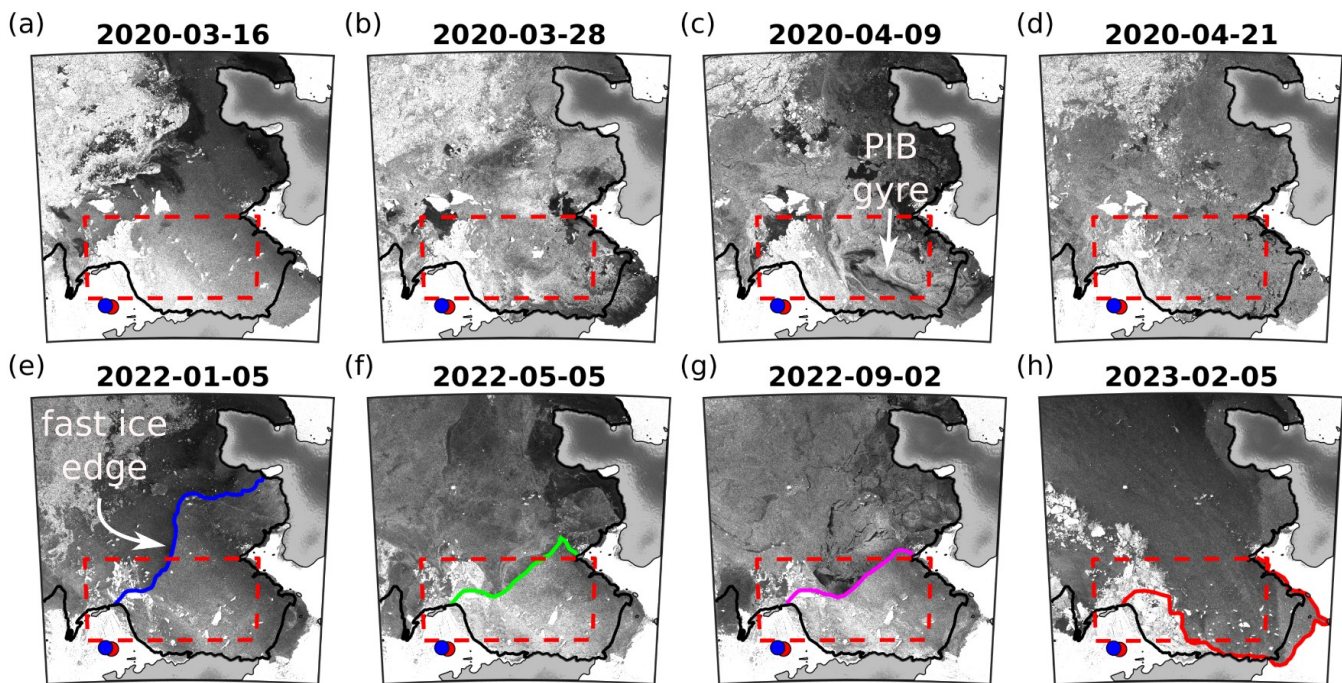


Figure 9: Daily-binned temperature records from DTS at (a) Channel Camp and (d) Cavity Camp. (c) Last temperature profile before the August 2021 - January 2023 gap and the first measurement in January 2023, highlighting cooling in the upper water column. Dotted, dashed, and solid black lines indicate the depths of shallow, mid-depth, and deep ocean sensors. (b) and (e) Waterfall diagram of the last 100 DTS profiles at Channel Camp and Cavity Camp, showing abrupt cooling between 300 and 400 m depth. The temperature range of each line is presented in (f), with an example of the last DTS profile from October 2021 (red). Note that there is a period with no data in August and September 2021 at Cavity Camp. The DTS profiles shown in the waterfall plots were smoothed for visualization with a running mean of 40 sample points (corresponding to ~10 m along the cable).

478 3.8 Sea ice conditions in PIB: formation and breakup of fast ice

479 We examine the multi-year evolution of sea ice coverage in PIB to identify the potential drivers of variability in
480 hydrographic properties beneath TEIS. At the start of our observational period in the austral summer of 2019/20, PIB was
481 largely free of sea ice (Fig. 10i), with open water extending from TEIS to the ice front of Pine Island Glacier (Suppl. Video).
482 As surface air temperatures dropped below -10°C through March 2020 and winds remained generally calm (Fig. B1), thin
483 first-year sea ice began to form (Fig. 10a/b). By late March and into April 2020, a major sea ice breakout event occurred,
484 driven by strong easterly winds exceeding 20 m s^{-1} . These winds fractured the newly formed ice and redistributed it,
485 revealing an active PIB gyre in satellite SAR imagery, marked by the cyclonic movement of sea ice (Fig. 10c). By mid-April
486 winds calmed to around 5 m s^{-1} and air temperatures stayed below -10°C (Appendix B), promoting sea ice formation by
487 latent heat loss and leading to near-complete sea ice coverage in PIB (Fig. 10d). This coverage persisted through the
488 following two austral summers (2020/21 and 2021/22).

489 The fast ice cover remained until January 2022 after which the fast ice front gradually retreated (Fig. 10f/g),
490 eventually breaking up in October 2022 and leading to open-water conditions in PIB once again by February 2023 (Fig.
491 10h).

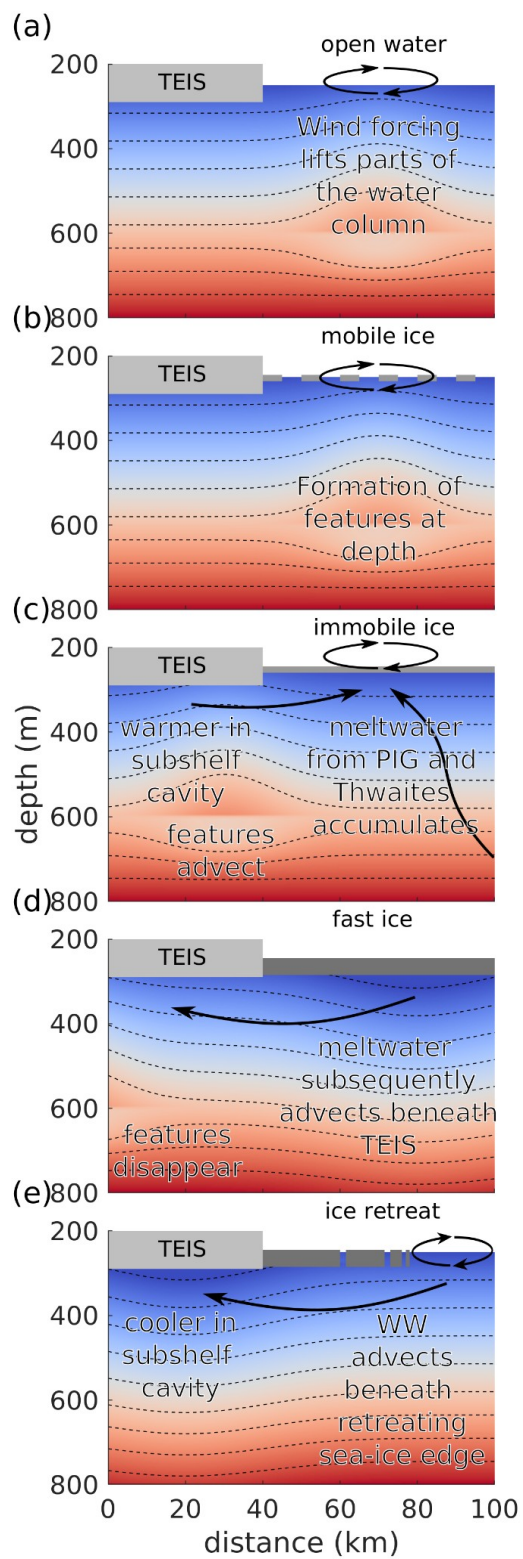


493 **Figure 10:** Co-evolution of PIB sea ice and Thwaites sub-ice shelf ocean densities. Panels (a-d) present Sentinel-1A SAR images
 494 depicting a first-year, sea ice breakout occurring between mid-March and late-April 2020. Panels (e-h) show the retreat of the multi-year,
 495 fast-ice edge to the grounding line of Pine Island Glacier. The dashed red rectangle shows the sea ice concentration sampling box. The
 496 black line indicates the position of the ice-shelf front and grounding line (Bindshadler et al., 2011). Red and blue dots denote Channel
 497 Camp and Cavity Camp locations on TEIS. Panel (i) shows sea ice concentration time series in PIB. Panels (j) and (k) display time-series
 498 data of ocean water density anomalies at these sites across various depths. Grey dashed lines indicate the times of SAR image capture
 499 shown in panels (a-h), grey bars in panels (j,k) indicate periods when the measured current speeds were elevated.

500 **4 Discussion**

501 During the April 2020 sea ice breakout, we observed the first event of opposing density anomalies between the shallow/mid-
 502 depth and deep sensors, with anomalies exceeding 0.03 g kg^{-1} (Fig. 10j/kf/g). Similar anomalies occurred in July 2020, as
 503 well as in February and April 2021, when thin first-year sea ice is moving around PIB. However, these events disappeared
 504 after May 2021, when the now second-year sea ice became more firmly fastened across PIB (Fig. 10e).

505 We propose that the events of anomalous temperatures are driven by processes causing heaving and sinking around an
 506 expanding layer at 600 m depth, which marks the top of the mCDW layer. During the events, water-mass properties change
 507 both upward and downward (Fig. 7), suggesting the influence of gyre-scale features moving through the water column and
 508 driving its transient evolution (Fig. 8). This interpretation is supported by the DTS profiles, which reveal periodic excursions
 509 of water-mass properties centered around 600 m depth (Fig. 9a/d). Additionally, during these events, the hydrographic
 510 properties shift back and forth along a distinct trajectory, indicating that no mixing of water masses occurs.



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512 **Figure 11:** Schematic representation of the interactions between sea ice dynamics and hydrographic variability.

513

514 **4.1 Gyre-scale features formed during mobile, first-year sea ice breakouts**

515 In April 2020, the layers within 50 m of the ice base experienced significant warming following the passage of a gyre-scale
516 feature (Fig. 8a). At 316 m depth, where the shallow CTD is located, we observe a faster NNW-directed current (Fig. 6b).
517 During this time, PIB is covered by mobile, first-year sea ice (Fig. 10a-d), and southerly winds blow across the ice-shelf
518 surface (Fig. 6a). Density at shallow levels increases throughout the month by approximately 0.04 g kg^{-1} (Fig. 10j). Cross-
519 wavelet analysis reveals significant covariance between wind speed and density fluctuations at shallow depths (Fig. 6f), as
520 well as between current speed and density (Fig. 6g). This suggests that winds drive surface waters away or along the ice-
521 shelf front toward open water. Sea ice formation through latent and sensible heat loss to the atmosphere then leads to brine
522 rejection and explains the increase in shallow layer density. During the subsequent events in July 2020 and February 2021,
523 the anomalously warm period at shallow levels associated with the advecting features between 400 and 800 m depth is not
524 observed (Fig. 8b/c). We therefore interpret the anomalously warm period in the layer closest to the ice base following the
525 first event as a wind-driven, localized anomaly, likely facilitated by the open water surface to the north of Thwaites Pinning
526 Point (Fig. 1a).

527 In July 2020, we observed a subsurface feature without any associated warming in the layer closest to the ice base (Fig.
528 8b). Unlike the April 2020 event, the currents at shallow depth were directed toward the SSW, indicating that the observed
529 feature was advected from the NNE beneath TEIS (Fig. 6b). There is significant covariance between wind speed and density
530 fluctuations at shallow depths on timescales exceeding one month (Fig. 6f), and between current speed and density (Fig. 6g),
531 suggesting that winds drove the formation of this feature. Unfortunately, this period is not covered by the Aquadopps at
532 Cavity Camp, but the Aquadopps at Channel Camp confirm significant covariance between current speed and density
533 fluctuations at shallow levels but not at depth (Fig. 6g/j). The DTS record at Channel Camp captured most of this event,
534 showing warming between 400 and 600 m depth and cooling between 600 and 800 m in early July (Fig. 7b). However, the
535 DTS record ends in early August 2020, before the event concluded (Fig. 8b). We interpret this event as being driven by wind
536 stress toward the NNE, where PIB remained covered by mobile, first-year sea ice (Fig. 10e) to transmit the prolonged wind
537 forcing into the ocean.

538 In February 2021, we captured the clearest event occurring between 400 and 800 m depth (Fig. 8c). Similar to the July
539 2020 event, shallow currents were directed toward the SSW (Fig. 6b). However, unlike July 2020, current speed variability
540 at 316 m depth did not significantly co-vary with ERA5 wind speed (Fig. 6f) or with density variability at this depth. This
541 suggests that the near-isothermal layer, observed between 300 and 400 m depth (Fig. 7c), likely formed due to turbulent
542 mixing, independent of the deeper event. At mid-depth, and within the warming part of the water column (400–600 m),
543 currents flowed toward the SSE (Fig. 6h). At greater depths, within the cooling part of the feature (600–800 m), currents

shifted from SSE at mid-depth to SSW at depth (Fig. 6d). Current variability at Cavity Camp influenced density fluctuations on timescales of up to a month (Fig. 6i), with an even clearer signal at Channel Camp (Fig. 6j). During this period, PIB remained covered by first-year sea ice (Fig. 10e), while open water areas with mobile sea ice were present in northern PIB. We therefore suggest that the captured features in February 2021 as well as July 2020 originated from this area before advecting beneath TEIS.

In summary, the analysis of current directions and speeds suggests that the source region of the events lies to the NE of TEIS.

4.2 Conditions during immobile, multi-year fast ice cover

After May 2021, no further events were observed at mid-depth (Fig. 10k). During this period, the second-year sea ice in PIB reached its maximum extent, becoming fastened between the ice edge of TEIS to the west and Antarctica's coastline to the east (Fig. 10). This fast-ice platform stretched over 150 km from Thwaites Pinning Point to the grounding line of Pine Island Glacier. The extensive, immobile fast ice effectively isolated the ocean from atmospheric wind stress. Hydrographic data reveal an increasing meltwater content at both shallow and mid-depth levels until July 2021 (Fig. 5a/b). This observation aligns with the findings of Zheng et al. (2022) and Dotto et al. (2022), who suggest that prolonged fast ice coverage in PIB facilitates the accumulation of ice shelf meltwater beneath the sea ice cover, extending beneath TEIS. This meltwater likely originates from a combination of subshelf melting beneath TEIS and melting along the deep grounding lines of Thwaites and Pine Island Glacier. The resulting meltwater-enriched plumes rise through the water column due to their relative buoyancy, reaching shallower layers. Unfortunately, all Aquadopp current meters malfunctioned during this period, preventing a determination of the source region for these water masses, but the numerical model tracer tracking results shown by Dotto et al. (2022) demonstrated that such a flow from the ice shelves upstream of Thwaites is feasible.

4.3 Fast-ice breakout and increased WW advection

The retreat of the fast ice edge began at the end of the austral summer in January 2022 (Fig. 10e), when a significant portion of multi-year fast ice in northeastern PIB broke up, exposing open water (Fig. 10e/f). During the following winter, surface cooling by the atmosphere likely allowed WW to recharge in this open-water region, contributing to the observed cooling in the upper half of the water column within the TEIS cavity (Fig. 9c). However, whether WW originated specifically from this newly exposed area or was supplied by enhanced advection of a colder WW variety remains uncertain, as both processes could explain the observed cooling in our DTS record and WW properties change both from year to year, and spatially.

Evidence supporting WW advection, rather than cooling driven by meltwater-enriched water masses, comes from the shallow MicroCAT, which indicates a concurrent decrease in the mCDW-derived meltwater content toward the WW mixing line in late 2021 (Fig. 5a). Another possible explanation for the cooling is increased subglacial outflow, but grounding-line discharge is typically associated with lower salinity and minimal change in potential temperature (Davis et al., 2023). Given these factors, we conclude that enhanced WW advection is the most likely cause of the observed cooling.

576 4.4 Potential formation mechanisms of the observed events

577 Our results support the narrative of Zheng et al. (2022) that variability in subshelf oceanography is influenced by sea ice
578 conditions in PIB. The novelty of our study lies in the finding that different sea ice types correspond with and may lead to
579 characteristic signatures in the subshelf water column. Mobile unconfined sea ice generates surface stress on the ocean,
580 driving circulation similar to wind forcing on open ocean water (Fig. 11a/b). Strong winds in PIB lifts mid-depth isopycnals
581 and facilitates the formation of gyre-scale features (tens of kilometers) which are subsequently advected beneath TEIS,
582 altering the thermal structure between 400 m and 800 m depth over several weeks. In contrast, we hypothesize that when PIB
583 is covered by persistent, near-stationary, or landfast multi-year sea ice, the transfer of wind stress into the ocean is inhibited
584 (Fig. 11c), which may prevent the formation and advection of these features. An extended duration of fast ice coverage leads
585 to overall warmer conditions beneath TEIS (Dotto et al., 2022) and the accumulation of meltwater in the upper ocean layers,
586 driven by sub-ice-shelf melting and buoyant meltwater from the deep grounding lines of Pine Island and Thwaites Glaciers
587 (Fig. 11d). As the sea ice edge retreats landward, colder WW is advected beneath the ice shelf in the upper layers, while
588 variability in mCDW at depth occurs primarily on tidal timescales (Fig. 11e), contrasting with the longer variability observed
589 under sea ice-covered conditions. We hypothesize that after the fast ice breakout in January 2023, when data collection
590 ended, the mid-depth features reappear as sea ice and ocean conditions continue to interact with wind forcing.

591 Different types of sea ice play a significant role in shaping the oceanographic variability beneath TEIS, supporting
592 the ideas presented by Zheng et al. (2022) and Dotto et al. (2022). The main distinction between this study and that of Dotto
593 et al. (2022) is the availability of a longer oceanographic record that captures changes in hydrographic properties as sea ice
594 cover in PIB evolves, along with a more extensive use of the DTS dataset to examine the vertical extent and timing of
595 changes within the subshelf cavity. While Zheng et al. (2022) and Dotto et al. (2022) suggested that a cyclonic PIB gyre lifts
596 isopycnals in PIB, causing them to sink beneath TEIS and resulting in colder conditions, our study reveals a delayed,
597 contrasting response at depth. We observe warming between 400 and 600 m depth and cooling between 600 and 800 m
598 following an active, cyclonic PIB gyre. The PIB gyre spans approximately 50 km and transports around 1.5 Sv of water,
599 reaching depths of about 700 m (Thurnherr et al., 2014). Considering the gyre's depth range, our observed events are centred
600 around 600 m depth, which may explain the upward displacement of isopycnals above this level. However, the mechanism
601 responsible for the opposing effect at greater depths remains an open question and requires further investigation within a
602 numerical modelling framework.

603 Recent numerical simulations of the Amundsen Sea suggest that the ice-shelf cavities beneath Thwaites and Pine Island
604 Ice Shelves are favourable environments for submesoscale eddies ($O(0.1\text{--}10\text{ km})$, $O(1\text{ day})$; Shrestha et al., 2024). These
605 eddies transport heat vertically toward the ice shelf base, potentially enhancing basal melting in a positive feedback loop
606 (Shrestha et al., 2024). However, identifying their formation mechanisms remains challenging due to the lack of direct
607 observations within the ice-shelf cavity. We anticipate that our dataset will help constrain these mechanisms. The features
608 we observe, however, exhibit larger horizontal and temporal scales ($O(10\text{--}100\text{ km})$, $O(1\text{ month})$) and a greater vertical extent

609 (O(100 m)) than the O(10 m) submesoscale eddies simulated by Shrestha et al. (2024). Additionally, while their modelled
610 eddies formed behind bathymetric sills at depth, lifting mCDW upward, our observed features display an opposing signal,
611 centred around 600 m depth, temporarily pushing mCDW downward.

612 Fluctuations in thermocline depth, where temperatures rapidly increase from 0 °C to +1 °C, separating the cold WW
613 above from the warm mCDW below, have been linked to wind stress variations over the open ocean in PIB (Webber et al.,
614 2017). These fluctuations have been associated with changes in basal melt rates beneath Pine Island Ice Shelf on a similar
615 timescale to the features we observe (O(1) month, Davis et al., 2018). While wind stress primarily drives isopycnal
616 displacement within the thermocline, where vertical density gradients are strongest, this mechanism produces a uniform
617 response throughout the water column and does not explain the opposing trends we observe, which instead manifest as
618 periodic thickening centred around 600 m depth.

619 Mooring observations near the front of Getz Ice Shelf have shown that WW deepening beyond 550 m is associated
620 with strong easterly winds and reduced sea ice cover, originating about 100 km from the mooring site. This process
621 generates intra-layer waves that propagate toward the ice shelf, temporarily cooling the water by 1-2 °C at 586 m depth over
622 O(10) day-timescales (Steiger et al., 2021). Because our events exhibit warming between 400 and 600 m depth, this cooling
623 mechanism directly contradicts our observations, ruling out these waves as the driving force behind the observed features.
624 However, non-local pumping may have contributed to the increased advection of WW between July 2021 and January 2023,
625 during which the upper half of the water column beneath TEIS cooled by 1.2 °C (Fig. 9c).

626 4.5 Implications

627 Our results highlight the oceanographic variability beneath TEIS which implies a need for improved basal melt
628 parameterizations in coupled ice-ocean models. The observed events consistently advect at around 600 m depth (Figs. 8 and
629 9), increasing water temperatures between 400 to 600 m depth and potentially enhancing basal melting in regions where ice
630 thickness reaches similar depths, such as along the deep grounding lines of Pine Island and Thwaites Glaciers. By lifting
631 isopycnals closer to the ice-shelf base, these events contribute to localized warming beneath the ice-shelf base and they may
632 accelerate basal melt, with near surface layers potentially continuing to warm in the weeks following an event at greater
633 depths (Fig. 8a).

634 Simple depth-dependent melt parameterizations often overestimate heat and salt exchange at the ice-ocean interface,
635 leading to unrealistic projections of grounding-line retreat (Seroussi et al., 2017), and would miss the dynamic events
636 described here. While even the most advanced models, such as those used in Naughten et al. (2023), provide sophisticated
637 representations of basal melting beneath West Antarctic ice shelves, including 3D ocean circulation, sea-ice interactions, and
638 atmosphere-ocean fluxes, they still rely on quasi-steady parameterizations of ice-ocean interactions (e.g., the three-equation
639 formulation; Holland & Jenkins, 1999). As a result, they may underrepresent transient processes like those observed in this
640 study. Our field data suggest that changes in sub-ice shelf circulation can occur on shorter timescales than those typically

641 resolved in these models. Thus, while coupled models include the major physical components, they may not yet capture the
642 episodic and fine-scale variability in ocean forcing and melt response revealed by high-resolution observations.

643 Since sea ice formation, presence and motion play a crucial role in redistributing heat, salt, and momentum, its impact
644 on basal melt rates beneath neighbouring ice shelves and the deep grounding lines must be accounted for. Our findings
645 emphasize the importance of incorporating oceanographic processes that link evolving ocean conditions to ice-shelf
646 melting (Yu et al., 2018). Observations, such as the dataset presented in this study, provide essential constraints for refining
647 coupled ice-ocean models and improving projections of Thwaites Glacier's future evolution and the potential collapse of
648 WAIS.

649 **5 Conclusion**

650 Our measurements revealed coupled atmosphere-ice-ocean interactions that could only be captured using the AMIGOS-3
651 system, which was designed to track long-term water mass movements throughout the water column as PIB sea ice coverage
652 evolved. We observed distinct events linked to open ocean conditions or during mobile sea ice cover, where mid-depth
653 waters warm while waters near the seabed temporarily cool over a few weeks. Under a closed fast ice cover in PIB, these
654 events disappear, allowing deep water from Thwaites Trough to penetrate beneath the TEIS. This water mass competes with
655 warmer waters from PIB, which extend far westward reaching beneath TEIS. However, when the fast ice edge retreats across
656 PIB, these competing water masses diminish at depth and upper level waters cool substantially through the increased
657 advection of WW. This highly dynamic system likely influences the basal melting of Thwaites Glacier and other glaciers
658 draining into the Amundsen Sea.

659 The recent decline in Antarctic sea ice, marked by more extreme annual fluctuations, suggests that the events we
660 observed may become more frequent as sea ice coverage continues to decrease. Reduced sea ice will not only provide less
661 insulation from atmospheric variability but may also allow atmospheric forcing to penetrate even deeper into the water
662 column than previously recognized, influencing the variability of mCDW near the seabed.

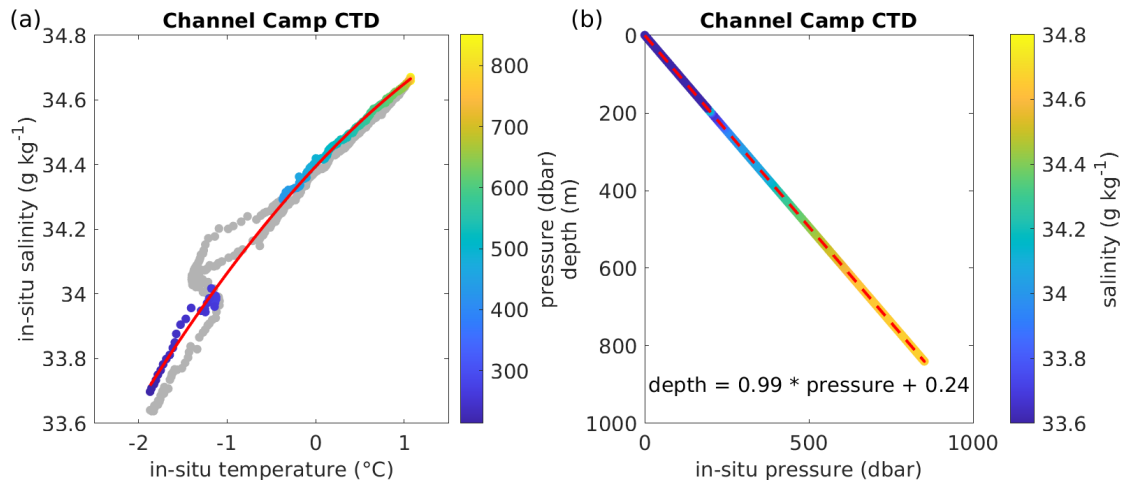


Figure A1: CTD cast at Channel Camp. (a) Relationship between in-situ temperature and salinity from CTD profiling on January 12, 2020. Coloured dots indicate the data points used to derive a polynomial fit (red curve), excluding the thermocline to reflect long-term averages. (b) Relationship between in-situ pressure and depth below the ocean surface, with the linear fit shown as a dashed red line. Note the transition from freshwater in the borehole to saltwater in the ocean cavity around 200 m depth.

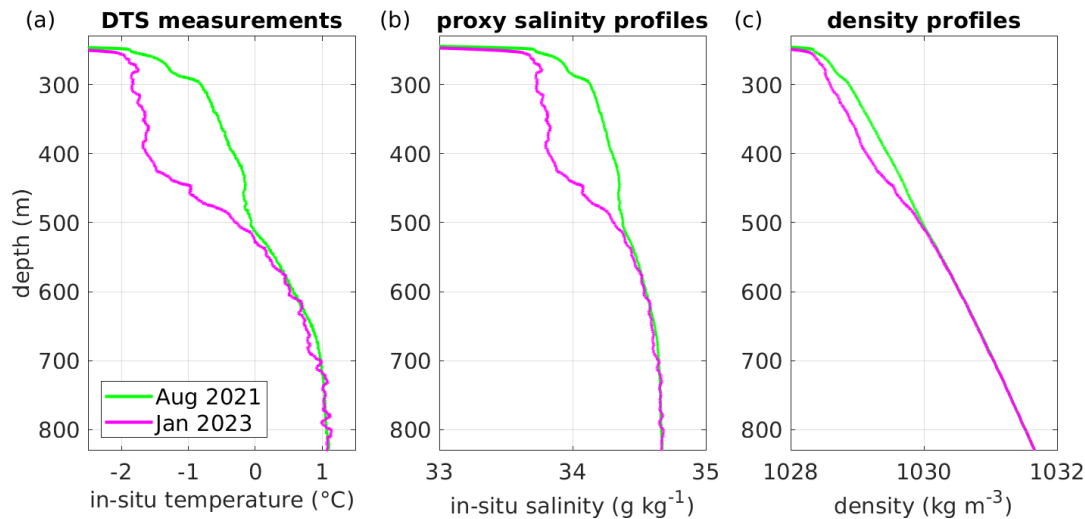
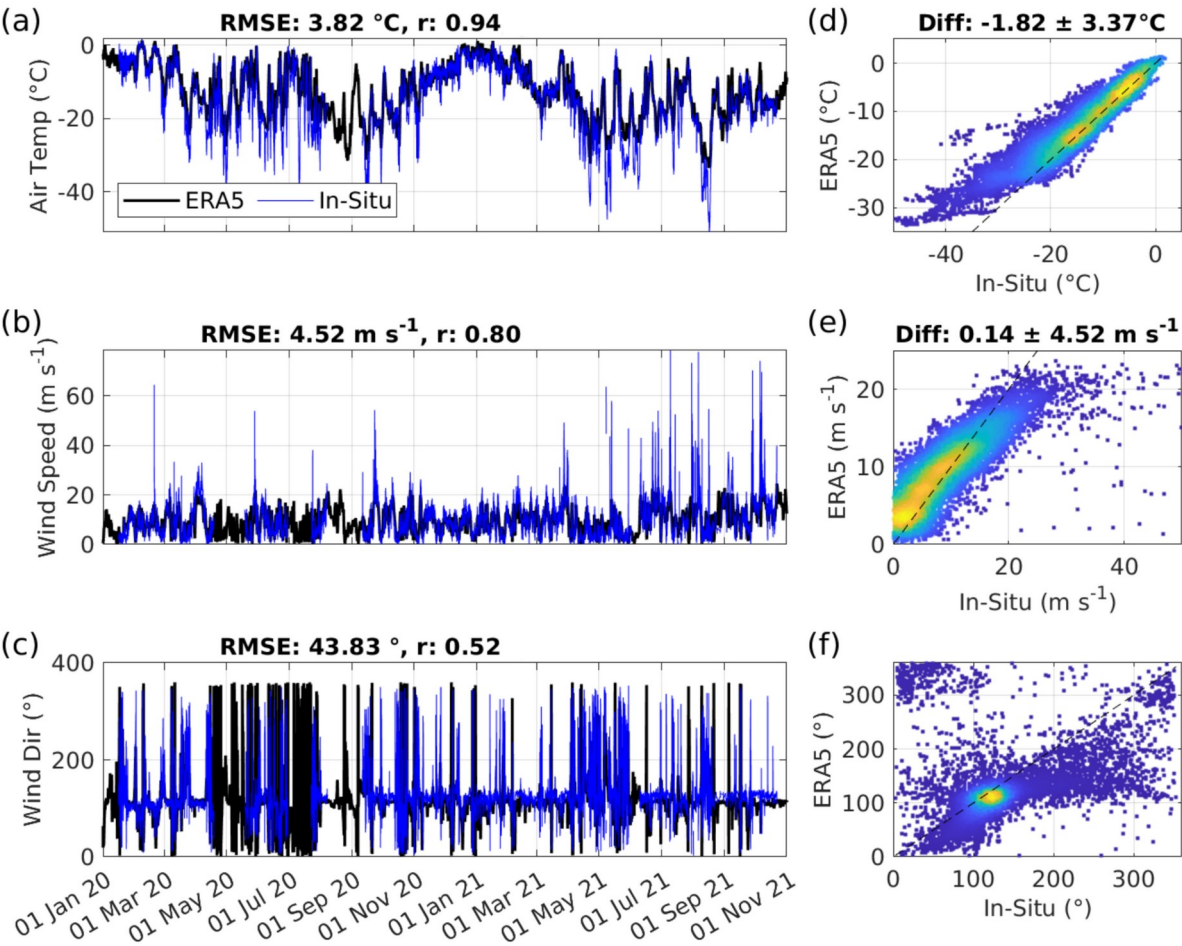


Figure A2: Cooling beneath Channel Camp. (a) In-situ temperature profiles. (b) In-situ salinity profiles derived from the polynomial relationship established during CTD profiling on January 12, 2020 (see Fig. A1). (c) Corresponding seawater density profiles. Note the freshening observed in the upper half of the water column.



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Figure B1: AMIGOS-3 versus ERA5: (a-c) Time series of air temperature, wind speed, and wind direction showing available in-situ data. (d-f) Scatter plots showing the relationship between the in-situ and ERA5 data for each variable, with colours indicating point density, where warmer colours correspond to higher point density. Mean differences and their standard deviations are calculated as in-situ data minus ERA5.

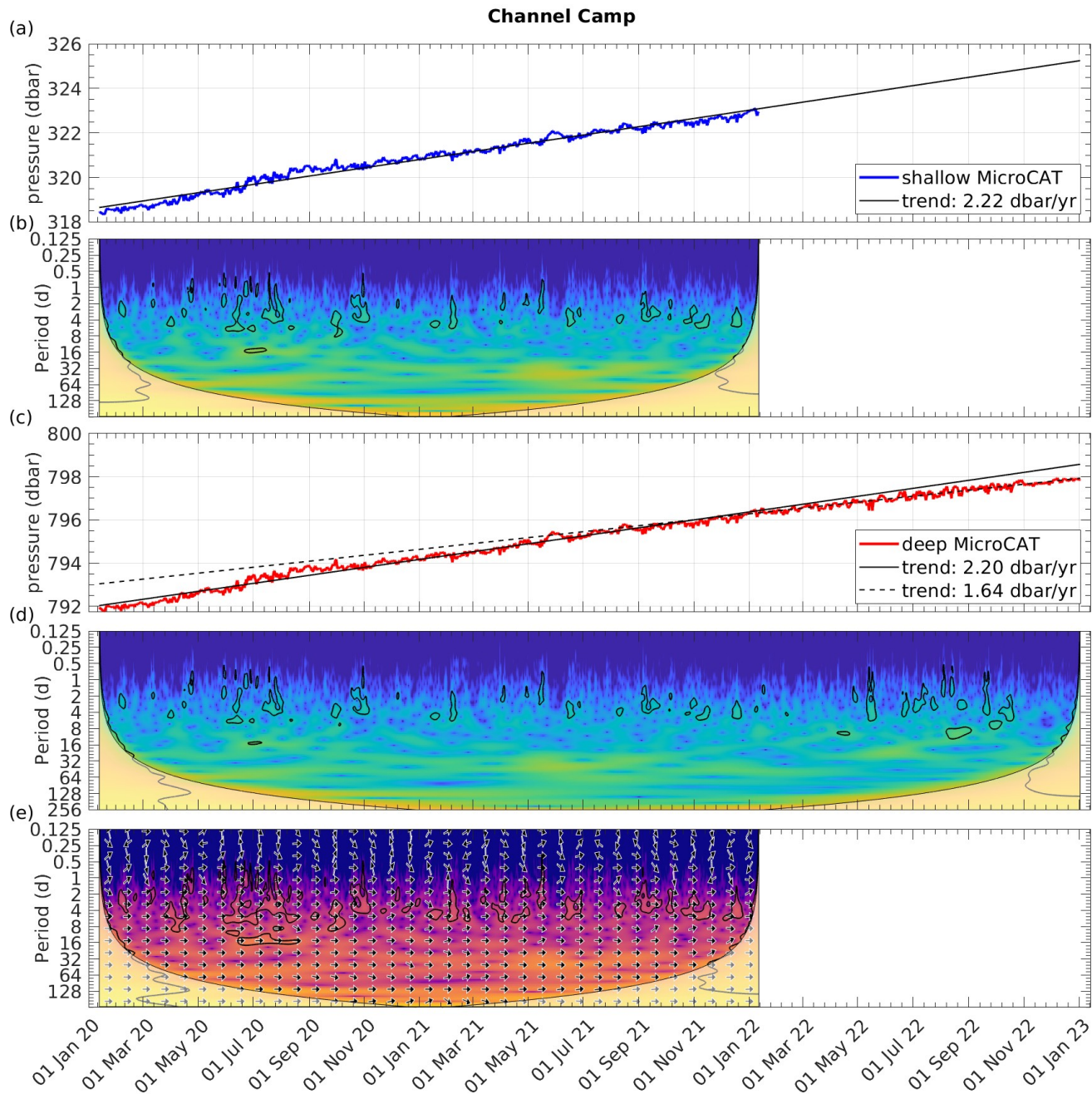


Figure C1: Pressure records from Channel Camp for (a) shallow levels and (c) near the seabed. Panels (b) and (d) display the continuous wavelet transforms of the two time series. Panel (e) shows the cross-wavelet transform between the two pressure records for their overlapping time period.

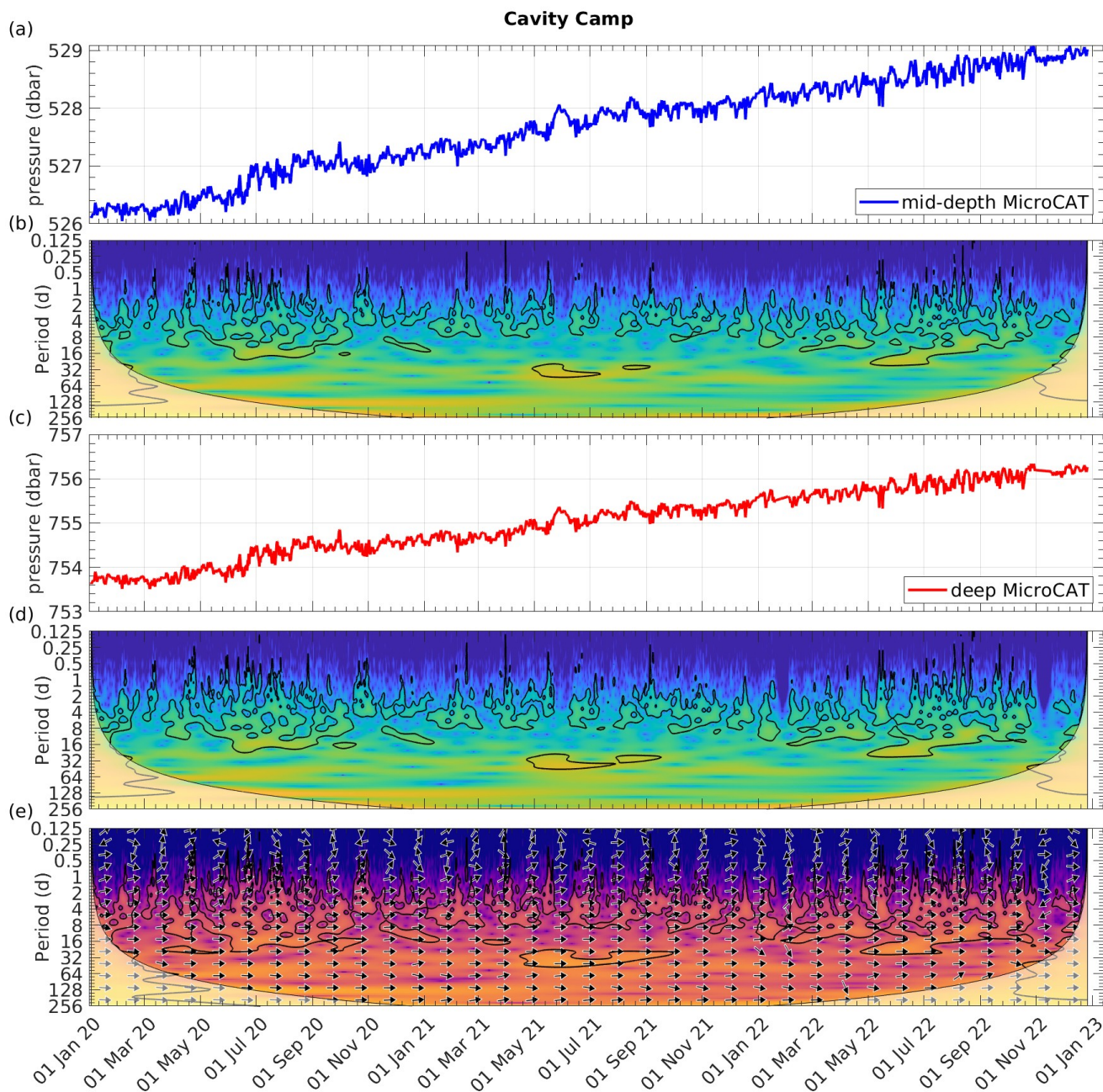


Figure C2: Pressure records from Cavity Camp for (a) mid-depth levels and (c) near the seabed. Panels (b) and (d) display the continuous wavelet transforms of the two time series. Panel (e) shows the cross-wavelet transform between the two pressure records for their overlapping time period.

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692 **Code availability**

693 Python code for retrieving daily sea ice concentration can be found at https://github.com/tsnow03/thwaites_amigos.

694 The MATLAB Gibbs-SeaWater (GSW) Oceanographic Toolbox is available from <http://www.teos-10.org/>. MATLAB
695 software for wavelet analysis can be found at <https://github.com/grinsted/wavelet-coherence>.

696 **Data availability**

697 The AMIGOS-3, borehole CTD and DTS data from Cavity Camp and Channel Camp are available from the United States
698 Antarctic Program Data Center (USAP-DC) at <https://www.usap-dc.org/view/project/p0010162>. The ship-based CTD
699 dataset from February 2019 is available at [https://www.bodc.ac.uk/data/published_data_library/catalogue/10.5285/e338af5d-](https://www.bodc.ac.uk/data/published_data_library/catalogue/10.5285/e338af5d-8622-05de-e053-6c86abc06489/)
700 [8622-05de-e053-6c86abc06489/](https://www.bodc.ac.uk/data/published_data_library/catalogue/10.5285/e338af5d-8622-05de-e053-6c86abc06489/). Autonomous underwater vehicle data are available at
701 <https://researchdata.se/en/catalogue/dataset/2020-193-1>. ERA5 reanalysis data are available from
702 <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-single-levels?tab=overview>. Sentinel-1 imagery is available from
703 the Copernicus Open Access Hub (<https://scihub.copernicus.eu/>). The sea ice concentration dataset is available from the
704 University of Bremen, at https://data.seaice.uni-bremen.de/amsr2/asi_daygrid_swath/s3125/.

705 **Author contributions**

706 CTW conceived the study, led data analysis, produced the figures and drafted the manuscript. TS provided sea ice
707 concentration time series and contributed to writing. TSD contributed to the discussion of hydrographic properties. SWT
708 calibrated the DTS data and processed the MicroCAT data. TAS developed the AMIGOS-3 system. ECP led field work,
709 collected and processed borehole CTD casts. ECP and KJH are principal investigators of the TARSAN project. All authors
710 discussed results, implications, edited text and approved of the final manuscript.

711 **Competing interests**

712 Karen J. Heywood serves as Co-Editor-in-Chief of *Ocean Science*.

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914 **Supplementary materials**

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