



Calving response to lake thermal regime at an ice sheet margin in west Greenland

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Abstract. Ice-marginal lakes are increasingly widespread around the Greenland Ice Sheet (GrIS), but their influence on ice dynamics remains poorly constrained. Elsewhere, recent work has shown that ice-contact lakes are more thermally complex than previously thought, with potential implications for rates of melt undercutting and calving losses. Here, we assess the relationship between lake thermal regime and calving frequency at a lake-terminating outlet glacier in west Greenland using in situ observations collected during the summers of 2024 and 2025. Lake water temperature varied across the summer season, exhibiting a distinct warm phase during August characterised by a positive thermal gradient and brief period of stratification. Calving activity was dominated by waterline and collapse events, with event frequency peaking during the warm lake phase. Statistical modelling indicates that this was also the period when lake water temperature and calving frequency were most strongly associated. These findings suggest that intraseasonal changes in lake thermal structure may modulate calving behaviour, likely through enhanced melt undercutting. Overall, we highlight the potential for thermally driven calving to become an increasingly important mechanism of ice-loss at lake-contact margins of the GrIS with continued climate warming, and thus lake warming.



1 Introduction

Ice-marginal lakes have grown in number and extent over recent decades, forming within topographic basins revealed during ongoing glacier retreat (Shugar et al., 2020). These lakes can accelerate glacier flow, enhance thinning, and promote terminus recession, especially by modifying subglacial water pressures and facilitating calving (Warren and Kirkbride, 2003; Tsutaki et al., 2013, 2019; King et al., 2019; Pronk et al., 2021). Glaciers that transition from land- to lake-terminating margins may therefore become partially decoupled from climate-driven atmospheric forcing alone, with rates of mass loss increasingly influenced by thermo-mechanical processes at the ice-water interface (Benn et al., 2007; Sutherland et al., 2020; King et al., 2019; Sato et al., 2022). Indeed, ice discharge via calving has been shown to account for more than 30% of total mass loss from some lake-terminating glaciers in Patagonia and the Himalaya (Tsutaki et al., 2011; King et al., 2019; Minowa et al., 2021) and can accelerate inland velocities by effectively de-buttressing the terminus (Pronk et al., 2021).

Calving in freshwater settings occurs via several mechanisms, including enhanced melt of the submerged ice face, with melt rates typically greatest at the waterline. Thermal erosion promotes the formation of a notch which undercuts, and subsequently destabilises, the overlying ice cliff, leading to subaerial calving (Röhl, 2006; Minowa et al., 2017). This process can also leave a protruding ice foot at depth vulnerable to detachment via buoyant forces (Benn et al., 2007; Sugiyama et al., 2019). Observations from New Zealand (Kirkbride and Warren, 1997), Patagonia (Minowa et al., 2026), Sweden (Dye et al., 2025), Alaska (Caldwell et al., 2025) and the Himalaya (Watson et al., 2020) have shown that the style, frequency and magnitude of calving events are sensitive to a range of lake-specific factors (Benn et al., 2007). Alongside bathymetry, water depth, and ice cliff height, lake water temperatures have been identified as a potentially significant control on calving regime, especially where they exceed 1°C (Warren and Kirkbride, 2003; Haresign and Warren, 2005; Minowa et al., 2017).

Recent work has demonstrated that ice-marginal lakes can be substantially warmer and more thermally complex than previously thought. In situ measurements from high mountain regions (Sugiyama et al., 2016, 2021; Scofield et al., 2025) show water temperatures peaking in summertime, when a positive thermocline separates relatively warm surface waters from cooler, dense waters at depth. Although deep water temperatures appear moderately stable, surface temperatures are influenced by changes in insolation and wind speed, fluctuating over both diurnal and intraseasonal timescales (Sugiyama et al., 2016; Minowa et al., 2017; Watson et al., 2020; Scofield et al., 2025; He et al., 2026). This variability in thermal regime has implications for rates of frontal ablation. Comparisons of lake temperature with glacier geometry and calving behaviour have revealed a positive association between temperature and ice loss, driven in part by the rate-controlling effect of temperature on thermal undercutting (Warren and Kirkbride, 2003; Minowa et al., 2017; Sugiyama et al., 2019; Dye et al., 2025). However, this relationship remains poorly constrained because measurements of the thermal conditions of ice-marginal lakes are sparse, and datasets which present coincident records of calving are rarer still.



This paucity of in-situ data is especially true for the lacustrine margins of the Greenland Ice Sheet (GrIS), where ice-marginal lakes are increasingly widespread and their influence on mass loss is expected to grow over the coming decades (Carrivick et al., 2022; How et al., 2025). Understanding the processes defining the rate of this ice loss is important, not least because the GrIS is amongst the largest contributors to contemporary sea level rise (Edwards et al., 2021). However, calving is particularly difficult to parameterise and without direct observations of lake thermal conditions and their effect on frontal ablation, lake-ice interactions cannot be meaningfully realised in ice sheet models used to project future change (Carrivick et al., 2020, 2022).

Lake-terminating outlet glaciers of the GrIS have retreated more rapidly, and exhibited faster velocities, than their land-terminating counterparts in recent years (Mallalieu et al., 2021; Harpur et al., 2026). In the only in situ study of calving at a GrIS lake-contact margin to date, Mallalieu et al. (2020) showed that once winter lake ice cover dispersed in early summer, calving was dominated by events at the waterline and larger collapses. This regime was attributed to melt-undercutting and thermo-erosional notch development, which progressively weakened the ice cliff and represented the primary mechanism of margin retreat. The presence of a waterline notch implied thermal stratification of lake waters but, without concurrent measurements of water temperature, the implications for calving were inferred, rather than empirically demonstrated. Although recent remote sensing observations have indicated that surface water temperatures vary considerably between lakes at the margin of the GrIS and are warming over time (Carrea et al., 2025; How et al., 2025), profile measurements, i.e. throughout the water-column, remain limited.

This study aims to characterise the influence of lake thermal conditions on calving regime at a lake-terminating outlet glacier of the GrIS. Using lake temperature measurements and timelapse camera imagery acquired during two consecutive summer seasons, we examine relationships between lake thermal variability and the frequency and style of calving events. These calving metrics are further evaluated alongside meteorological conditions and modelled surface melt.

2 Methods

2.1 Study site

Our field data are predominantly derived from an unnamed 1.75 km² ice-marginal lake and associated outlet glacier in South-West Greenland, ~100 km inland of the coast and 447 m above sea level (67.06°N, -50.03°W; Fig. 1). The lake features two topographically overdeepened basins, each hosting a distinct calving front separated by an exposed bedrock sill. We focus on the western basin (~0.37 km²), where bathymetric surveys conducted in 2024 record a maximum lake depth of 41 m, decreasing to between 8 and 25 m adjacent to the ice front (Fig. 1b). The calving face is 452 m wide, laterally constrained by topography, and has recently retreated at an average rate of 26 m yr⁻¹ (2016 to 2025; calculated using the Google Earth Digitisation and Margin Change Quantification tools (Lea, 2018)).

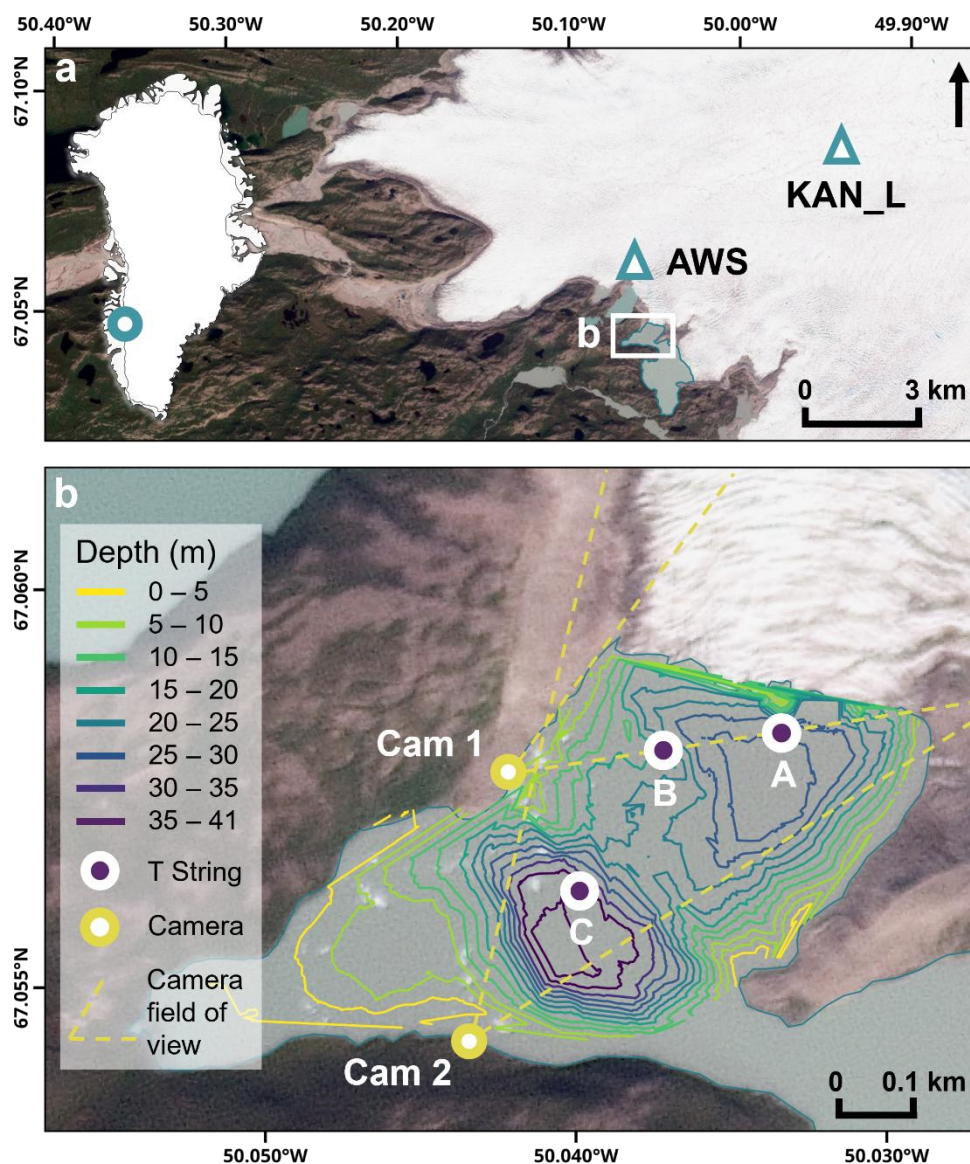


Figure 1: (a) Location of the study site in western Greenland. Triangles show the two Automatic Weather Stations (AWS and KAN_L) located outside the bounding box of panel b. (b) Bathymetry of the western basin of the study lake recorded in July 2024 shown with contour lines generated at 3 m intervals. Circles mark the locations of the three temperature strings (T String in legend) and two trail cameras. Background imagery is a PlanetScope scene from 4th August 2024 (Planet Team, 2026).

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2.2 Lake water temperature

Vertical lake temperature profiles were recorded at two mooring locations (String A and String B) over 72 days from the 11th of July to the 20th of September 2024, and at a third location (String C) over 11 days from the 1st to 12th of August 2025 (Fig. 1). Strings A and B were positioned relatively close to the calving front, while maintaining sufficient distance to minimise the risk of entanglement with calved ice, and at locations of differing lake depth to capture spatial variability in lake thermal regime. The following summer, String C was positioned farther from the calving front to mitigate iceberg interference during deployment and to characterise the thermal regime within a deeper section of the lake.

Each mooring comprised HOBO Tidbit MX2203 temperature sensors installed along a length of static cord, held by a buoy at the lake surface and anchored by a sandbag at the bed. In summer 2024, sensors were placed just below the surface and at depths of 2, 5, 10 and 25 m (String A) and 2, 5, 10 and 17 m (String B), reflecting variations in bathymetry at the two locations. On collection, String A was found to have moved ~100m to a position closer to String B, which remained stationary. In 2025, String C held a denser array of sensors placed near the surface and at subsequent depths of 3, 6, 9, 12, 15, 18, 21, 24, 27 and 29 m. This increased sensor density was adopted after assessing the 2024 observations to improve the detection of temperature gradients and thermocline movements over the shortened observational window. The bottom sensor on each string sat approximately 40 cm from the lakebed. Each sensor logged measurements at 10-minute intervals, with a manufacturer-stated accuracy of $\pm 0.2^{\circ}\text{C}$ and response time of seven minutes.

2.3 Meteorological data

Meteorological conditions during summer 2024 were recorded using a Campbell Scientific automatic weather station (AWS) installed at an accessible location on the ice sheet surface, 539 m a.s.l. and 295 m from the ice front (Fig. 1a). The AWS was fitted to record air temperature (manufacturer-stated accuracy $\pm 0.4^{\circ}\text{C}$), relative humidity ($\pm 2\%$), wind speed ($\pm 0.5 \text{ m/s}$) and direction ($\pm 5^{\circ}$), incoming shortwave radiation ($\pm 5\%$) and outgoing longwave radiation ($\pm 5\%$). Each variable was measured at one-minute intervals and recorded as the average (air temperature, wind speed and direction, solar radiation) or maximum value (relative humidity) every thirty minutes between the 1st of July and 20th of September 2024. While all instruments continued logging for the full observational period, uneven melt at the base of the AWS compromised its stability on the 15th of August, limiting usable data to the first 45 days of observations.

To extend the air temperature timeseries beyond this date, a multivariate linear regression model was applied using data from the PROMICE KANG_L AWS located 6.7 km northeast of the study lake (Fig. 1a; Fausto et al., 2021; How et al., 2022). Local air temperature was modelled as a function of PROMICE air temperature and additional meteorological variables measured at the study site. Specifically, predictor variables included PROMICE air temperature, relative humidity and wind direction, with the latter represented using sine and cosine components. Air temperature was estimated using Eq. (1):



$$T_{\{met\}} = \beta_0 + \beta_1 T_{\{prom\}} + \beta_2 RH + \beta_3 \cos(\theta) + \beta_4 \sin(\theta) + \varepsilon$$

where $T_{\{met\}}$ is the air temperature at the study site ($^{\circ}\text{C}$), $T_{\{prom\}}$ is the PROMICE air temperature ($^{\circ}\text{C}$), RH is relative humidity (%), θ is wind direction (radians), β_0 is the intercept, $\beta_1 - \beta_4$ are regression coefficients, and ε is the residual error.

135 Model performance was evaluated using a split-sample validation approach. The model was trained on data from 1st – 31st of July 2024 and validated against an independent period (1st – 14th of August 2024). During validation, the model achieved an R^2 of 0.70 and a root mean square error (RMSE) of 0.87°C , indicating good agreement between observed and predicted air temperatures. Following validation, the model was refitted using the full period of overlap (1st July – 14th August 2024), resulting in an R^2 of 0.79 and an RMSE of 0.86°C for the training period. This final model was then applied to reconstruct air
140 temperature for 15th August – 20th September 2024.

For summer 2025, since no AWS was assembled at the field site, all meteorological data were derived from the PROMICE KAN_L AWS.

2.4 Surface ablation model

Surface ablation between the 11th of July and 14th of August 2024 was estimated using the physically based surface energy
145 balance model presented by Brock and Arnold (2000). Melt energy (Q_m) at the location of the AWS was calculated as the sum of the principal surface energy fluxes (Eq. 2):

$$Q_m = SWR + LWR + SHF + LHF$$

where SWR and LWR are the net shortwave and longwave radiation fluxes, and SHF and LHF are the turbulent sensible and latent heat fluxes, respectively. Energy fluxes were calculated at hourly intervals and converted to melt rate, which is
150 expressed in millimetres water equivalent (mm w.e.) per hour. Fluxes directed towards the surface are defined as positive. Under the assumption that subsurface heat flux and the energy contribution from rainfall are negligible, melt is calculated as the residual of the surface energy balance, such that Q_m represents the energy available for melting. Melt was assumed to occur when Q_m was positive and the surface temperature reached the melting point.

Surface albedo was prescribed as a constant value of 0.42, derived from a map of broadband albedo at the study site for 2017
155 (Tedstone and Cook, 2020). Aerodynamic roughness length was set to 0.005 m, consistent with values reported for weathered bare ice in the southwest Greenland ablation zone (Smeets and van den Broeke, 2008).

A sensitivity analysis was conducted to test the influence of different parameters on the modelled melt output. Surface albedo and aerodynamic roughness length were varied independently across a range of plausible values, while all other parameters were held constant. Model sensitivity was evaluated using the range of total melt produced under each



160 parameterisation. Varying surface albedo between 0.35 and 0.50 resulted in total modelled melt ranging from 1041 to 646
mm w.e., equivalent to ± 12 mm w.e. day⁻¹ (47%). In contrast, varying roughness length between 0.0001 and 0.01 m
produced modelled melt differences of only ± 4 mm w.e. day⁻¹ (15%). Although this analysis indicates large potential
uncertainty associated with albedo, since the baseline albedo value of 0.42 is constrained by site-specific observations, this is
treated as an upper bound estimate. A more realistic uncertainty is therefore derived from the root-sum-square of roughness
165 uncertainty and the reported model error (± 1.2 mm w.e. day⁻¹; Brock and Arnold, 2000), giving an overall uncertainty of
approximately ± 4.2 mm w.e. day⁻¹. This corresponds to ~16% of the total modelled melt, indicating that while uncertainty is
moderate, the overall magnitude and temporal variability of melt can be considered robust.

2.5 Calving observations

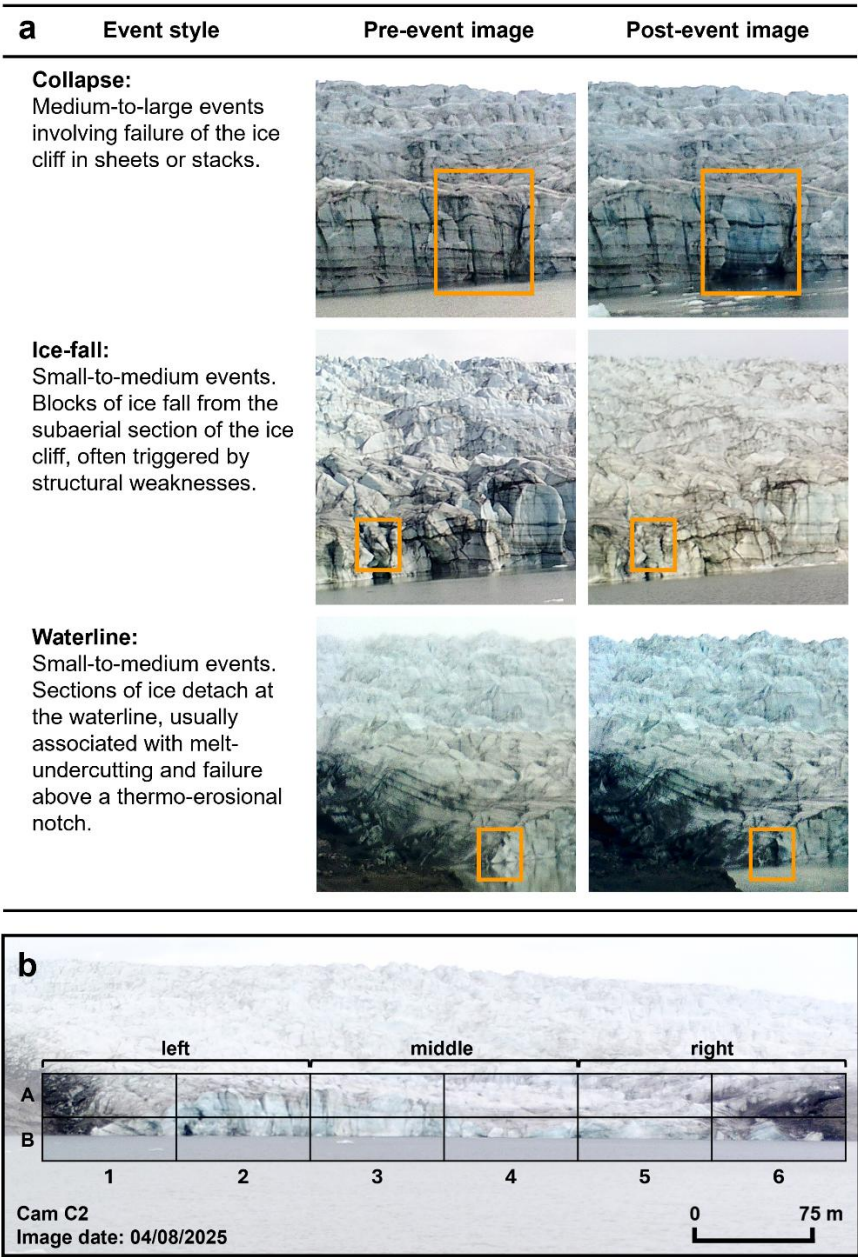
Time-lapse photography has been widely applied to monitor glacier dynamics, owing to its suitability for generating
170 temporally resolved datasets over periods much longer than typical field campaigns (e.g. Mallalieu et al., 2017; Minowa et
al., 2018; How et al., 2019). In this study, calving activity was recorded using two trail cameras (Ltl Acorn 5210A 12MP)
installed at fixed positions facing the ice front (Fig. 1). In 2024, images were taken three times daily at 9 am, 12 pm and 6
pm local time, providing consistent daylight coverage throughout the summer. In 2025, given the briefer deployment period
of 12 days, images were taken hourly during daylight hours between 5 am and 11 pm.

175 Individual calving events were identified and classified in terms of their location, style and size through manual inspection of
consecutive images, with events defined by observable changes to the ice front geometry between successive frames (How et
al., 2019; Kneib-Walter et al., 2021). To characterise the spatial distribution of calving, the ice front was subdivided into a
grid of 12 cells (A1-A6 and B1-B6) and each event assigned to one or more cells according to its spatial extent (Fig. 2b).
Event counts within each of the 12 individual cells were deemed too small for quantitative comparison, so for purposes of
180 analysis events were subsequently grouped into the left (including cells A1-B2), middle (A3-B4) and right-hand (A5-B6)
sections. Events spanning two sections were assigned to the section containing the largest proportion of the event. Since the
camera position and therefore the field of view varied slightly between 2024 and 2025, spatial classifications are not
comparable between years in an absolute sense. Differences in daily calving-event frequency between the three lake thermal
phases (described in section 3.1.1) were assessed using pairwise Mann–Whitney U tests, with statistical significance defined
185 as $p < 0.05$.

Calving styles were determined using an adapted version of the criteria presented by Kirkbride and Warren (1997) and
Mallalieu et al. (2020). Here, due to limitations in image resolution and the scale of observable features within the time-lapse
imagery, the classification scheme is simplified to include three broad categories of failure mechanism: collapses, icefalls,
and waterline events (Fig. 2). To distinguish between styles, events were assessed based on the point of failure initiation
190 relative to the waterline, the proportion of the ice-cliff height involved, and whether failure occurred as detachment of a



discrete block or more extensive face collapse. Calving event size was classified manually based on the proportion of the ice cliff height spanned by the event, with small events defined as <25% of the maximum ice cliff height, medium events as 25–75%, and large events as >75%.



195 **Figure 2:** (a) Criteria defining the three observed styles of calving based on their appearance in successive timelapse images. (b) Demonstration of grid cells applied to timelapse imagery of the ice front. Note that the given scale is approximate.



2.6 Statistical analysis: Generalised additive models

A series of generalised additive models (GAMs) were used to assess the influence of lake thermal conditions at String A, surface melt, and atmospheric forcing on daily calving rates during summer 2024. The modelling framework was designed to address three key aims: (i) to constrain the role of lake thermal conditions as a driver of calving, (ii) to quantify the contributions of other environmental factors, and (iii) to assess whether these contributions vary through the melt season.

GAMs extend traditional regression models by representing relationships between a response variable and multiple predictors as smooth, non-linear functions, rather than assuming strictly linear relationships (Wood, 2017). This approach allows both the magnitude and shape of individual predictor effects to be explored, making it well suited to assessing how various factors contribute to driving calving rates (Crippa et al., 2024). For predicting count data, the general form of the model is given by Eq. (3):

$$\log(E[Y_i]) = \beta_0 + \sum_{j=1}^p f_j(x_{ij})$$

where $(E[Y_i])$ is the expected value of the response variable (here, daily calving count) for observation i , β_0 is the model intercept, p is the number of predictors, and $f_j(x_{ij})$ represents a smooth function describing the potentially non-linear relationship between the predictor j and expected daily calving frequency. A log link function was used to account for the count-based response.

2.6.1 Data preparation

Calving observations were aggregated to daily total counts, while meteorological and limnological variables were averaged to daily resolution. Lake water temperatures were derived from String A due its proximity to the calving front. Candidate predictor variables included surface melt, air temperature, lake surface temperature, and lake thermal stratification (defined as the temperature difference between the surface and 25 m depth). To investigate the temporal response of calving to lake conditions, alternative representations of surface temperature and stratification were derived before model fitting. These comprised same-day absolute values, discrete 1- and 2-day lags, 48- and 72-h rolling means, and day-to-day change, calculated as the current daily value minus that of the preceding day.

2.6.2 Model specification

All models were fitted using a Poisson error distribution with a log link function. Predictor effects were modelled as smooth terms using penalised regression splines. Spline complexity was standardised across models by fixing the number of basis functions ($n_{splines} = 5$) for all smooth terms. This ensured consistent model flexibility and allowed comparison across



225 model configurations.

All models were fitted using the *pyGAM* package in Python, which performs automated smoothing parameter selection (Servén and Brummitt, 2018). Model performance was evaluated using pseudo- R^2 (explained deviance) and the Akaike Information Criterion (AIC).

2.6.3 Modelling framework

230 Alternative temporal formulations of lake surface temperature and thermal stratification were first evaluated to determine whether daily calving was more strongly associated with absolute lake conditions, short-term variability, or preceding conditions (Table S1). For each variable, single-predictor Poisson GAMs compared the same-day daily mean, day-to-day change, discrete 1- and 2-day lags, and 48- and 72-h rolling means. All screening models were fitted to a common set of 70 daily observations to ensure direct comparison of model performance.

235 A structured modelling approach was then used to isolate the effects of distinct predictors. Initial models focussed on lake thermal controls over the full July to September period. Predictor variables included lake surface temperature and stratification lagged by 48 hours, as well as air temperature and surface melt. Lake regime phase (as described in section 3.1.1.) was optionally included as a categorical term to account for intraseasonal shifts in lake thermal regime associated with observed overturning events, which in 2024 occurred on the 1st of August and 1st of September. Models sequentially
 240 examined surface temperature alone; surface temperature with regime phase; surface temperature with lagged stratification; and surface temperature, lagged stratification and regime phase. Additional models allowed the surface-temperature response, and subsequently both the surface-temperature and stratification responses, to vary among lake-regime Phases 1, 2 and 3 using phase-specific smooths.

Because the initial screening identified day-to-day change in surface temperature as the strongest standalone temperature
 245 formulation (Table S1), an equivalent sensitivity analysis repeated the same model hierarchy with day-to-day change in surface temperature substituted for absolute surface temperature. Two-day-lagged stratification was retained unchanged, allowing the influence of temperature representation to be evaluated while holding the remaining model structure constant. Lake thermal predictors were subsequently evaluated alongside other environmental drivers using models containing lake temperature, 2-day-lagged stratification, air temperature, and surface melt. These models were restricted to the common
 250 period for which all covariates were available, yielding 33 daily observations from the 13th of July to 14th of August 2024 after application of the 2-day stratification lag. Models were fitted first with common smooth responses across observations and then with lake-regime phase included as a categorical term. An additional exploratory model allowed each environmental-driver response to vary by lake-regime phase. The all-driver analysis was conducted for both the absolute-temperature and temperature-change formulations.



255 Nested additive GAMs fitted to identical observations were compared using analysis of variance (ANOVA) to test whether
 the reduction in deviance associated with addition of a predictor significantly improved model fit. Comparisons were
 restricted to genuinely nested model sequences. Phase-specific models containing separate smooths were therefore evaluated
 using AIC and explained deviance, rather than being included in the nested ANOVA sequence. To complement the GAM
 results, a one-way ANOVA test was used to assess whether mean daily calving-event frequency differed among lake regime
 260 phases 1, 2 and 3.

3 Results

3.1 Summer lake thermal structure

Lake water temperatures are reported for the two observational periods in 2024 and 2025. Lake temperatures from 2024
 extend across the full summer season (July – September), characterising temporal variability at String A and String B. The
 265 2025 lake water temperature measurements span 1st – 12th of August and are reported for a single string (String C; Fig. 1)
 with measurements recorded densely throughout the water column.

3.1.1 2024 lake thermal structure

Thermal conditions at String A and String B show similar patterns of evolution over the summer 2024 observation period
 (Fig. 3). In the early season, and until the 1st of August, the lake generally warms at all depths (hereafter Phase 1). The mid-
 270 season (August) is characterised by temperatures rising above 4°C, peaking, and then cooling gradually towards 4°C on the
 1st of September (Phase 2). Towards the end of the season (September), temperatures continue to drop below 4°C at all
 depths as winter approaches (Phase 3).

At both String A and String B, mean daily temperature peaked in the second week of August at all depths, reaching a
 maximum of 6.6°C at the surface (String A, 11th August) and 4.5°C at 25 m (String A, 8th August) (Fig. 3). Temperature
 275 minima at String A occurred in July at the surface (2.3°C, 11th July) and at 25 m depth (2.5°C, 10th July), and on the 21st of
 September at 2 m (1.91°C), 5 m (2.22°C) and 10 m (2.27°C). At String B, minima occurred in September at all depths but 17
 m, where the lowest daily mean temperature was in July (2.5°C).

At the onset of Phase 1, all five sensors on String A recorded temperatures between 2.5 and 2.7°C, indicating a well-mixed,
 isothermal water column in the early season (Fig. 3). Water temperatures gradually warmed at all depths throughout July,
 280 increasing from 2.4°C on the 10th to 3.4°C on the 31st at the surface of String A. Temperatures between the surface and 10 m
 increase with depth and feature a diurnal pattern. At 25 m, however, this signal is strongly attenuated, with lower diurnal
 variability and temperatures approximating the mid-range of values observed across the column (Fig. 4).



On the 1st of August, temperatures at String A merged across the water column before separating into a structure of positive stratification, i.e. cooling with depth, marking the beginning of Phase 2. The uppermost sensors (surface and 2 m) continued warming to peak values approaching 6 – 6.6°C (11th August), while deeper sensors (10 m and 25 m) warmed more moderately to 4 – 4.5°C. During this time, surface temperatures resumed a strong diurnal cycle which weakened with depth (Fig. 4). From the 12th of August, daily average temperature gradually decreased, until temperatures at all depths converged at 4°C on the 15th. Cooling across the water column continued throughout August. A mixed profile during this period was occasionally punctuated by episodes of stratification when surface temperatures exceeded 4°C. Lake water temperatures at all depths were significantly warmer ($p < 0.05$) during Phase 2 (August) than Phase 1 (July) and Phase 3 (September) (Fig. 5).

The third lake regime phase (Phase 3) began on the 1st of September, when overturning of the temperature profile produced a negative thermal gradient with a low range of temperatures across the column. During this phase, temperatures at all depths fell consistently below 4°C. The diurnal signal is seen only in occasional overnight cooling, restricted to the upper 2 m (Fig. 4). Temperatures at 10 m and above diverge from those at 25 m on the 13th of September, marking the separation of cool surface waters (epilimnion) from warmer bottom waters (hypolimnion) (Fig. 3).

Temperature change at String B was generally consistent with String A (Fig. 3). At the start of observations, all five sensors clustered between 2.5 – 3.0°C. Temperatures across all depths increased progressively, reaching a broad peak around early-to-mid August, during which near-surface sensors (surface and 2 m) exhibited maximum values of ~6 – 6.4°C, while deeper sensors (10 m and 17 m) warmed to ~4 – 4.5°C. Diurnal variability was most pronounced at the surface and 2 m sensors throughout the peak warming phase (Fig. 4). Following the mid-August maximum, temperatures declined across all depths into September, with all depths reconverging toward 2.5 – 3.0°C by mid-September. High-frequency variability at the near-surface sensors persisted throughout the cooling phase. Compared to String A, String B exhibits a relatively lower amplitude of short-term deviations from the mean. Stratification across the water column is generally weaker at String B, and July overturning occurs three days later than at String A.

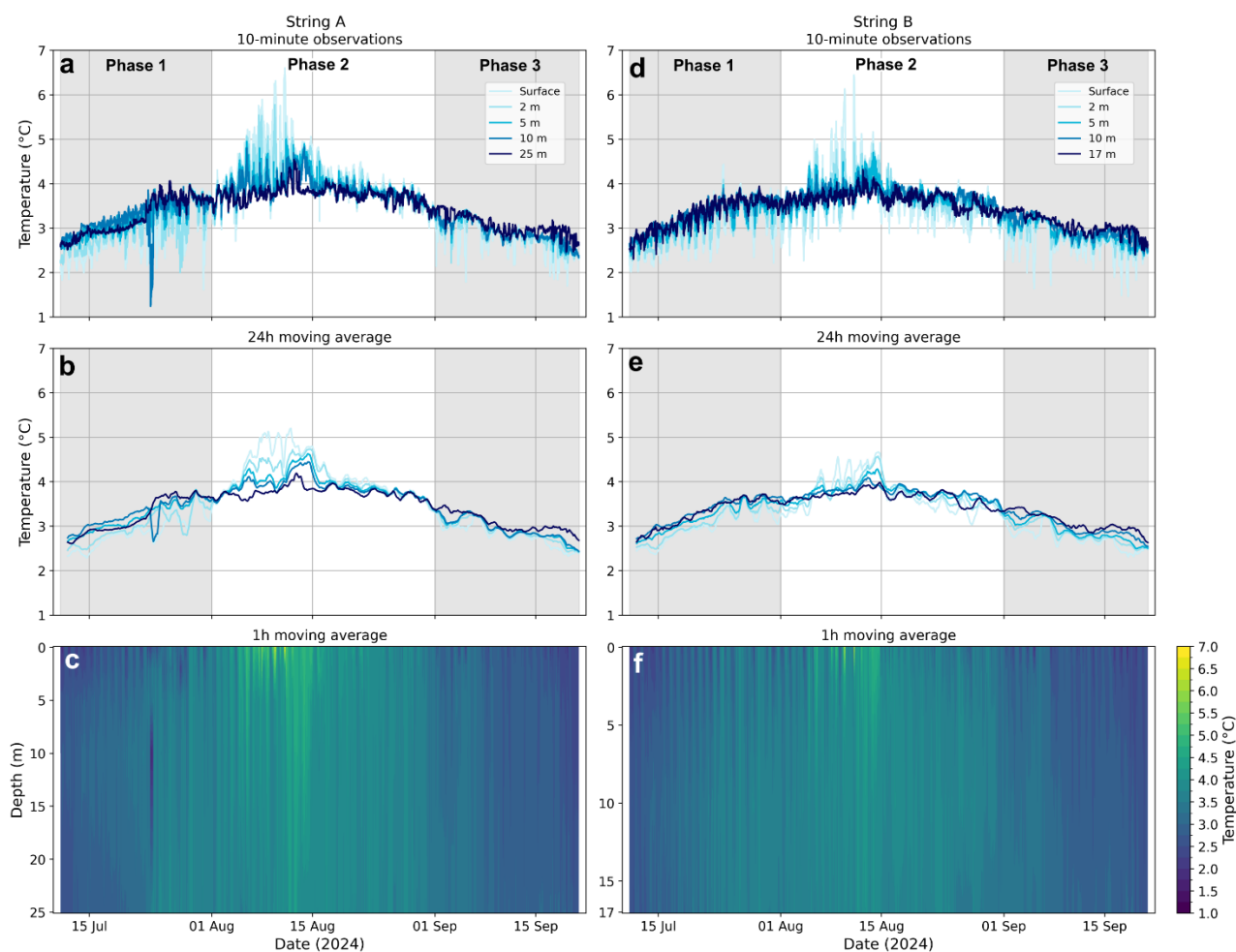


Figure 3: Evolution of lake water temperatures at String A and String B during summer 2024. Observations are displayed in raw form (10-minute resolution; panels a, d), as a 24-hour moving average (b, e) and as a 1 hr moving average (c, f), linearly interpolated between depths. Shading marks the three phases of lake thermal regime.

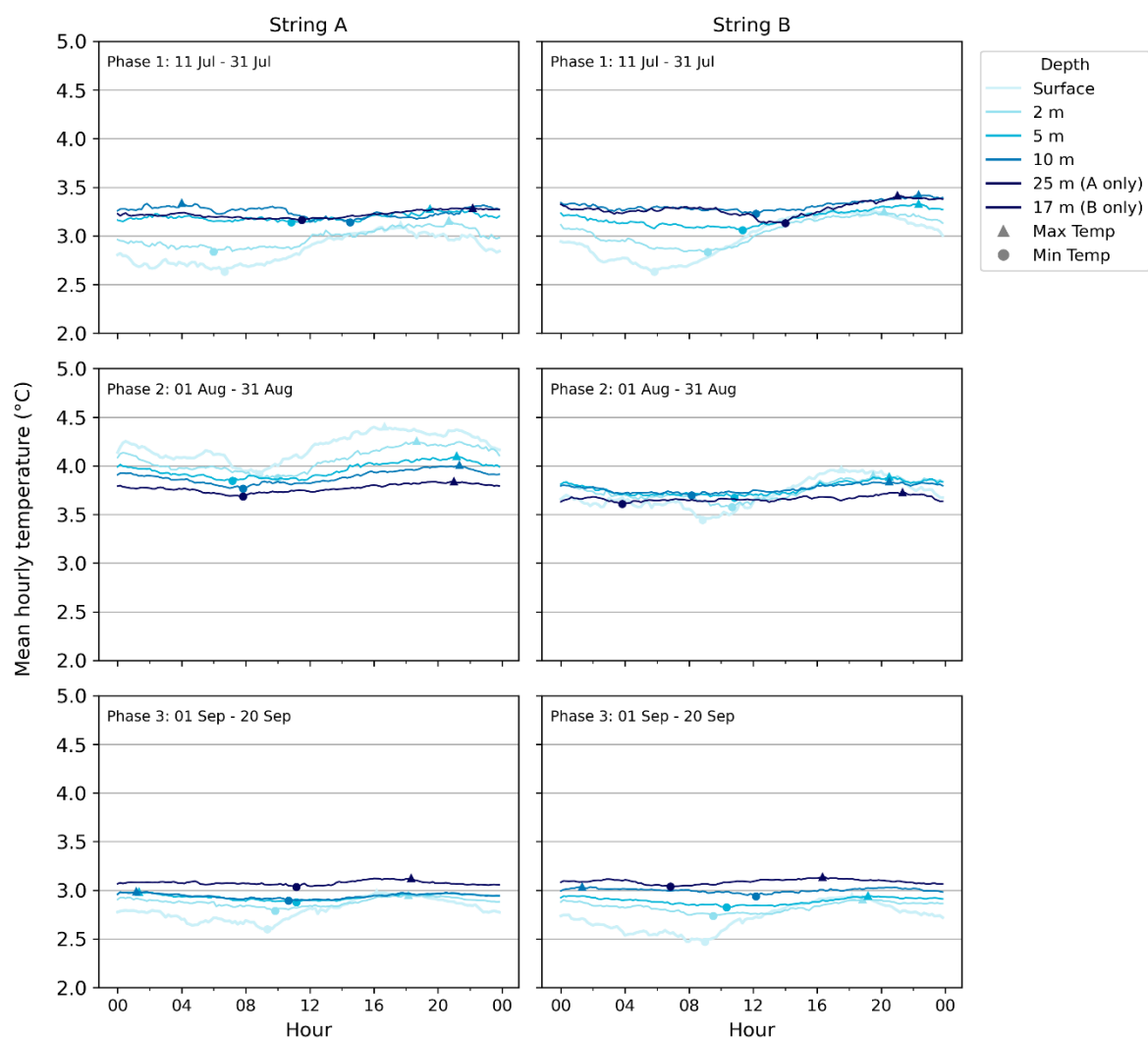
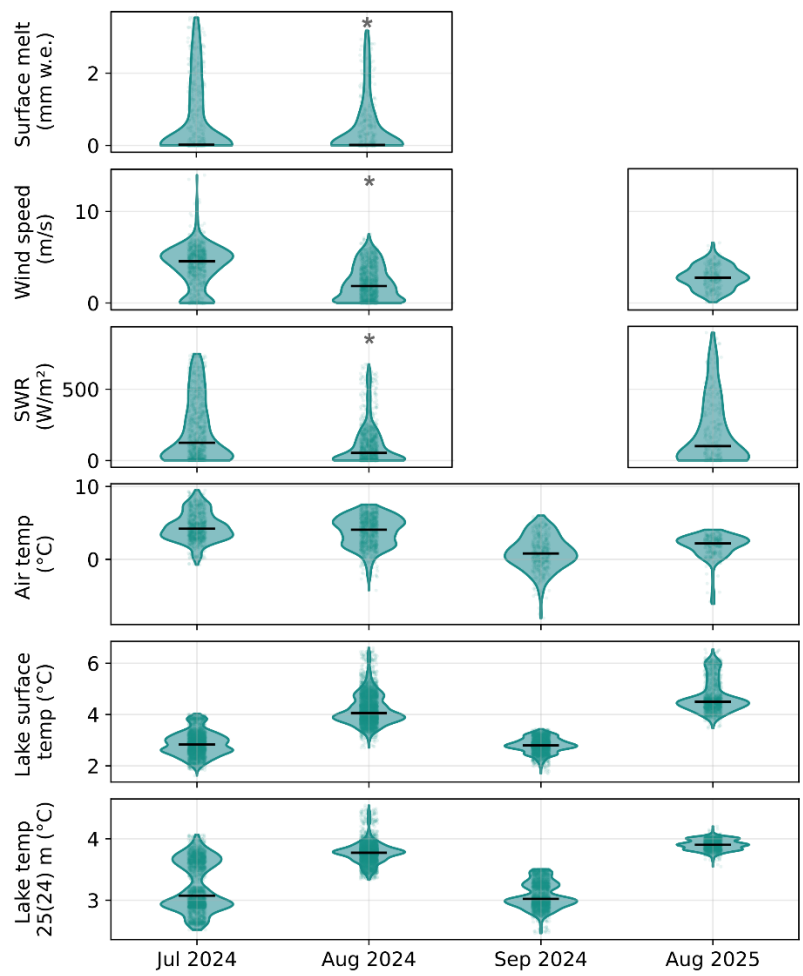


Figure 4: Average diurnal lake thermal regime at String A and String B, calculated as the mean temperature at each depth throughout the day within three periods across summer 2024.



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Figure 5: Lake water temperature, surface melt and meteorological conditions (air temperature, wind speed and shortwave radiation (SWR)) during each lake regime phase in summer 2024 and August 2025. Lake temp at depth is represented by measurements from 25 m deep during 2024 and 24 m deep during 2025. Lines indicate median values and shaded widths reflects data distribution. Asterisks mark observations which span a limited period of August 2024 (1st – 13th August).

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3.1.2 2025 lake thermal structure

Observations from the 1st to 12th of August 2025 show a persistent vertical temperature gradient, with temperatures decreasing with depth through the upper 18 m of the water column. Below this, temperatures generally warmed towards the lake bed (Fig. 6). Mean temperatures across the record ranged from 3.8°C (29 m) to 4.7°C (surface), comparable to mean observed over the same period in 2025 (3.9°C at 25 m, 4.5°C at the surface) (Fig. 5). At depths 12 m and below,

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temperatures remained close to 4°C (12 m: 3.4 – 4.8°C) and exhibited limited temporal variability; mean daily temperatures at these depths varied only between 3.5°C (12 m) and 4.0°C (12 m).

In contrast, the upper water column displayed pronounced high-frequency variability during the first week, when temperatures ranged from 3.4°C (9 m) to 6.5°C (surface). The amplitude of variability decreased rapidly below 9 m, and diurnal fluctuations became progressively dampened at depth (Fig. 7).

From the 7th of August onwards, near-surface variability diminished, with shallow-water temperatures converging towards 4 – 4.5°C, and the mean diurnal range across all depths decreasing from 0.7°C (1st – 6th Aug) to 0.4°C (7th – 12th Aug) (Fig. 7).

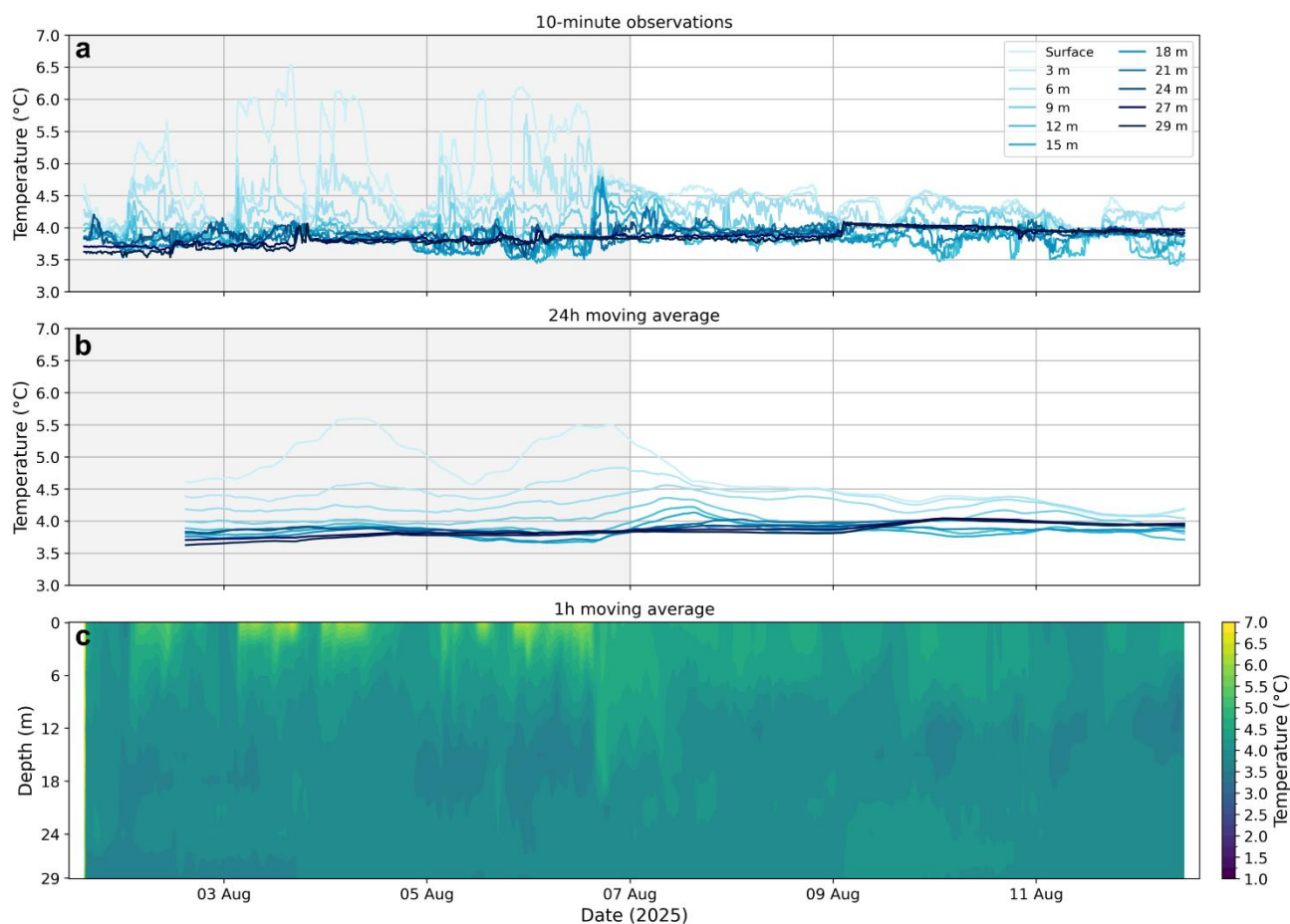


Figure 6: Evolution of lake water temperatures, from surface to 29 m depth during summer 2025. Observations are displayed in raw form (10-minute resolution; panel a), as a 24-hour moving average (b) and 1-hour moving average (c). Gray shading in (a) reflects distinct regime phases.

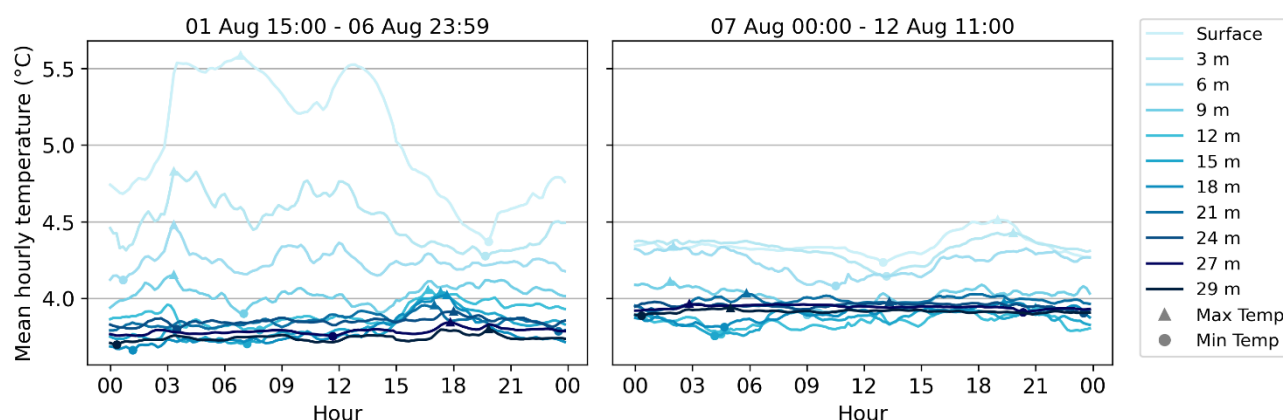


Figure 7: Average diurnal lake thermal regime during 12 days of August 2025, calculated as the mean hourly temperature at each depth within two phases.

3.2 Meteorological conditions

3.2.1 2024 meteorological conditions

The combined AWS and modelled air temperature data show variations between a maximum of 9.5°C on 18th July and a minimum of -8.1°C on the 9th of September, oscillating throughout the summer (Fig. 8). Mean daily temperature increased from 3.2°C at the start of the record (11th of July) to a peak of 8.0°C on the 18th of July, before cooling through late July into early August. In August, mean daily air temperature ranged from a maximum of 6.9°C on the 12th to a minimum of -2.9°C on the 31st, with an overall cooling trend from mid-August onwards, reaching -0.2°C by the 19th of September.

Incoming shortwave radiation broadly corresponded with air temperature, exhibiting maximum and minimum daily means of 298.7 W m² on the 12th of July and 1.7 W m² on the 13th of August. Wind speed ranged between 0 m s⁻¹ and 14.0 m s⁻¹, periodically aligning with air temperature.

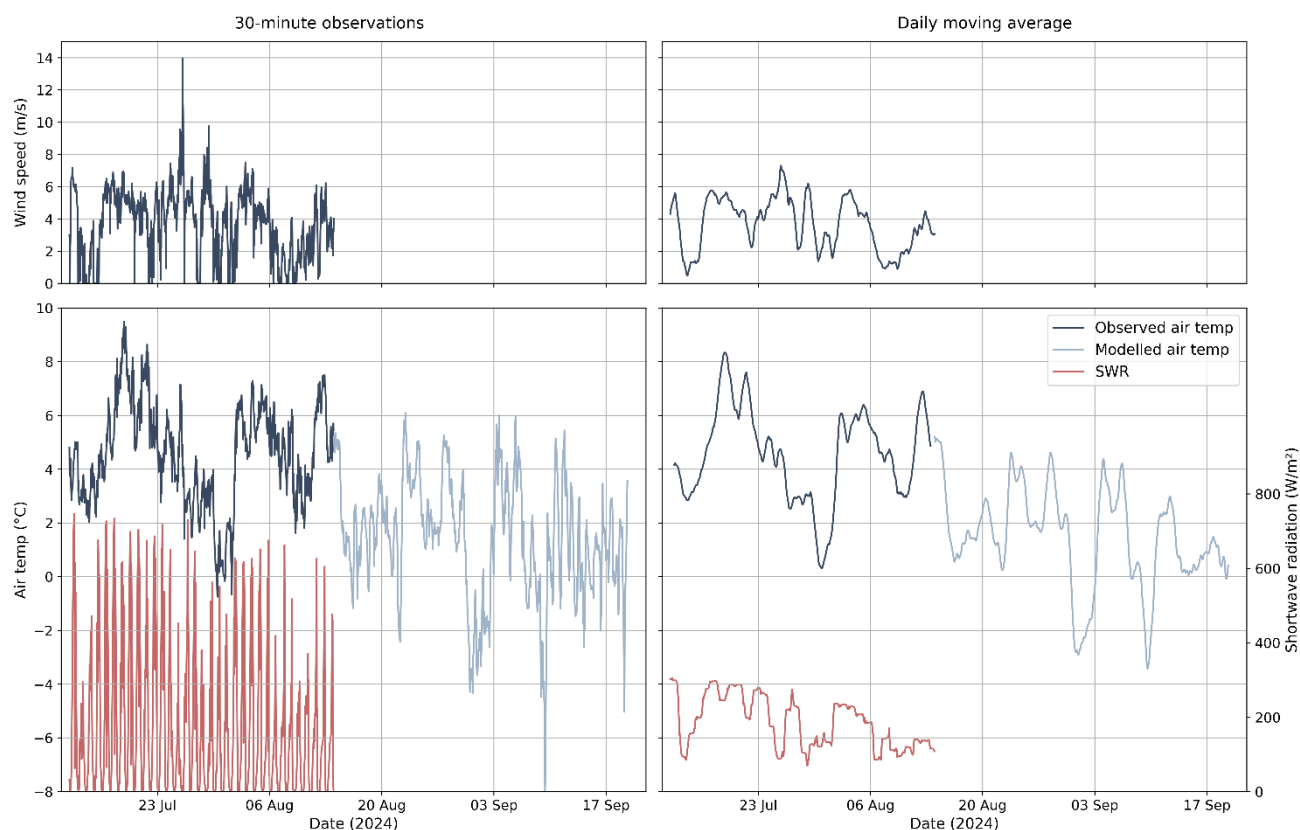


Figure 8: 30-minute observations and daily moving average of wind speed, incoming short-wave radiation (SWR), and observed and hourly modelled air temperature during summer 2024.

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3.2.2 2025 meteorological conditions

From the 1st to 12th of August 2025, the PROMICE KAN_L meteorological data show minimum and maximum air temperatures of -6.1°C and 4.1°C, respectively (Fig. 9). Mean daily air temperature generally ranged between 1.9°C and 3.1°C but dropped below 0°C on the 7th (-0.9°C) and again on the 12th (-0.5°C). This cooling is consistent with a decrease in daily mean incoming shortwave radiation on the 6th of August, preceding the air temperature minimum. Wind speed varied throughout the observation period, ranging from 0 to 6.7 m s⁻¹, with peak wind speeds coinciding with the air temperature minimum on the 7th of August.

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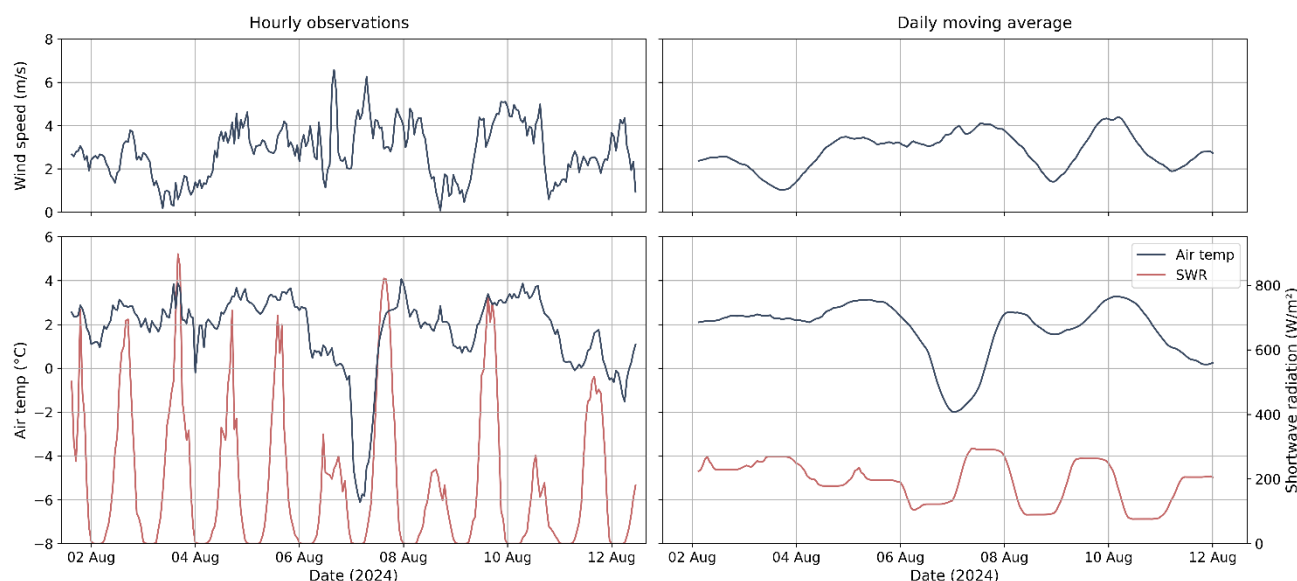


Figure 9: Hourly PROMICE KAN_L observations and daily moving average of wind speed, incoming short-wave radiation and air temperature during August 2025.

3.3 Modelled surface ablation

Surface melt modelled over 1st of July – 14th of August exhibits a strong diurnal signal with a mean rate of 9.0 mm w.e. day⁻¹, totalling 306.26 mm w.e. (Fig. 10). Increases in daily melt from the 12th – 22nd of July and 1st – 2nd of August correspond with warming air temperatures, while reduced melt 30th of July – 1st of August aligns with lower air temperatures over the same period. Maximum daily melt of 29.9 mm w.e. occurs on 22nd of July, as does the maximum hourly melt rate of 7.96 mm w.e.

The relative contributions of individual melt components are also shown in Fig. 10. Shortwave radiation (SWR) provides the dominant positive contribution to melt throughout the record and exhibits the strongest diurnal variability. Longwave radiation (LWR) and sensible heat flux (SHF) contribute smaller positive fluxes, whereas latent heat flux (LHF) is predominantly negative and offsets melt across the entire period.

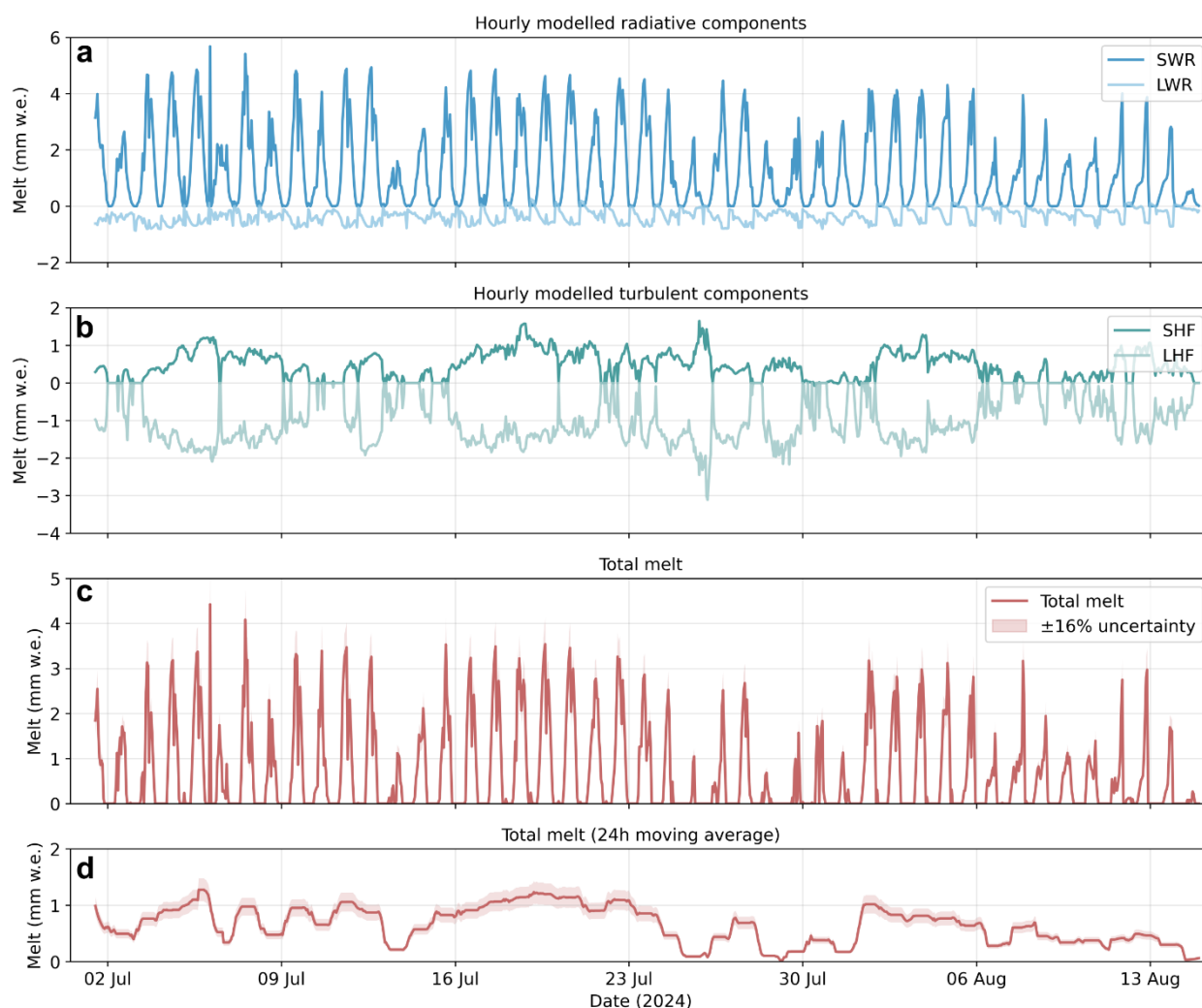


Figure 10: Hourly modelled melt contributed by individual (a) radiative and (b) turbulent modelled melt components, and (c) hourly total and (d) 24-hour moving average total modelled melt, 1st of July – 14th of August 2024.

3.4 Calving observations

3.4.1 2024 calving observations

A total of 137 calving events were identified in timelapse imagery taken over 72 days in summer 2024 (12th of July – 20th of Sept). These events are evenly distributed across the calving front. Large events ($n = 48$) are clustered in the middle of the calving front, whilst small ($n = 49$) and medium ($n = 40$) events occur predominantly on the left- and right-hand sides, where the calving face is shorter and has a shallower gradient (Fig. 11).



Calving activity varied over the observation period, ranging between zero and six events per day. When analysed within the three phases of distinct lake thermal regime, mean daily event frequency was lowest in Phase 1 (11th – 31st of July; 1.67 events per day), peaked in Phase 2 (2.13 events per day), and declined again in Phase 3 (1st – 20th of Sept; 1.80 events per day). However, despite differences of ~20% in mean daily calving rate among lake regime phases, these differences were not statistically significant (one-way ANOVA, $F(2,69) = 0.56$, $p = 0.57$), and standard deviations within each block were comparable (1.41 to 1.62 events per day), reflecting consistently high day-to-day variability throughout the summer.

The composition of calving events varied across phases. Small events dominated in July (0.76 per day) and early August (1st – 14th of Aug; 0.79 per day), while the number of medium events increased in late August (15th – 31st of Aug; 1.1 per day), when they exceeded both small (0.60 per day) and large (0.93 per day) events. Large events were most frequent in August (0.80 per day), coinciding with peak lake temperatures at all depths and reflecting an increase in the number of collapse events. By September, small, medium, and large events occurred at equal rates (0.60 per day).

3.4.2 2025 calving observations

Between 3 pm on the 1st of August and 11 am on the 12th of August 2025, a total of 49 calving events were identified (Fig. 11). Events were clustered to the left ($n = 23$) and middle ($n = 19$) of the calving front. Small, medium and large events were evenly represented in the record, though their spatial distribution differed: large events were concentrated on the left of the calving front ($n = 10$), medium events were most frequent in the middle ($n = 9$), and small events were distributed across both the left and middle sections ($n = 10$ and $n = 4$ respectively).

Calving activity ranged from one to 11 events per day, peaking on the 4th of August. When separated into two periods corresponding to phases of distinct lake thermal regimes, mean daily event frequency was higher in Phase 1 (4.83 events per day; 1st – 6th of August) than in Phase 2 (3.5 events per day; 7th – 12th of August), though this difference was not statistically significant ($p = 0.627$). Variability was high in both periods ($\sigma = 3.54$ and 2.25 events per day respectively). The size composition of events was broadly similar across periods, with small, medium, and large events each contributing comparably to total activity. Large events occurred at a slightly lower daily rate in Phase 1 (1.5 events per day) than Phase 2 (1.8 events per day), whilst small and medium event rates both declined from Phase 1 (1.7 and 1.7 per day) to Phase 2 (1.21 and 1.21 per day). Collapses were predominant across the 12 days of observation.

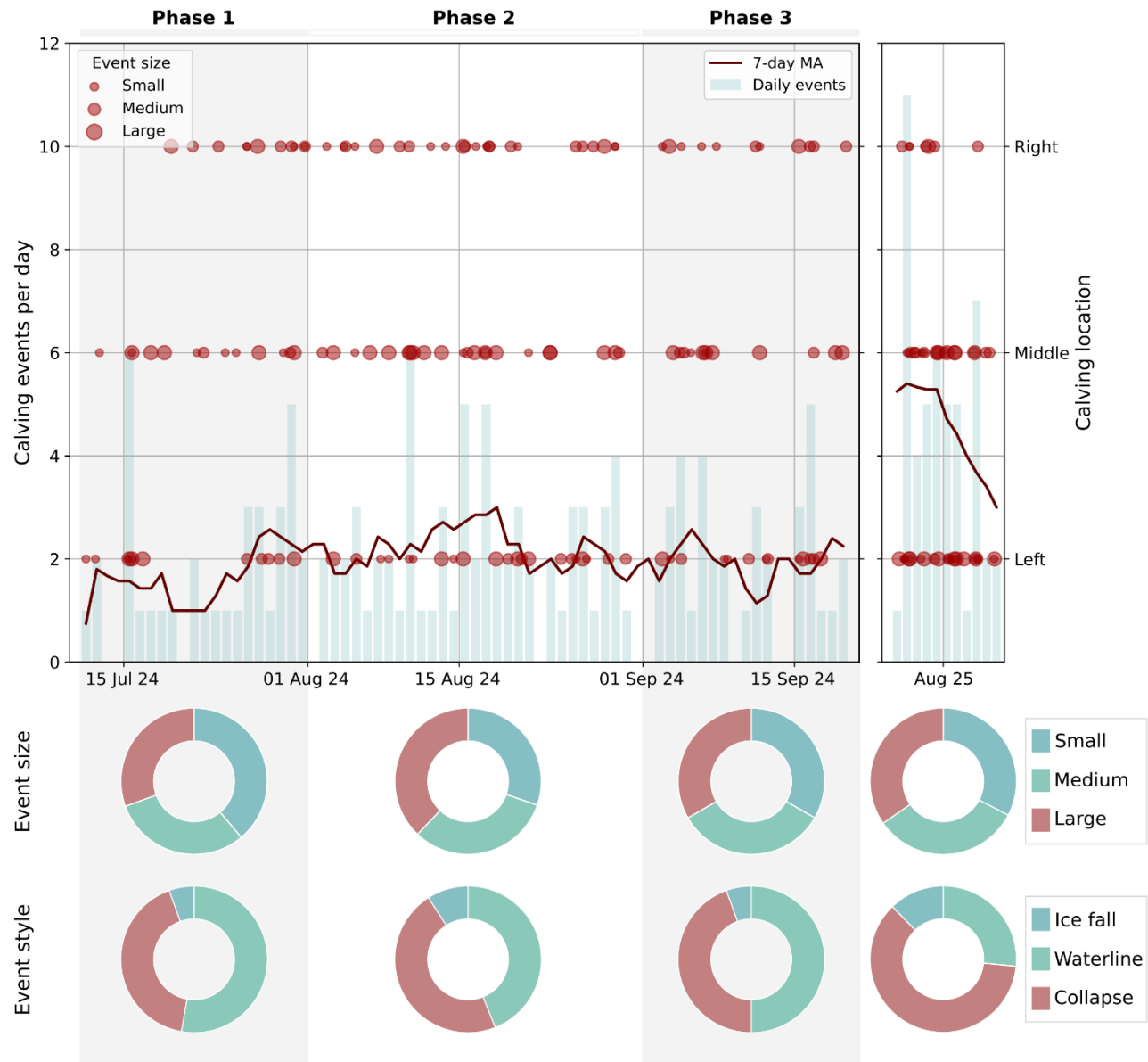


Figure 11: Calving events during summer 2024 and August 2025. Scattered points represent individual events, where point size reflects event size. Bars show number of events per day, overlaid by a 7-day centred moving average. Shaded columns represent the three lake thermal regime phases in summer 2024. Donut plots show the proportion of calving events of each size and style during the three phases of lake thermal regime and August 2025.



3.5 Statistical modelling

A series of GAMs were fitted to assess the influence of lake thermal conditions on daily calving rates during summer 2024 (Table 1). Lake conditions were represented using lake surface temperature and 48-hour lagged stratification at String A. Screening of alternative temporal formulations (Table S1) showed that day-to-day change in surface temperature explained more variation in calving when considered alone than absolute surface temperature (pseudo- $R^2 = 0.095$ and 0.064 , respectively). However, this advantage was not maintained when stratification and other environmental predictors were included. Absolute surface temperature was therefore retained in the primary models, while results from equivalent models using day-to-day change in surface temperature are reported in Table S2.

The model combining lake surface temperature and 48-h lagged stratification explained 10.8% of variability in daily calving frequency (model A3; pseudo- $R^2 = 0.108$; AIC = 242.124). Including lake thermal regime phase as a fixed effect increased explained deviance slightly (model A4; pseudo- $R^2 = 0.125$) and strengthened the apparent statistical association between lake surface temperature and calving ($p = 0.004$), although the phase term itself was not significant ($p = 0.546$). Nested-model comparisons indicated that neither the addition of lagged stratification to the temperature-only model nor the subsequent addition of lake-regime phase significantly improved model fit (ANOVA, $p = 0.173$ and 0.460 , respectively, Table S3).

The partial effect curves for model A4 (Fig. 12) indicate a non-linear response to lake surface temperature, with calving rates increasing up to $\sim 4^\circ\text{C}$ before weakening at higher temperatures, while stratification exhibits a weak positive effect with substantial uncertainty. The regime term captures residual temporal variability not explained by lake temperature and stratification, with lower residual variability in Phase 2 (August) compared to phases 1 (July) and 3 (September), suggesting that predictor-calving relationships may vary between lake thermal regime phases.

Allowing lake condition relationships to vary among thermal regime phases increased explained deviance to 15.2% (model A6), the highest value among the lake-only models, suggesting some sub-seasonal variation in the fitted relationships. However, this increase in explanatory power was accompanied by a slightly higher AIC (248.383), and none of the phase-specific lake effects was statistically significant. Within A6, the strongest temperature-calving association occurred during Phase 2 ($p = 0.077$).

More complex models including air temperature and surface melt rates as additional predictors and applied to a shorter period (11th July – 14th August) increase overall explanatory power (model A7; $R^2 = 0.322$), but none of the predictors are statistically significant.

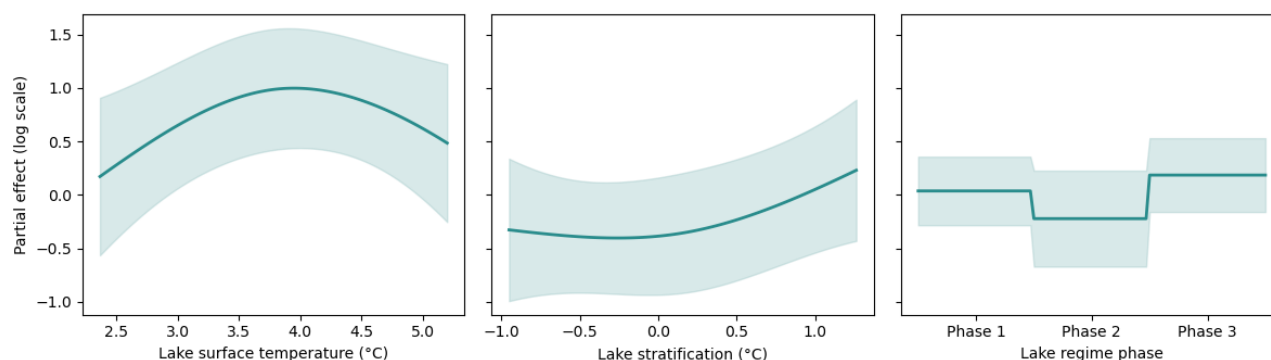


Figure 12: Partial-effect plots from model A4 showing the relationships between lake surface temperature, 48-h lagged lake stratification, lake thermal-regime phase and daily calving frequency. Solid lines represent the estimated strength of smooth and fixed effects and shaded areas indicate 95% confidence intervals. Partial effects are shown on the model's log scale, with higher values indicating a greater positive contribution to expected daily calving frequency after accounting for the other terms in the model.



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Table 1: Summary of GAM performance and predictor significance for calving rates during summer 2024 and each of the three lake regime phases. Model fit is reported as pseudo R^2 and AIC, while p values reflect the significance of each predictor variable.

Model	Period	Pseudo- R^2	AIC	Predictor	p
A1	11/07/24 – 20/09/24	0.064	242.748	Daily average lake surface temperature	0.000013
A2	11/07/24 – 20/09/24	0.089	244.285	Daily average lake surface temperature	0.000015
				Lake thermal regime phase	0.430
A3	11/07/24 – 20/09/24	0.108	242.124	Daily average lake surface temperature	0.021
				Daily average lake stratification (48h lag)	0.209
A4	11/07/24 – 20/09/24	0.125	244.186	Daily average lake surface temperature	0.004
				Daily average lake stratification (48h lag)	0.389
				Lake thermal regime phase	0.546
A5	11/07/24 – 20/09/24	0.088	247.545	Daily average lake surface temperature (Phase 1)	0.170
				Daily average lake surface temperature (Phase 2)	0.096
				Daily average lake surface temperature (Phase 3)	0.037
				Lake thermal regime phase	0.038
A6	11/07/24 – 20/09/24	0.152	248.383	Daily average lake surface temperature (Phase 1)	0.429
				Daily average lake surface temperature (Phase 2)	0.077
				Daily average lake surface temperature (Phase 3)	0.271
				Daily average lake stratification, 48h lag (Phase 1)	0.154
				Daily average lake stratification, 48h lag (Phase 2)	0.290
				Daily average lake stratification, 48h lag (Phase 3)	0.592
				Lake thermal regime phase	0.504
A7	11/07/24 – 14/08/24	0.322	112.878	Daily average lake surface temperature	0.224
				Daily average lake stratification (48h lag)	0.753
				Daily average air temperature	0.469
				Daily average surface melt rate	0.201
A8	11/07/24 – 14/08/24	0.340	113.746	Daily average lake surface temperature	0.074
				Daily average lake stratification (48h lag)	0.820
				Daily average air temperature	0.597
				Daily average surface melt rate	0.124
				Lake thermal regime phase	0.439



485 4 Discussion

4.1 Lake thermal regime

Our observations from an ice-marginal lake in west Greenland suggest a summertime thermal regime which bears some similarities to that of ice-contact lakes elsewhere (Sugiyama et al., 2021; Bird et al., 2022; Scofield et al., 2025). Typical of dimictic lakes, the summertime seasonal evolution of lake temperatures in 2024 is defined by two overturning events (Fig. 3), which occur when water at all depths crosses a thermal threshold of 4°C and is therefore at its densest (Sugiyama et al., 2016). These events mark the onset and ending of a period of relatively warm water temperatures with a positive thermal gradient (hereafter, ‘warm period’). Specifically, overturning on 1st August 2024 initiates a reorganisation of the thermal structure, where warm surface waters (epilimnion) are separated from cool waters at depth (hypolimnion) via the development of a thermocline. This separation allows the epilimnion to heat or cool more readily in response to changes in air temperature, solar radiation and wind speed, fluctuating over diurnal and intraseasonal timescales (Woolway et al., 2014). Although temperatures at depth are more stable than at the surface, the average diurnal range ($\pm 0.4^{\circ}\text{C}$ at 25 m, Fig. 4) is generally greater than has been observed at ice-marginal lakes elsewhere (Scofield et al., 2025). At Thulagi Lake in the Manaslu region of the Himalaya, for instance, the summertime diurnal range is limited to $\pm 0.2^{\circ}\text{C}$ at 30 m depth (Scofield et al., 2025). Further, whilst temperature variability at depth in Thulagi Lake is distinctly decoupled from changes at the surface, we observe a similar, albeit attenuated, pattern of variability at depth compared to the surface, suggesting that the thermocline observed here is perhaps weaker than at other ice-contact lakes.

In late August 2024, cooling of surface waters to $\sim 4^{\circ}\text{C}$ indicates the collapse of the thermocline, marking the end of the warm phase. If thermal stratification is defined by a difference of at least 1°C between the surface and lake bed (Woolway et al., 2014), we observe stratification during only 12 days of the warm period at both String A and String B in 2024 (Fig. 3). This is considerably briefer than the warm periods of ice-contact lakes in different climatic regions and topographic settings; in the Himalaya and Patagonia, for example, lakes exhibit positive thermal gradients for two to three months a year, with stable stratification persisting for weeks or months (Sugiyama et al., 2021; Scofield et al., 2025; He et al., 2026). We suggest that the relative brevity of the stratification observed here reflects both regional climate and comparatively shallow lake depths (max 29 m). Ice-contact lakes occupying steep-sided mountain valleys are considerably deeper than our study lake, with maximum depths of 580 m, 245 m, and 78 m recorded at Brazo Upsala in Patagonia, in southeastern Tibet, and at Thulagi Lake in the Himalaya, respectively (Sugiyama et al., 2016; Scofield et al., 2025; He et al., 2026). Shallow depths surveyed here (Fig. 1) may force a weaker thermocline, which would be more easily eroded by changes in thermal input and mixing on particularly cold or windy days. Indeed, we observe earlier overturning and greater structural instability at String B (Fig. 3), located in a shallower area (17 m) than String A, suggesting that thermal regime may be modulated by bathymetric variability even within a single lake.

Our in situ data can be contextualised against satellite-derived lake surface water temperatures reported by Carrea et al.



(2025) for years 1995 - 2022, who found that ice-marginal lakes in west Greenland typically exceeded 4°C in early- to mid-July, ~three weeks early than in our observations (Fig. 3). Carrea et al. (2025) also report lake surface temperatures generally falling below 4°C in late September, approximately three weeks after breakup of the positive thermal gradient at our study lake. Despite these differences, Carrea et al. (2025) highlight variability in the duration of the warm period both among individual lakes and over time, suggesting that the shorter warm period observed here is not necessarily anomalous, but rather reflects inherent variability in lake thermal regimes. Collectively, these findings highlight the considerable heterogeneity in sub-seasonal thermal evolution among ice-marginal lakes in west Greenland and emphasise the importance of local controls, including, for example, lake bathymetry and meltwater inputs, in modulating their response to atmospheric forcing (Boehrer and Schultze, 2008).

4.2 Calving regime

At the nearby Russell Glacier, Mallalieu et al. (2020) observed a summertime calving regime driven by melt-undercutting, which dominated ice front retreat over interannual timescales. Directly comparing our observations to this record, and other records, of calving activity is difficult due to the lack of one consistent method of quantifying and classifying calving events. However, the style of calving observed at our study lake is broadly similar to that at Russell Glacier, whereby events of all styles recorded here predominantly (92%) feature ice loss at the waterline.

We record a sub-seasonal shift in calving regime between July and September 2024, characterised by an increased proportion of medium and large calving events during Phases 2 and 3 (Fig. 11). Waterline and collapse events remained the dominant calving style throughout the summer, indicating undercutting of the ice cliff as a primary mechanism of terminus instability. This interpretation is supported by field observations of a pronounced waterline notch at the ice front. Combined with the crevassed ice surface, which reflects steep bed topography immediately upstream of the terminus, this suggests that progressive undercutting promoted the outward toppling of overhanging ice blocks. This is consistent with the calving mechanism originally described by Kirkbride and Warren (1997) at a temperate lake-terminating glacier in New Zealand and subsequently observed across a range of climatic and topographic settings (Warren and Kirkbride, 2003), including Patagonia (Sugiyama et al., 2021), west Greenland (Mallalieu et al., 2020), and Arctic Sweden (Dye et al., 2025).

Several of these glaciers also exhibit buoyancy-driven calving, typically expressed as the rapid emergence and rotation of large icebergs at the glacier front (Warren and Kirkbride, 1997; Warren et al., 2001; Benn et al., 2007; Minowa et al., 2018; Sugiyama et al., 2019). While we did not identify any such events from the timelapse imagery taken during the 2024 and 2025 field seasons, the absence of direct observations of the subaqueous ice-cliff geometry means that we cannot rule out buoyant calving as a contributor to ice loss at our study site.

To assess whether buoyancy driven calving is required to explain subaqueous terminus retreat, we consider the annual retreat rate of the subaerial cliff, which averaged 26 m yr⁻¹ between 2016 and 2025. If most subaqueous retreat occurs when the lake



is largely ice-free, approximately 90 days in 2024, this corresponds to an average retreat rate of $\sim 0.3 \text{ m d}^{-1}$ (Mallalieu et al., 2020; Carrea et al., 2025). This value lies at the lower end of subaqueous melt rates at lake-terminating glaciers exposed to water temperatures similar to, or slightly warmer than, those measured here (Haresign and Warren, 2005; Diolaiuti et al., 2006; Sugiyama et al., 2016; Minowa et al., 2017). This rate suggests that sustained subaqueous melting during the summer season could plausibly account for observed retreat without invoking buoyant calving as an additional mechanism, provided that an extensive submerged ice terrace is absent.

4.3 Relationship between lake temperature and calving

Our observations suggest that sub-seasonal changes in lake thermal regime modulate calving behaviour at a lake-terminating margin of the GrIS. During Phase 1 (early summer), prior to lake overturn, the absence of thermal stratification implies that energy available for melt is distributed relatively uniformly across the subaqueous ice front. Calving activity was lower than in other phases despite the presence of a heavily crevassed and structurally weakened terminus, suggesting that the ice cliff was preconditioned for failure but not undergoing rapid destabilisation. At the onset of Phase 2, the ‘warm period’, lake overturn establishes a positive thermal gradient, with water temperatures peaking throughout the column and exhibiting enhanced variability near the surface. Calving frequency is 28% higher than in Phase 1, with a lower proportion of small events in favour of large collapses. This is also the period during which lake temperature exhibited the strongest association with calving activity in the most comprehensive phase-varying model (Model A6; Table 1). As lake temperatures subsequently declined and the water column transitioned towards an inverse thermal structure during Phase 3, calving activity decreased. Together, these observations suggest that changes in both lake thermal structure and lake temperature may influence calving regime (Fig. 13).

This interpretation is consistent with previous observations from lake-terminating glaciers, where seasonal warming of proglacial lakes has been related to enhanced frontal ablation through increased subaqueous melt (Minowa et al., 2017; Sugiyama et al., 2021; Dye et al., 2025). Röhl (2006) demonstrated that thermal stratification plays an important role in this relationship by promoting the concentration of thermal energy within a narrow elevational range near the waterline, facilitating thermoerosional rates sufficient for notch development. Interestingly, our models suggest that the influence of lake temperature on calving is greatest at $\sim 4^{\circ}\text{C}$ (Fig. 12), coinciding with the overturning events that bound Phase 2. This finding may reflect the particularly dynamic nature of the thermal regime during these transitional periods, when the water column is actively reorganising and heat is being redistributed vertically, potentially driving short-lived but relatively rapid periods of undercutting before the column stabilises into a more stratified or fully mixed configuration (Röhl, 2006). Overall, our observations suggest that, at this grounded and relatively slow-flowing lake-contact margin of the GrIS, sub-seasonal variability in lake thermal conditions may modulate calving rates by controlling the development of thermal notches and the subsequent failure of overhanging ice.



Our study lake occupies a relatively passive sector of the ice margin, where mean annual ice flow velocities at the terminus
 580 are slow ($\sim 30 \text{ m yr}^{-1}$; Gardner et al., 2019). At faster-flowing glaciers, dynamic calving driven by high longitudinal strain
 rates and crevassing typically dominate over thermally-driven processes (Benn et al., 2007), potentially obscuring the lake
 temperature signal we observe here. This has implications for the application of our findings to other lake-contact margins;
 thermally-driven calving may be most significant for ice flux at slow-moving margins, which could become increasingly
 prevalent as the GrIS continues to thin and potentially decelerate in response to ongoing climate change (Tedstone et al.,
 585 2015).

Importantly, our findings may have broader implications for the future evolution of lake-contact margins of the GrIS. As
 proglacial lakes expand around the GrIS, the influence of lake thermal processes on seasonal calving and terminus retreat
 may become increasingly significant, particularly at slow-flowing, grounded lake-contact margins where glacier dynamics
 are relatively passive (Carrivick et al., 2022). Non-ice-contact lakes in the Arctic have experienced progressively earlier
 590 spring ice break-up and longer ice-free seasons in response to atmospheric warming over recent years (Smejkalova et al.,
 2016). Satellite-derived lake surface temperatures suggest that these trends extend to ice-marginal lakes in Greenland, where
 the onset of positive thermal gradients is advancing by approximately five to eight days per decade, while the end of this
 warmer phase with positive stratification has remained comparatively stable (Carrea et al., 2025). If the period of positive
 thermal stratification is lengthening, and thermally-driven undercutting is most effective during stratified conditions as our
 595 observations suggest, a longer stratification season may translate into a prolonged window of enhanced calving activity at
 lake-contact margins, and thereby faster rates of ice-margin retreat (see Mallalieu et al., 2020).

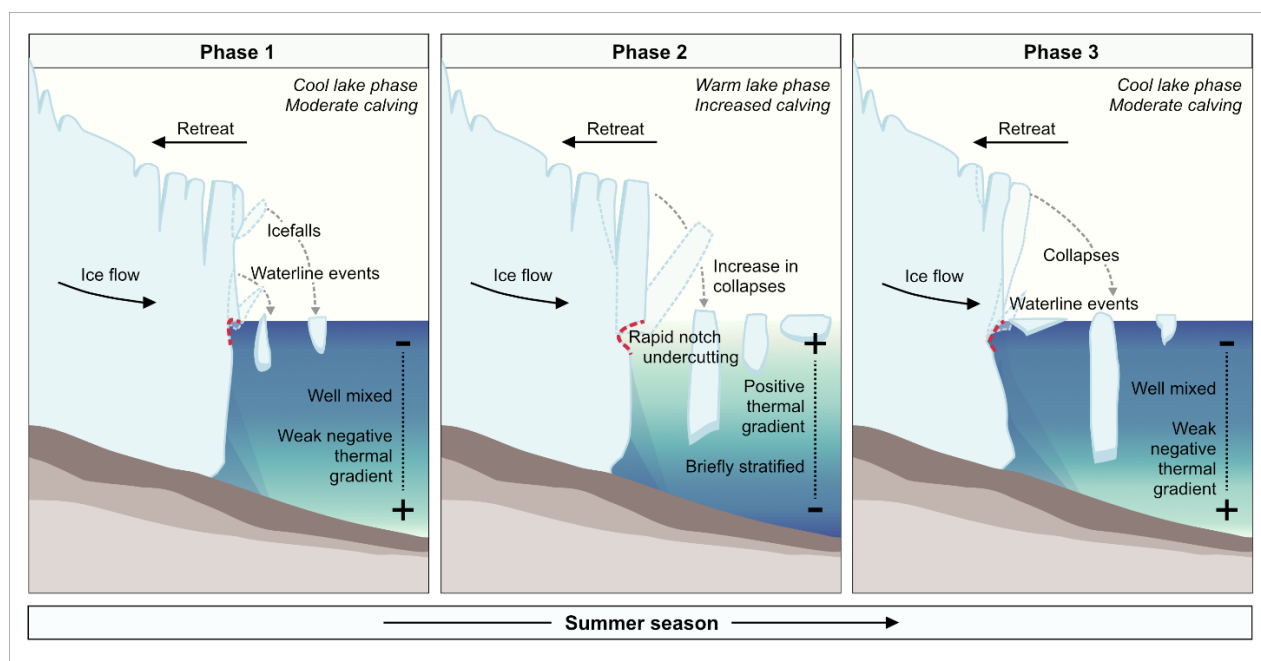


Figure 13: Schematic diagram of the three phases of lake thermal regime and associated calving behaviour.



5 Conclusions

600 We present concomitant records of lake water temperature and calving frequency, supplemented with measurements of meteorological conditions and modelled surface melt, at a lake-terminating outlet glacier in west Greenland. Our observations reveal a sub-seasonally variable lake thermal regime featuring a distinct, albeit brief, warm phase characterised by surface water temperatures above 4°C and the development of a positive thermal gradient. Calving was dominated by waterline and collapse events, consistent with a regime driven primarily by notch development and melt-undercutting at the ice-water interface. Calving frequency was highest during the warm lake regime phase, which was also the period when the association between lake water temperature and daily calving rate was strongest in the most comprehensive of the statistical models. These findings suggest that intraseasonal changes in lake thermal structure can modulate calving behaviour, likely by enhancing rates of melt undercutting when surface waters exceed 4°C. Since these observations derive from a relatively slow-flowing and grounded glacier margin, they indicate that sub-seasonal variations in lake thermal regime can influence calving activity where dynamic controls are assumed to be comparatively weak. Continued atmospheric warming is expected to favour longer ice-free seasons and therefore extended periods of thermal stratification in Greenland's ice-marginal lakes. If thermally driven undercutting is enhanced under such conditions, lake thermal regime may become an increasingly important control on retreat and mass loss at lake-contact margins of the Greenland Ice Sheet.

Data availability

615 All data will be made available via the UK Polar Data Centre upon publication.

Author contributions

The research was conceptualised by C.M.H with the help of M.W.S, D.J.Q, L.T and J.C. C.M.H collected the field data, with assistance from P.A.T, D.M.R, A.C.S, D.J.Q, J.M, J.A, L.D.R and H.B. C.M.H performed statistical modelling with guidance from M.K. C.M.H analysed the data, prepared the figures and wrote the manuscript, with input from M.W.S, 620 D.J.Q, L.T, J.C. and A.C.S. All authors reviewed the manuscript.

Competing interests

The authors declare that they have no conflict of interest.

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