

# Turbulent Dynamics of Buoyant Melt Plumes Adjacent Near-Vertical Glacier Ice

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## Key Points:

- Robotic observations at a submerged near-vertical iceberg face capture turbulent dynamics of buoyant melt plumes and background currents
- Buoyant plumes extend 20-50 cm from the boundary, undulate on 100-s periods, and drive horizontal turbulent transports.
- Buoyant plumes can be more effective than horizontal flows in energizing boundary layer turbulence and heat flux.

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**Abstract**

At marine-terminating glaciers, the interplay between meltwater buoyancy and local currents control turbulent exchanges. Because of challenges in making centimeter-scale measurements at glaciers, turbulent dynamics at near-vertical ice-ocean boundaries are poorly constrained. Here we present the first observations from instruments robotically-bolted to an underwater ice face, and use these to elucidate the tug-of-war between meltwater-derived buoyancy and externally-forced currents in controlling boundary-layer dynamics. Our observations captured two limiting cases of the flow. When external currents are weak, meltwater buoyancy energizes the turbulence and dominates the near-boundary stress. When external currents strengthened, the plume diffused far from the boundary and the associated turbulence decreases. As a result, even relatively weak buoyant melt plumes are as effective as moderate shear flows in delivering heat to the ice. These are the first *in-situ* observations to demonstrate how buoyant melt plumes energize near-boundary turbulence, and why their dynamics are critical in predicting ice melt.

**Plain Language Summary**

Melting glaciers are projected to produce several inches of sea level rise over the next few decades. Despite this threat, the fundamental fluid dynamics which drive melt at tidewater glaciers remain poorly characterized. This is primarily attributed to challenges associated with measuring the temperature and velocity of ocean water at the submerged cliffs of actively calving glaciers. To this end, we have developed a robotically-deployed instrument that can be bolted to a glacier's face. This instrument is capable of measuring temperature and kinetic energy of ocean waters within a few inches of the ice, representing the first measurements of their kind. Our observations demonstrate the ways in which meltwater at ice boundaries can accelerate melt. In particular, the meltwater tends to be less salty (and hence lighter) than the nearby ocean waters (which are salty, warm and heavy), so the meltwater rises along the ice face, creating an energetic, near boundary flow. With our new measurements, we show that these flows are as important as large-scale currents in providing energy to the ice to fuel melt. We anticipate these data will help our community create more accurate models of ice melt needed to predict the advance or retreat of marine ice cliffs of Greenland, Alaska and Antarctica.

## 1 Introduction

Directly quantifying the rate of ice-melt at the near-vertical cliffs of marine-terminating glaciers is a challenge due to the boundary's inaccessibility to traditional forms of sampling. The ice melt-process is also complicated because the thermodynamics depend on how local buoyancy production (from melt) combines with the external forcing (temperature  $T$ , salinity  $s$ , velocity  $\vec{u}$ ) to control energy flow across the ice boundary. The heat flux  $j_q$ , for example, is ultimately set by strong thermal gradients near the diffusive scales, which provides buoyant energy that fuels the turbulent energy cascade. In turn this cascade also intensifies the near-boundary thermal gradients that drive melt and supply buoyant energy (Fig. 1A). This feedback loop is at odds with traditional turbulence theory that often assumes isotropy, homogeneity and the ability to separate the spatial scales of energy sources (large) and sinks (small).

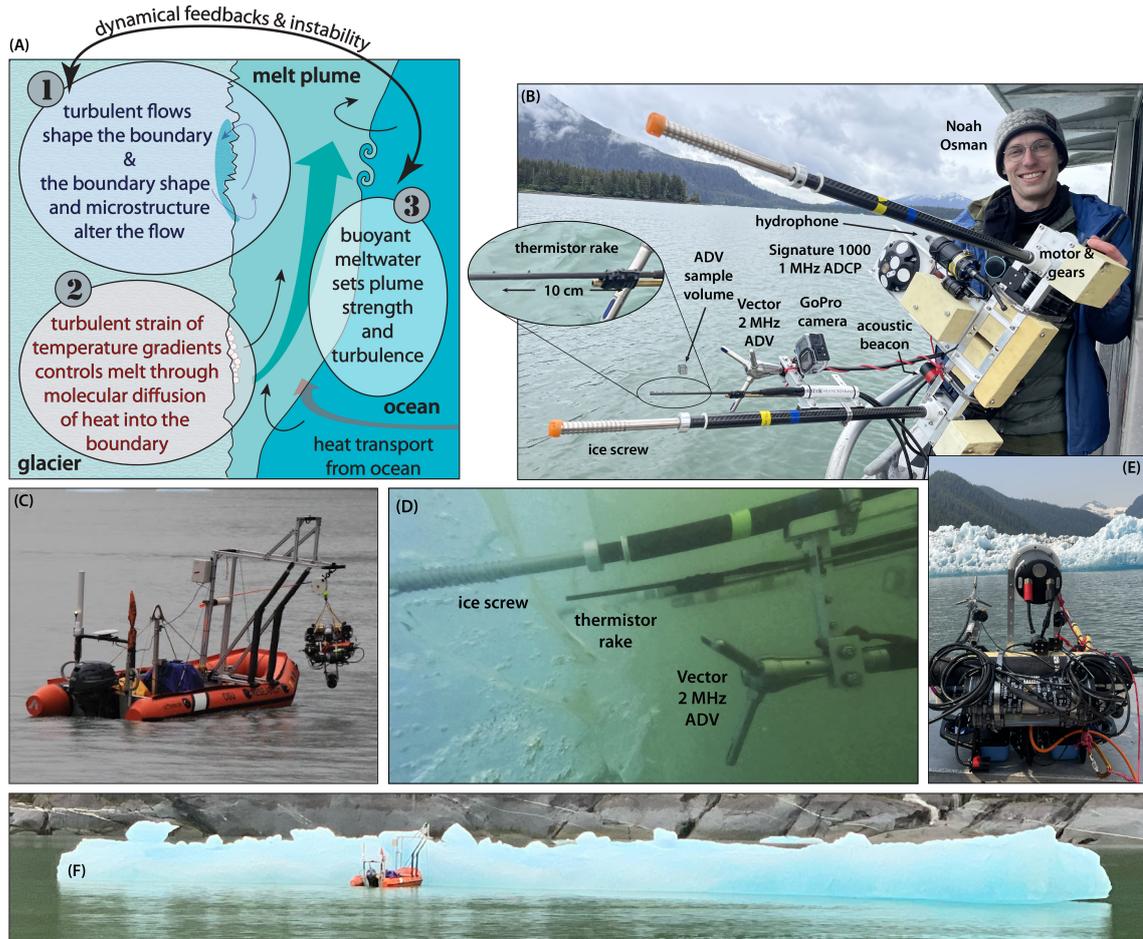
At an ice face, we hypothesize that meltwater detaches from the boundary in fine-scale turbulent sweeps, similar to those observed under sea ice (Fer et al., 2004) and in atmospheric boundary layers (Kline et al., 1967), but here producing buoyant energy at the same small scales that dominate viscous dissipation. Meltwater buoyancy thus injects additional momentum at very small scales – near the viscous tail of a downscale turbulent energy cascade likely fueled both by the large-scale buoyant forcing (Xu et al., 2013) alongside a zoo of classic ocean- (Garrett, 1989) and fjord-specific (Bendtsen et al., 2021) turbulence sources. Further complicating the dynamics are the energy exchanges as parcels entrain buoyancy from the background stratification as they move vertically against gravity, which can represent either a source or sink of energy (Magorrian & Wells, 2016; Kimura et al., 2014).

Beneath gently-sloping, near-horizontal ice-ocean interfaces, meltwater buoyancy drives along-ice flow. However, this buoyancy also provides static stability, so turbulent exchanges primarily occur through hydrodynamic instability such as Kelvin-Helmholz billows (Smyth, 1999). In 1984, a comprehensive set of observations of turbulent melt dynamics in the Marginal Ice Zone of the Greenland Sea were acquired (McPhee et al., 1987). MCPhee et al. (1987) used these data to create an empirical model to predict melt from  $T$ ,  $s$ , and the turbulent stress  $\tau$ , which formed the basis for the canonical three-equation melt parameterization (Holland & Jenkins, 1999). Because the stability of the ice permitted detailed, high-accuracy measurements to be obtained, this parameterization (based on ice-melt thermodynamics and three empirical coefficients derived from those experiments), remains the community's primary and only way to predict melt beneath ice shelves (Jenkins et al., 2010) if the relevant flow ( $\vec{u}$ ,  $T$  and  $s$ ) can be prescribed.

77 As the ice interface approaches vertical, meltwater can generate sufficient buoyancy to be-  
78 come convectively unstable and directly energize turbulence, as demonstrated in the laboratory  
79 by Josberger and Martin (1981). Because the entrainment of warm ocean water increases with  
80 plume strength, the melt process creates a positive feedback (Fig. 1A) that further energizes the  
81 plume to enhance melt. Eckert and Jackson (1950) created a framework for characterizing tur-  
82 bulent free-convection boundary layers adjacent to a heated plate (for air at  $Pr = 1$ ), and their  
83 study remains highly relevant today (Parker et al., 2021). However, the ice-melt problem is more  
84 complex because: (1) turbulence is generated by both buoyant convection and shear production  
85 (Josberger & Martin, 1981; Zhao et al., 2024); (2) melt can be driven by both salinity or thermal  
86 gradients, each which diffuse and influence density in different ways (Gade, 1979; Kerr & Mc-  
87 Connochie, 2015); (3) vertical gradients of ocean properties (such as density) affect buoyancy  
88 production of turbulent energy and the growth of turbulent plumes (i.e., Magorrian and Wells (2016));  
89 and (4) in addition to buoyancy, other sources of velocity like internal waves (Cusack et al., 2023)  
90 or mean currents (Jackson et al., 2020; Zhao et al., 2023) affect shear production of turbulent en-  
91 ergy.

92 Theoretical models (i.e. Wells and Worster (2008)) provide a framework to describe plume  
93 evolution, but still require turbulence closure derived from laboratory experiments (McConnochie  
94 & Kerr, 2017), numerical experiments (Gayen et al., 2016), or observational analogies (McPhee  
95 et al., 1987). At the geophysical scale, empirical models have been developed that assume sim-  
96 plified geometries and turbulence closure. For example, Jenkins (2011) used the framework of  
97 MacAyeal (1985) to couple buoyant plume theory with the 3-equation melt model (McPhee et  
98 al., 1987) to predict the downstream flow evolution. By prescribing an idealized plume geom-  
99 etry, this framework has been used to predict the freshwater distribution from a localized sub-  
100 glacial discharge (Cowton et al., 2015; Carroll et al., 2016) and also for distributed melt (Magorrian  
101 & Wells, 2016; Jackson et al., 2020).

102 To date, there are no experiments analogous to the 1984 sea-ice observations (McPhee et  
103 al., 1987) that could be used to constrain the flow and meltrate parameterization for a vertical ice  
104 face. In addition to uncertainty in values of drag and transfer coefficients, there is also debate about  
105 how to formulate the coupled models themselves. Part of this debate stems from observations  
106 of glacier face ablation (Sutherland et al., 2019) and the existence of large-scale meltwater in-  
107 trusions (Jackson et al., 2020) that imply significantly higher meltrates than predicted with the  
108 above theories as applied in their commonly-used forms. It has been suggested that the bound-  
109 ary layers are energized by external currents which increases the turbulent transfer coefficients;



**Figure 1.** (A) Cartoon illustrates the interplay between (1) ice morphology (2) turbulent and molecular transports across the ocean-ice interface, and (3) melt-driven buoyant plumes that energize the boundary layer. (B) Meltstake sensors are configured to measure these dynamics with minimal disturbance to the flow. (C) Remotely-operated vessel and winched ROV. (D) Meltstake as deployed 12:40 29 May showing ice structure and the sensors' proximity to the interface; the ADCP is outside the frame of view. (E) A Meltstake riding atop the delivery ROV on deck; iceberg from Xeitl Síť in background. (F) Remote deployment in progress.

110 anecdotal evidence suggests this is not unreasonable, i.e., Cusack et al. (2023); Slater et al. (2016);  
 111 Jackson et al. (2020). Other factors – like energy from exploding air bubbles observed in the lab  
 112 (Wengrove et al., 2023), or ice roughness and channelization observed beneath ice shelves (Stanton  
 113 et al., 2013; Watkins et al., 2021) – may also be at play here. It is the purpose of this note to de-  
 114 scribe the first detailed observations of the turbulent flow at a near-vertical glacier-ice face, and  
 115 to demonstrate how plume buoyancy and external velocities contribute to melt-dynamics. A con-  
 116 current paper (Weiss et al, in prep) will extend the analysis of these data to quantify melt rates  
 117 and assess bias and uncertainty in current melt parameterizations.

## 2 Methods

### 2.1 The Glacier Meltstake

The Meltstake is a submarine device (Fig. 1B-F) that is remotely bolted to a near-vertical glacier-ice face to directly measure melt and the spatial structure of near-boundary velocity, temperature and turbulence. It is called a “Meltstake” in analogy to the subaerial ablation stakes used by glaciologists to measure ice accumulation and ablation in the field. It is designed to be a stable platform to observe the flow in a reference frame fixed to the ice and in ways that minimize the system’s thermal and hydrodynamic impact on melt dynamics.

The body of the Meltstake is suspended outward from the ice on two, 61-cm long carbon fiber tubes, chosen for their mechanical stiffness and low thermal conductivity  $5 \times 10^{-6} \text{m}^2/\text{s}$  (Macias et al., 2019). Ice screws mounted on the ends of these 16-mm diameter tubes turn using Blue Robotics T200 motors at 23:1 reduction. Each screw-assembly rotates within a 25-mm carbon sheath to allow instruments to be rigidly attached at various distances from the ice. A Raspberry Pi “brain” controls drilling power, schedule and underwater communications with a remotely-operated vehicle (ROV) through a long-range 28 kHz Delphis Subsea modem.

The Meltstake is transported to the ice face using a BlueRobotics ROV, equipped with a Ping360 imaging sonar and video camera for underwater navigation. The Meltstake is pinned to the ROV and held in place with a Newton linear actuator. The ROV can be deployed from either a robotic vessel equipped with a remotely-operated winch or a traditional vessel. High-power Ubiquiti Rocket WiFi allows remote operation of the ROV/Meltstake from several kms away using the standard QGroundControl software. Acoustic messages sent from the ROV trigger drill operations. Once the the Meltstake “bites” into the ice, the ROV releases from it. The freed ROV can then monitor the meltstake, request further manual drilling, update the Meltstake’s autonomous schedule, or request its release to return to the surface. The unit is rated to 100-m depths, ballasted 10 N buoyant, and has a flasher and GPS/satellite beacon for recovery.

### 2.2 Experimental setting and measurements

Boundary-layer measurements were made at a freely-floating iceberg with 10-m draft,  $\sim 20$  km down-fjord from Xeitl Síť (also called LeConte Glacier) in Southeast Alaska. We deployed the Meltstake on a vertical face of the ice at 6.5-m depth starting at 20:40 UTC, May 29, 2023. As the iceberg melted, we sent acoustic “drill” commands (at 21:39 and 23:05) to advance the Meltstake and move sensors closer to the ice interface. At 23:48 it was released and recovered.

149 At 00:46, May 30 it was again delivered to the same iceberg at 8.5-m depth, drilled further at 01:46,  
150 and released at 02:20.

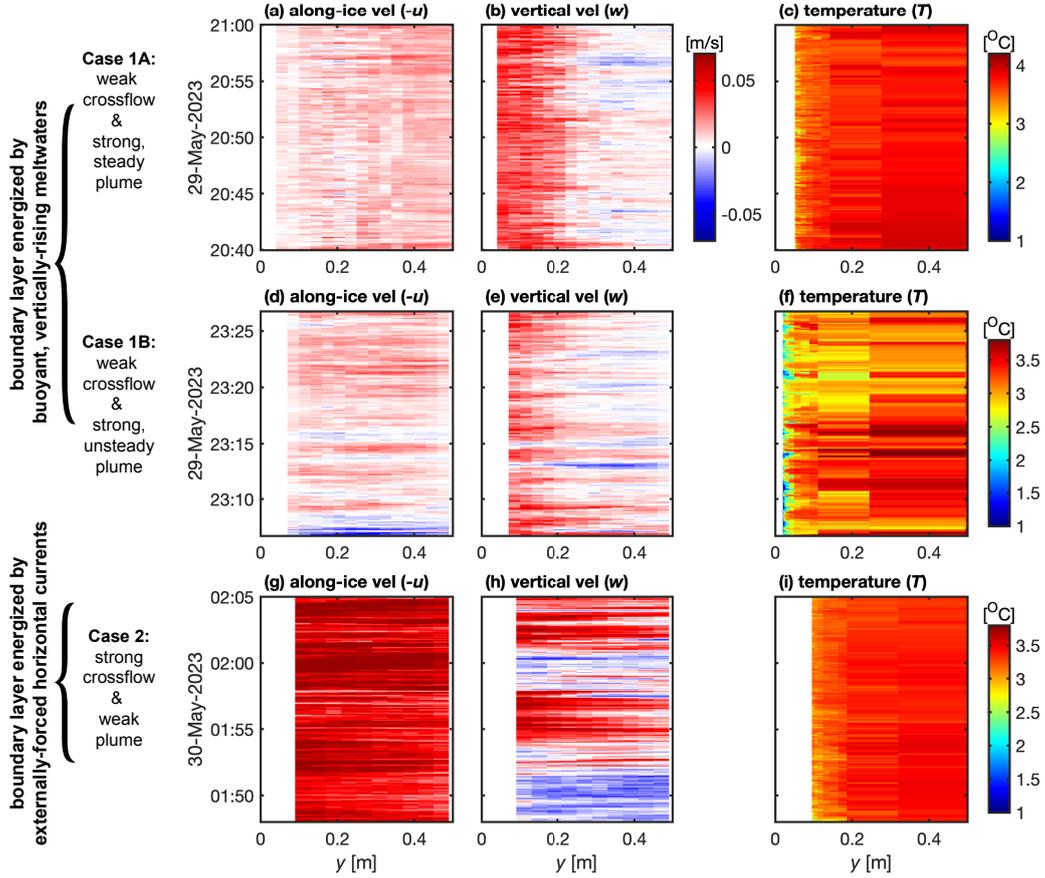
151 Velocity was imaged with a 5-beam Nortek 1000 kHz Acoustic Doppler Current Profiler  
152 (ADCP) in pulse-coherent mode (4 Hz sampling with 3-4 cm bins). Because of high acoustic backscat-  
153 ter from ice, ADCP data are contaminated by spurious reflections from sidelobes at ranges that  
154 exceed the distance of the closest transducer-to-ice distance. We thus attempted to orient the ADCP  
155 so that the 4 slant beams encounter the ice at approximately the same range. We use a right-hand  
156 coordinate system in which  $x$  is along-ice,  $y$  is horizontal and positive away from the ice, and  
157  $z$  is up. ADCP data were recorded in along-beam coordinates and used for two purposes: (1) op-  
158 posing beams were combined to determine the bulk vertical ( $w$ ) and along-ice ( $u$ ) velocity over  
159 the 10-70 cm footprint of the spreading beams; (2) along-beam velocities were used to compute  
160 (i) the velocity  $v$  from the central beam and (ii) turbulent statistics of the flow using the struc-  
161 ture function method of Wiles et al. (2006) as implemented by Thomson et al. (2016). Echo backscat-  
162 ter from a Nortek Vector Acoustic Doppler Velocimeter was used to determine meltrate.

163 Near-boundary temperature was measured using a thermistor “T-rake,” a horizontal array  
164 of eight, fast-response thermistors, each exposed downward into the expected flow at distances  
165 of 2, 4, 7, 12, 23, 39, 58 and 84 mm from the tip of a carbon tube (Fig. 1B, S1 and supplement).  
166 Three fast-response RBR Solos provided additional temperature measurements at 10, 35 and 60  
167 cm from the ice. Salinity was obtained from nearby vertical profiles using a RBR Concerto CTD  
168 and ranged from 27.4-28.4 within the  $\pm 1$ -m depth range around each deployment.

### 169 **3 Observations of buoyancy- and externally-forced boundary layers**

170 Here we examine three time periods that illustrate the range of flow patterns observed (Fig. 2).  
171 The first two cases represent a boundary layer energized by the vertical rise of buoyant meltwa-  
172 ter, which we term “buoyancy-forced.” The third is an example we term “externally-forced,” be-  
173 cause horizontal velocities were significantly stronger than those of the vertically-rising flows.

174 **Case 1A: Quasi-steady buoyant plume.** Shortly after the Meltstake was deployed (at 6.5 m depth,  
175 and under weak,  $u \sim 1$  cm/s, crossflow conditions), a quasi-steady plume was observed to flow  
176 vertically up the ice at 2-4 cm/s within  $\sim 20$  cm of the ice (Fig. 2a-c). During this time, the strongest  
177 temperature anomalies (indicative of melt waters) were only observed by sensors within a few  
178 millimeters of the T-rake tip, and  $\sim 5$  cm from the ice.



**Figure 2.** Horizontal, along-ice velocity (a,d,g), vertical velocity (b,e,h), and temperature (c,f,i) within 0.5 m of the ice interface for three twenty-minute periods. Between Case 1A and 1B (at 6.5 m depth), the Meltstake was advanced 6 cm further into the ice, placing the thermistor rake within 2 cm of the ice, but also increasing ADCP sidelobe contamination; Case 2 was a separate deployment at 8.5 m depth. Distance from ice ( $y$ ) was computed acoustically for  $-u$  and  $w$  (Weiss et al., 2024) and using equation 3 to determine  $y_o$  for  $\hat{T}$ ; note that the ice melted 3-5 mm during each 20-min period (Weiss et al., 2024), so we treat  $y$  independent of time for these plots.

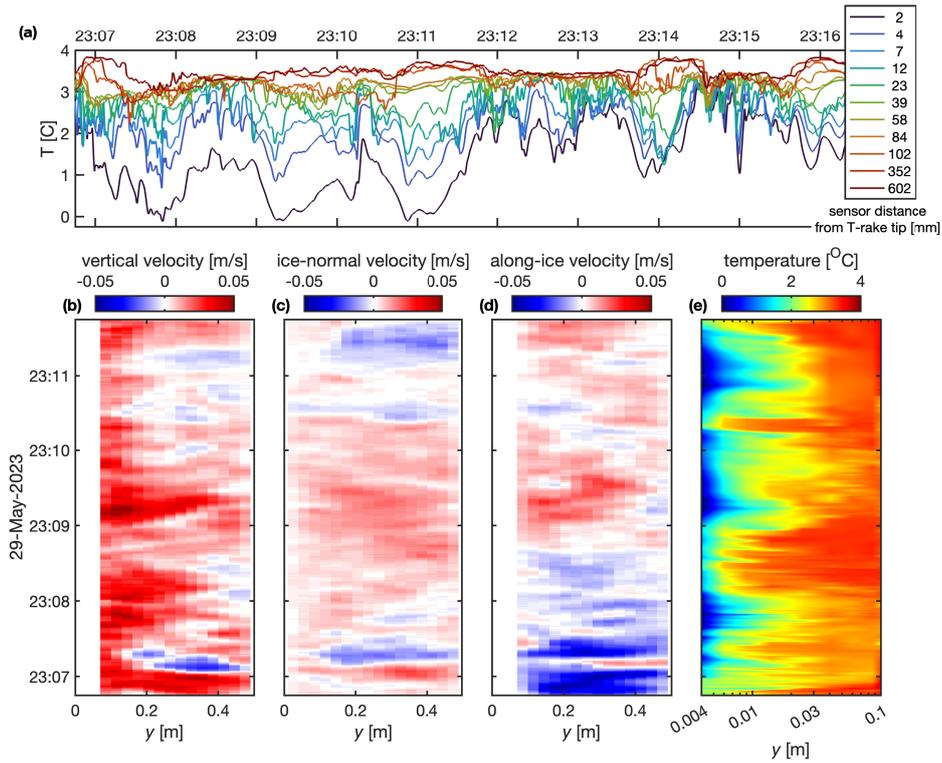
179 **Case 1B: Strongly-undulating buoyant plume.** As time evolved the buoyant plume became  
 180 more variable in time, weakened in magnitude, and decreased in thickness (Fig. 2d-f). The cross-  
 181 flow also became slightly unsteady (but still weak), undulating with similar timescales as the ver-  
 182 tical plume. The Meltstake was also advanced towards the ice between 1A and 1B, yielding  $T$   
 183 observations within 2 cm from the ice. Temperatures most distant from the ice were observed  
 184 to increase slightly, and pulses of low-temperature waters were swept 2-10 centimeters from the  
 185 ice, contrasting the weaker thermal anomalies in case 1A. Far from the boundary ( $y > 10$  cm),  
 186  $w$  alternates sign on  $\sim 100$  second intervals; these pulses appear correlated with temperature.  
 187 For example, between 23:14 and 23:18 there are several strong vertical velocity reversals that co-  
 188 incide with warm pulses, which could be interpreted as turbulent eddies drawing warm ambient  
 189 fluid towards the boundary.

190 **Case 2: Strong crossflow.** After the Meltstake was released and re-drilled into the ice at 8.5-m  
 191 depth, the iceberg had moved and tidal flows strengthened, exposing the ice to stronger currents  
 192 (Fig. 2g-i). At this time,  $-u$  averaged 6 cm/s,  $w$  was highly variable but upward ( $\sim 1-1.5$  cm/s)  
 193 on average, and both undulated with O(5 min) period;  $-u$  and  $w$  are correlated and somewhat  
 194 out-of-phase (the weakest  $w$  generally correspond to the largest  $-u$ ). Temperature anomalies (in-  
 195 dicating the presence of meltwater) were observed close to the boundary.

#### 196 **4 Character of turbulence in the buoyant plume**

197 To glean insight into turbulent dynamics energized by meltwater buoyancy, we examine  
 198 the undulating plume case (Case 1B) in more detail. We focus on the 5-10 minutes following drilling  
 199 (at 23:05) and we look in detail at the 11 individual thermistors in the context of the near-boundary  
 200 velocities (Fig. 3). During the first 5 minutes, the T-rake was in closest proximity to the ice (see  
 201 supplement), such that the innermost thermistor (2 mm from the T-rake tip; midnight blue in Fig.  
 202 3a) was on average 4 mm from the ice.

203 These temperature data demonstrate a turbulent melt-and-extrude cycle, whereby the first  
 204 phase of the eddy draws warm water toward the boundary to initiate melt, and the second phase  
 205 sweeps the cold meltwaters away from the ice. This pattern can be seen in the traces in Fig. 3a:  
 206 at times when  $T$  rises at the outer sensors, temperatures at the inner sensors cool. For example,  
 207 at 23:11, 23:14 and 23:16, the 3 outer sensors (red traces in Fig. 3a) warm together, while the in-  
 208 ner five sensors (blue-green traces) cool in unison. These cold pulses – which reached as low as  
 209  $0^\circ$  C at times – are the signatures of melt emerging from the boundary. Following these (i.e., at



**Figure 3.** Details of the boundary-layer layer illustrate the dynamics of the unsteady plume: (a) Ten-minute segment of temperature data from the 11 sensors used in Fig. 2f. Lower panels show a zoom-in on the first five minutes of that record on May 29, 2023: (b) vertical velocity, (c) ice-normal velocity (positive/red is away from the ice), (c) along-ice velocity, and (d) temperature, plotted against logarithmic distance coordinates to highlight the smallest scales near the ice boundary. In (c),  $v$  is from the ADCP center beam so is least-contaminated by acoustic sidelobes and provides unbiased data almost to the ice surface.

210 23:12) are periods in which the temperature of all sensors coalesce together, and are the times  
 211 when warm waters make their closest contact to the ice, presumably temporarily enhance melt.

212 Anomalies exceeding  $\sim 1.5$  °C (below ambient  $T_a$ ) were detected 25 mm from the bound-  
 213 ary, and coherent across all sensors, indicating a pathway for meltwater to be swept out from the  
 214 laminar sublayer into the outer layer by turbulence. During these events, the ice-perpendicular  
 215 velocity (Fig. 3c) was directed away from the face at approximately 1 cm/s, extended 10s of cm  
 216 from boundary, and varied coherently in all three velocity components. This cycle of perturba-  
 217 tions – that brings warm water towards the ice and extrudes cold meltwaters away from the bound-  
 218 ary – undulates on 100-s periods, and is the signature of a horizontal eddy-transport of heat that  
 219 fuels melt.

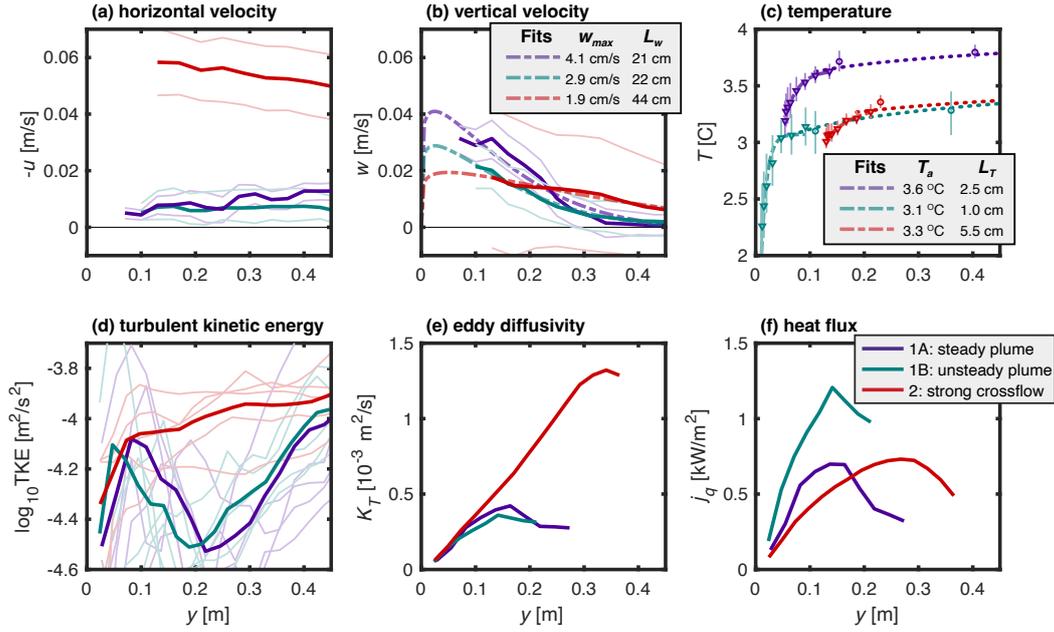
## 220 5 Quantitative differences in flow patterns

221 To compare the flow characteristics during each of the example time periods, we compute  
 222 mean profiles of the near-boundary velocity, temperature, turbulent energy and heat transport (Fig.  
 223 4). Fits of  $w$  and  $T$  to empirical functions are used to determine spatial scales, magnitudes and  
 224 gradients, which we use to determine  $\tau$  and  $j_q$ , both of which are important parameters to pre-  
 225 dict melt. Consistency between direct turbulence observations and  $\tau$  derived from Eckert and Jack-  
 226 son (1950)'s self-similar profiles provides confidence in our interpretations.

227 **Velocity:** For a convection flow driven by buoyancy from a heated vertical plate, Eckert and Jack-  
 228 son (1950) derived similarity solutions for a Prandtl number ( $Pr = 1$ ) flow. They find the ver-  
 229 tical velocity  $\hat{w}$

$$\hat{w}(\hat{y}) = w_1 \hat{y}^{1/7} (1 - \hat{y})^4. \quad (1)$$

230 is a function of the nondimensional distance from the wall  $\hat{y} = y/\delta$ , where  $\delta$  is assumed to vary  
 231 slowly in  $z$  and represents the distance over which the solution is valid ( $\hat{w} \geq 0$ );  $w_1$  is a con-  
 232 stant. We use this form to characterize the observed plumes' vertical velocity  $w(y)$  by minimiz-  
 233 ing  $\sum (w(y) - \hat{w}(y))^2$  to determine  $w_1$  and  $\delta$  over 20-minute durations. For this solution, the  
 234 peak velocity is  $w_{max} = 0.5372w_1$  and the plume width, defined by  $\hat{w}(L_w)/w_{max} = 1/e$  is  
 235  $L_w = 0.304\delta$ . As shown in Fig. 4(b), these fits represent the data well in the region we have  
 236 observations, and indicate a factor-of-two increase in plume width ( $L_w = 44$  cm) during pe-  
 237 riods of strong crossflow compared to that during weak ( $L_w = 21 - 22$  cm). Because  $w_{max}$   
 238 decreased for large  $L_w$ , the total vertical transport,  $Q_{plume} = \int_0^\delta \hat{w} dy = 0.146w_1\delta$  was ob-  
 239 served to be similar for each of the three cases: 76, 56 and 78 cm<sup>2</sup>/s.



**Figure 4.** Mean and turbulent characteristics of the observed boundary layers: (a) along-ice velocity  $-u$ , (b) vertical velocity  $w$ , (c) temperature  $T$ , (d) turbulent kinetic energy TKE, (e) turbulent diffusivity  $K_T$ , and (f) heat flux  $J_q$ . Each colored line represents a 20-minute average over the time periods shown in Fig. 2: steady plume (1A, purple), undulating plume (1B, turquoise), strong crossflow (2, red). Thin/light lines in (a-c) define the central 50% of the data. Gaps in (c) separate data from the temperature rake and RBR Solos (separated horizontally by 60 cm and hence responsible for offsets in  $T$ ). Light dashed lines in (b) and (c) represent eqns. 1, 3 with least-square-fit coefficients as indicated; in (c) fits to eq. 3 use the T-rake data (shown in thick dashed lines) and fits to eq. 2 use the outer 5  $T$  sensors (thin dashed lines). In (d), semi-transparent lines represent estimates of TKE from each of the 5 individual ADCP beams (heavy lines are the means).

240 **Temperature:** T-rake timeseries provide temperature and its gradient with sub-centimeter res-  
 241 olution and at close proximity to the ice boundary. Here we use these and Solo data to charac-  
 242 terize the thermal boundary layer (see Supplement for details), which we separate into an outer  
 243 and inner layer.

244 We begin by considering Eckert and Jackson (1950)'s similarity solution, for which the char-  
 245 acteristic lengthscale for  $T(y)$  and  $w(y)$  assumed the same ( $\delta$ ). In their form (applicable to air  
 246 ( $Pr = 0.7$ ) and requiring  $T = 0$  at the boundary), a substantial temperature gradient ( $O \sim 1\text{C/m}$ )  
 247 is predicted far from the boundary, which is not observed here (Fig. S2). Here we modify their  
 248 form by introducing  $\Delta T$  to allow for a lesser temperature drop (relative to ambient  $T_a$ ) in the outer  
 249 layer:

$$T = T_a - \Delta T(1 - (y/\delta)^{1/7}). \quad (2)$$

250 Fits to the outer 5 temperature measurements are roughly consistent with both this form and the  
 251 logarithmic scaling presented by Tsuji and Nagano (1988) (see Supplement), yielding a 0.2-0.3  
 252 °C drop in the outer boundary layer.

253 Close to the ice, the observed  $T(y)$  is inconsistent with (Eckert & Jackson, 1950) eq. (2).  
 254 Motivated by the early work of Smith (1972) and Tsuji and Nagano (1988), we consider an in-  
 255 ner layer shaped by molecular transports and having a different characteristic lengthscale  $L_T$ ,  
 256 and arbitrarily assume the following exponential form:

$$\hat{T}(y) = T_a - (T_a - T_i)e^{-y/L_T}. \quad (3)$$

257 Here we assume the ice temperature  $T_i = 0^\circ \text{C}$  and solve for  $T_a$ , the ambient (farfield) tem-  
 258 perature,  $L_T$ , the decay scale, and  $y_o$ , the T-rake offset by minimizing  $\sum_{n=1}^8 (T(y_n) - \hat{T}(y_n))^2$   
 259 for each of the  $n$  thermistors.  $T_a$  and  $L_T$  are shown Fig. 4(c);  $y_o$  was 5.4, 1.0 & 13 cm for cases  
 260 1, 2 & 3. The melt-plumes' thermal lengthscales ( $L_T = 1 - 4 \text{ cm}$ ) are a factor of ten smaller  
 261 than  $L_w (= 20 - 40 \text{ cm})$ ; like  $L_w$ ,  $L_T$  is largest during periods of strong crossflow. The con-  
 262 sequences of these differences are evident in the mean temperature profile (Fig. 4c and supple-  
 263 ment), where two length scales also emerge: one that controls visco-diffusive transports and shapes  
 264 the inner boundary layer ( $L_T$ ), and a second that characterizes energetic turbulent transports in  
 265 the outer boundary layer and diffuses (reduces) larger-scale gradients of  $T$  for  $y > 10 \text{ cm}$ .

266 **Turbulence:** Of relevance to ice melt is the near-boundary TKE, which we compute from along-  
 267 beam structure functions (Wiles et al., 2006) (Fig. 4d). We employ this technique because it does

268 not depend on relationships between acoustic beams, and hence relaxes assumptions of spatial  
 269 homogeneity. While TKE is relatively uniform in the strong crossflow (red line), it increases to-  
 270 wards the boundary (with a maximum at  $\sim 10$  cm) for both periods when melt-plume veloci-  
 271 ties dominated the KE. This suggests a different source of TKE in each case: shear production  
 272 during the strong crossflow vs. buoyancy production when the external flow weakened.

273 We calculate the horizontal turbulent heat flux as  $j_q = \rho c_p K_T dT/dy$  where  $\rho$  and  $c_p$  are  
 274 the density and heat capacity of seawater,  $K_T$  is the turbulent diffusivity and  $dT/dy$  the back-  
 275 ground temperature gradient. We estimate  $K_T \approx \kappa u' \ell$ , where  $\kappa = 0.4$  is von Karman's co-  
 276 efficient,  $u' \approx \sqrt{\text{TKE}}$ , and  $\ell$  is the lengthscale of the energy-containing eddies. In analogy to  
 277 Perlin et al. (2005), we modify the canonical law-of-the wall scaling (for which  $\ell$  is the distance  
 278 to the boundary) by limiting the characteristic lengthscale far from the boundary to be that of the  
 279 plume's eddies, which we approximate as  $w/(dw/dy)$ . Based on these *law-of-the-wall* modi-  
 280 fications and using Eckert and Jackson (1950)'s model (equation 1) to estimate plume eddy size,  
 281 we find  $\ell = \max(y, \hat{w}/(d\hat{w}/dy))$ , which increases linearly ( $\ell = y$ ) for  $y < 0.75L_w$  and then  
 282 decreases almost linearly to 0 at  $\ell = 3.3L_w$ .  $K_T$  is found to have similar magnitude and struc-  
 283 ture for all three cases;  $j_q$  is about twice as high for the unsteady plume as the other 2 cases. Note  
 284 that  $j_q = 1 \text{ kW/m}^2$  is equivalent to 1 cm/hour of ice melt.

285 Eckert and Jackson (1950)'s formulation also provides a convenient way to compute the  
 286 vertical stress at the ice boundary

$$\tau_w = 0.0225 \rho w_1^2 \left( \frac{\nu}{w_1 \delta} \right)^{1/4} \quad (4)$$

287 and has been found consistent with lab and numerical simulations of turbulent flow from a ver-  
 288 tically oriented source of distributed buoyancy (Parker et al., 2021; Zhao et al., 2024). We find  
 289  $\tau_w$  which is 0.0098 and 0.0053 Pa for the strong plumes, two to five times larger than  $\tau_w = 0.0022$   
 290 Pa for the case of a cross-flow. For comparison, the stress associated with the horizontal flow (as-  
 291 suming  $\tau_u = \rho C_d u^2$  with  $C_d = 2 \times 10^{-3}$ ) is 0.0072 Pa, similar to that of  $\tau_w$  in the plumes;  
 292  $\tau_u$  is roughly  $30\times$  smaller during weak crossflow.

## 293 **6 Interpretation**

294 Much of what we have learned about melt comes from limiting cases of weakly-turbulent  
 295 laboratory experiments (Josberger & Martin, 1981; McConnochie & Kerr, 2017), idealized nu-  
 296 merical modelling (Gayen et al., 2016; Zhao et al., 2024), measurements under horizontal sea

297 ice (McPhee et al., 1987), or inferences from farfield observations (Jackson et al., 2020). A re-  
298 maining challenge is understanding the connections between outer turbulent scales and molec-  
299 ular transports across a real ice interface, i.e., the exchanges of buoyancy, heat and momentum  
300 are fueled by dynamics sketched in Fig. 1A that have until now been largely studied in isolation  
301 or under idealized settings.

302 Our observations of iceberg-scale boundary layers are thicker and more energetic than those  
303 simulated in the lab or modelled numerically. Here, rising currents and their turbulence extend  
304 20-50 cm from the ice, contrasting the 1-10 cm lateral scales in simulated flows. And while the  
305 strongest temperature anomalies (a proxy for melt buoyancy) are confined within a 1-4 cm  $e$ -folding  
306 distance from the ice, the heat transport extends far from the boundary. Qualitatively, this is ev-  
307 idenced by the sweeps in  $T$  (figure 3), driven by eddies that cyclically advect warm waters to-  
308 ward the boundary and extrude meltwater across the plume on  $\sim 100$  sec timescales. These ed-  
309 dies are responsible for the turbulent heat flux  $j_q$  (Fig. 4f).

## 310 **7 Conclusions**

311 Recent observations of thick meltwater intrusions (Jackson et al., 2020) and unexpectedly  
312 high frontal ablation rates (Sutherland et al., 2019) have led to suggestions that Holland and Jenk-  
313 ins (1999) and Jenkins (2011)’s models need to be revisited. Some have suggested transfer co-  
314 efficients need to be modified (Jackson et al., 2020), others have suggested we need a new em-  
315 pirical model (Schulz et al., 2022), constrained by observations, that is “physically plausible,”  
316 but not physics based. Neither approach is particularly satisfying because they require arbitrary  
317 tuning of coefficients to match observations. The details of the physics are important.

318 Here we demonstrate the ways in which meltwater buoyancy energizes near-boundary tur-  
319 bulence adjacent to a near-vertical section of an iceberg originating from Xeitl Sít’ glacier. Im-  
320 portantly, when external sources of mechanical energy are weak, buoyant convection becomes  
321 dominant, driving vertical flows that enhance near-boundary turbulence. While these “meltwa-  
322 ter plumes” varied in character, their mean structure was well-described by fits to various pow-  
323 erlaw and exponential functions, and provide a means of quantifying scales of the flow.

324 While the character of real ambient melt plumes is similar to that predicted by theory (Wells  
325 & Worster, 2008), lab (Josberger & Martin, 1981) or numerical simulation (Gayen et al., 2016),  
326 the natural flows we observe are significantly more energetic. For example, the sole laboratory  
327 study to measure temperature within a turbulent boundary layer adjacent vertical melting ice (Josberger

328 & Martin, 1981) found turbulent temperature fluctuations to be confined within 2-10 mm of the  
329 ice, with fluid outside that layer being quiescent and only occasionally being entrained towards  
330 the boundary. In contrast, the boundary layer flows observed here are stronger, broader, and pro-  
331 duce higher heat fluxes than these idealized studies.

332 Our observations confirm that meltwater buoyancy can energize turbulence in the ice-adjacent  
333 boundary layer as effectively as a moderate external flow, plausibly driving similar melt rates in  
334 both cases. But what sets the TKE,  $j_q$  and controls the melt rate? While idealized studies provide  
335 some insight and intuition, the feedbacks that control melt cannot be determined from the local  
336 dynamics alone. For example, we have shown that a flow – forced ostensibly by the same exter-  
337 nal conditions – can have dramatically different character (compare Fig. 2 panels a-c with d-f).  
338 We hypothesise that the interplay between externally-driven turbulence and meltwater convec-  
339 tion is critical to the flow dynamics: both shear and buoyant production influence the coherent  
340 structures that are of first order importance of turbulent exchange across this boundary layer. Fur-  
341 ther direct observations that capture the phenomenology of *real* melt-driven boundary-layers and  
342 elucidate the range of dynamical possibilities are critical to inform the next generation of exper-  
343 iments and parameterizations.

## 344 8 Open Research

345 All data are available at the National Snow and Ice Data Center <http://nsidc.org/>

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