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### A proxy-based model of glacier thermal characteristics

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| Complete List of Authors:     | Martinez Gracey, Dia; Simon Fraser University, Earth Sciences<br>Flowers, Gwenn; Simon Fraser University, Earth Sciences<br>Jacquemart, Mylène; ETH Domain Federal Institutes of Technology, D-BAUG; Swiss Federal Institute for Forest Snow and Landscape Research WSL                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                                     |
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# A proxy-based model of glacier thermal characteristics

Dia MARTINEZ GRACEY<sup>1</sup>, Gwenn E. FLOWERS<sup>1</sup>, Mylène JACQUEMART<sup>2,3</sup>

<sup>1</sup>*Simon Fraser University, Department of Earth Sciences, Burnaby, BC, Canada*

<sup>2</sup>*Laboratory of Hydraulics, Hydrology and Glaciology, ETH Zürich, Zürich, Switzerland*

<sup>3</sup>*Swiss Federal Institute for Forest, Snow and Landscape Research, Sion, Switzerland*

*Correspondence: Dia Martinez Gracey <dia\_martinez\_gracey@sfu.ca>*

**ABSTRACT.** This work presents a spatially lumped, computationally efficient model for estimating glacier thermal characteristics at a regional scale. We formulated proxies for the major englacial heat sources (meltwater refreezing and vertical shear strain heating) and heat transport mechanisms (downstream advection and vertical diffusion). We subsequently calculated these proxies for all glaciers in Alaska (RGI-01) and Svalbard (RGI-07) for which sufficient input data are available (26,798 total). To relate glacier thermal characteristics to the computed proxies, we built a decision tree that classified the glaciers into five groups based on similar relative proxy values. We then cautiously interpreted each group based on its dominant thermal proxies and correspondence with glaciers of known thermal structure. Based on these interpretations, temperate glaciers would represent most of the glacierized area in Alaska (62%), whereas cold-layer polythermal glaciers would represent 73% of glacierized area in Svalbard. One unexpected thermal group emerged, which appears to comprise many small glaciers and some large glaciers (8% of glacierized area) in Alaska with high-elevation accumulation areas, resulting in a thermal structure with cold ice in the accumulation area and temperate ice in the ablation area. This structure has not yet been documented in Alaska, and may warrant further attention.

1 INTRODUCTION

Thermal structure describes the enthalpy distribution within a glacier (e.g. Cuffey and Paterson, 2010). A glacier is said to be temperate if all ice is at the pressure melting point, cold if all ice is below the pressure melting point, and polythermal if it contains both temperate and cold ice. (e.g. Cuffey and Paterson, 2010). Enthalpy exerts a first-order control on ice-flow velocity through the flow-law coefficient  $A$ , which varies by two and three orders of magnitude, respectively, over the observed ranges of water content and ice temperature (Duval, 1977; Cuffey and Paterson, 2010). The flow-law exponent,  $n$ , may also be temperature-dependent (Schohn and others, 2025). Despite its influence on ice dynamics, enthalpy is neglected in most current-generation global glacier models used for future projections of glacier mass change and sea-level contribution (e.g. Zekollari and others, 2025).

Locally, the thermal influence on glacier flow and hydrology links thermal structure to flooding, glacier surging and other fast-flow hazards (e.g. Gilbert and others, 2012; Clarke and Blake, 1991). A temperate glacier bed, which permits ice–bed decoupling and therefore sliding, is required for nearly all forms of fast glacier flow including surging and glacier detachments (e.g. Kääb and others, 2021; Vincent and others, 2015). The response of glacier thermal structure to climate warming is non-uniform, with some glaciers expected to get warmer (e.g. Kachniarz and others, 2025) while others get colder (e.g. Wilson and others, 2013; Huss and Fischer, 2016; Beer and others, 2026). Thus, understanding present-day thermal structures provides a basis for projecting future thermal structures and their implications for glacier hazards under climate change.

Glacier thermal structure has traditionally been determined directly by in-situ temperature measurements (e.g. Classen, 1970; Brown and others, 2009; Jacquemart and others, 2025), inferred from radar scattering in ice (e.g. Hagen and Sætrang, 1991; Gacitúa and others, 2015; Delf and others, 2022; Kachniarz and others, 2025), or modelled using thermomechanically-coupled models (e.g. Aschwanden and others, 2012; Gilbert and others, 2014a; Wilson and Flowers, 2013). However, these methods are time-, cost-, and/or computationally intensive and thus have only been deployed on a small number of glaciers worldwide. As a result, little is known about how thermal structure varies from glacier to glacier or region to region, though much is assumed. Recently, glenglat, a comprehensive database of glacier temperature measurements, was developed to consolidate data and allow for analyses at larger scales (Jacquemart and others, 2025). In tandem, work is being done to make thermomechanically-coupled models faster in order



to provide regional-scale estimates of thermal structure (e.g. Jouvét and others, 2024; Bonnet and others, 2025).

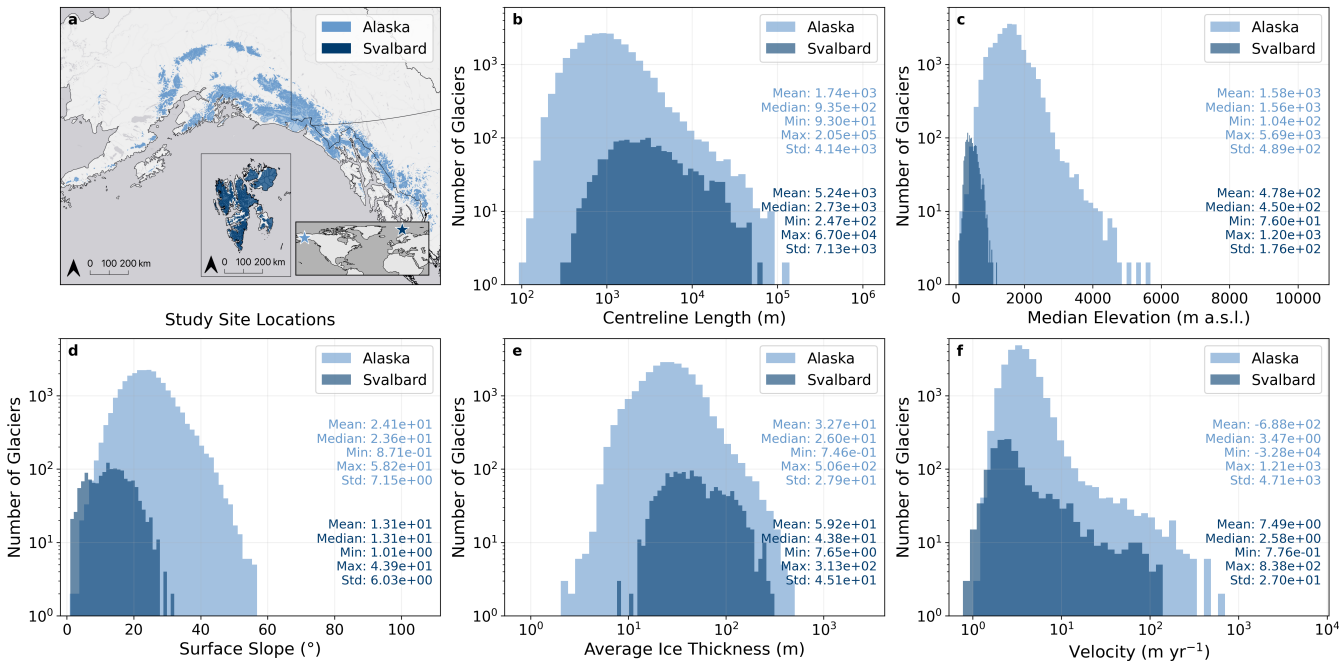
Both data and models have revealed a variety of polythermal regimes, such as those with predominantly temperate ice beneath a surface cold layer that extends to the bed only at the terminus (herein referred to as cold-layer polythermal), or those with predominantly cold ice and isolated regions of temperate ice near the bed (herein referred to as warm-core polythermal) (e.g. Irvine-Fynn and others, 2011). The presence and distribution of cold and temperate ice is determined by the relative importance of the heat sources and transport mechanisms within the glacier, which are in turn determined by both the external environment and internal glacier dynamics (e.g. Blatter and Hutter, 1991).

This work addresses the lack of regional-scale knowledge of glacier thermal structure by putting forth a simplified model to efficiently estimate the magnitude of major heat sources and transport mechanisms within a glacier. We do this by calculating proxy values for these quantities using publicly available datasets for glaciers across Alaska and Svalbard. We then use a decision tree to compare the individual proxy values in order to classify glaciers into five groups that can be associated with different glacier thermal structures.

## 2 STUDY AREA

In this study, we focus on the heavily glacierized regions Alaska (identified in the global Randolph Glacier Inventory 7.0 (RGI Consortium, 2023) as region 1 (RGI-01)) and Svalbard and Jan Mayen (RGI-07). Henceforth, “Alaska” refers to all RGI-01 glaciers for which proxies were calculated. This comprises glaciers that span southern Alaska, Yukon, and northwest British Columbia, but notably excludes glaciers in the Brooks and Aleutian Ranges (Fig. 1a), which suffer from insufficient data. “Svalbard” henceforth refers to all RGI-07 glaciers for which proxies were calculated. Glaciers on the islands Jan Mayen and Kvitøya (Fig. 1a) were excluded due to insufficient data. Alaska and Svalbard are globally significant contributors to sea-level rise, with Alaska contributing more to sea-level rise than any other region outside the ice sheets from 2000–19 (Hugonnet and others, 2021). Both regions also host a high concentration of surge-type glaciers (e.g. Sevestre and Benn, 2015; Lovell and others, 2026), further motivating our interest (e.g. Nolan and others, 2026).

Alaska and Svalbard exhibit a broad range of documented thermal structures (e.g. Jacquemart and others, 2025; Sevestre and others, 2015), reflecting the diversity of glacier geometry, dynamics, and environments across the regions (Fig. 1b-f). Alaska has a large climatic gradient from the maritime coast



**Fig. 1.** Study site and histograms of physical characteristics of glaciers in Alaska (light blue) and Svalbard (dark blue). (a) Location and relative size of Alaska and Svalbard glaciers. (b) Centreline length (RGI Consortium, 2023). (c) Median elevation from ASTER GDEM (Abrams and others, 2020). (d) Average surface slope (RGI Consortium, 2023). (e) Average ice thickness computed from Millan and others (2022) and Farinotti and others (2019). (f) Centreline velocity (Gardner and others, 2018; RGI Consortium, 2023). Statistics reported in (b)–(f) (mean, median, minimum, maximum, standard deviation) assume the units on the corresponding x-axes.

to the continental interior (Marcus and Ragle, 1970; Clarke and Holdsworth, 2002), whereas Svalbard experiences a more modest gradient with precipitation and temperature decreasing from the southwest to the northeast of the archipelago (Geyman and others, 2022). Nearly half of all glaciers in Svalbard are marine-terminating (Geyman and others, 2022), whereas <1% (56) glaciers in Alaska are marked as marine-terminating in RGI 7.0 (RGI Consortium, 2023). In Alaska, polythermal structures have been documented for a few small glaciers (e.g. Clarke and Goodman, 1975; Clarke and others, 1984; Wilson and others, 2013), whereas large valley glaciers have generally been assumed to be temperate (e.g. Clarke and Jarvis, 1976; Foy and others, 2011) despite a scarcity of direct observations (cf. Ochwat and others, 2021). Glacier thermal structures have been studied extensively in Svalbard, with both polythermal and cold regimes documented (e.g. Sevestre and others, 2015; Björnsson and others, 1996; Lovell and others, 2015; Jacquemart and others, 2025).

### 3 DATA AND METHODS

This work presents a spatially lumped, computationally efficient model in which a single proxy is calculated for each major heat source and transport mechanism per glacier. From these proxies, we gain insight into glacier thermal characteristics at a regional scale. Below, we describe the mathematical underpinnings of the model, how each proxy is calculated from readily accessible data, the decision tree used to group glaciers by their thermal characteristics, and the observations used to contextualize the resulting classification.

#### 3.1 Governing equation and simplifications

Glacier thermal structure is determined by heat sources and transport mechanisms as described by the conservation of energy (e.g. Aschwanden and others, 2012):

$$\rho_i \frac{\partial H}{\partial t} = \underbrace{Q}_{\text{heat production}} - \underbrace{\nabla \cdot (\rho_i \mathbf{v} H)}_{\text{advection}} + \underbrace{\nabla \cdot (k \nabla H)}_{\text{diffusion}}, \quad (1)$$

where  $\rho_i$  is the density of ice,  $H$  is enthalpy,  $t$  is time,  $\nabla$  is the gradient operator,  $\mathbf{v}$  is glacier velocity,  $k$  is the thermal conductivity of ice, and  $Q$  is the rate of heat production from all sources. Assuming ice is incompressible, the enthalpy gradient represents the change in internal energy of the system due to temperature changes in cold ice, or changes of water content in temperate ice.

This work is concerned with whether ice is temperate or cold, rather than its precise temperature or water content. When water content is not considered, the enthalpy of the ice is entirely due to sensible heat and related only to temperature. We therefore consider temperature, rather than enthalpy, as the thermodynamic quantity of interest, and Equation 1 becomes (van der Veen, 2013):

$$\begin{aligned} \frac{\partial T}{\partial t} \rho_i C = & \left. L_f M_f \right\} \text{meltwater refreezing} \\ & + \left. \begin{aligned} & \dot{\epsilon}_{xx} \sigma_{xx} + \dot{\epsilon}_{yy} \sigma_{yy} + \dot{\epsilon}_{zz} \sigma_{zz} \\ & + 2\dot{\epsilon}_{xy} \sigma_{xy} + 2\dot{\epsilon}_{xz} \sigma_{xz} + 2\dot{\epsilon}_{yz} \sigma_{yz} \end{aligned} \right\} \text{strain heating} \\ & - \left. u \frac{\partial T}{\partial x} \rho C - v \frac{\partial T}{\partial y} \rho C - w \frac{\partial T}{\partial z} \rho C \right\} \text{advection} \\ & + \left. k \frac{\partial^2 T}{\partial x^2} + k \frac{\partial^2 T}{\partial y^2} + k \frac{\partial^2 T}{\partial z^2} \right\} \text{diffusion}, \end{aligned} \quad (2)$$

where  $T$  is temperature,  $t$  is time,  $\rho_i$  is the density of ice,  $C$  is the heat capacity of ice,  $L_f$  is the latent

113 heat of fusion,  $M_f$  is the rate of melt refreezing,  $\dot{\epsilon}_{ij}$  is the strain rate tensor,  $\sigma_{ij}$  is the stress tensor,  $u$ ,  $v$ ,  
 114 and  $w$  are velocities in the  $x$ ,  $y$ , and  $z$  directions, respectively, and  $k$  is the thermal conductivity of ice. By  
 115 non-dimensionalizing and neglecting small terms including vertical advection, horizontal diffusion, and all  
 116 strain heating terms except that arising from vertical shear (see Supplementary Material §1), Equation 2  
 117 can be distilled into one dominant term for each heat source and transport mechanism:

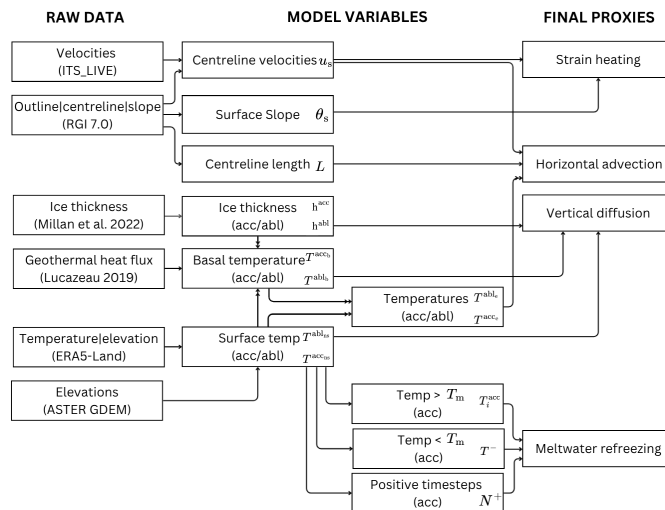
$$\begin{aligned} \frac{\partial T}{\partial t} \rho_i C = & \left. L_f M_f \right\} \text{ meltwater refreezing} \\ & + \left. 2\dot{\epsilon}_{xz} \sigma_{xz} \right\} \text{ strain heating due to vertical shear} \\ & - \left. u \frac{\partial T}{\partial x} \rho C \right\} \text{ horizontal advection} \\ & + \left. k \frac{\partial^2 T}{\partial z^2} \right\} \text{ vertical diffusion.} \end{aligned} \quad (3)$$

118 Here, meltwater refreezing and strain heating due to vertical shear are volumetric heat sources. Meltwater  
 119 refreezing describes the latent heat released when meltwater percolates into the permeable near-surface  
 120 (usually snow/firn) and refreezes, whereas the strain-heating term refers to the dissipative heat produced  
 121 due to ice deformation in the presence of basal resistance. Horizontal advection is the transport of heat due  
 122 to downstream ice flow, and vertical diffusion describes the conduction of heat due to vertical temperature  
 123 gradients within the ice column.

124 Thermomechanically coupled ice-flow models typically solve Equation 1 in two or three dimensions,  
 125 resulting in a distributed steady-state or transient enthalpy distribution within the glacier (e.g. Aschwanden  
 126 and others, 2012; Gilbert and others, 2014a; Benn and others, 2022, 2019). Our work, in contrast, constructs  
 127 proxies that can be calculated from existing datasets for each of the four terms on the right-hand side  
 128 of Equation 3, in an attempt to crudely but efficiently estimate glacier thermal characteristics without  
 129 thermomechanically coupled modelling.

### 130 3.2 Calculation of thermal proxies

131 Below we describe the proxy formulations for meltwater refreezing, strain heating, horizontal advection,  
 132 and vertical diffusion, as well as input datasets and their processing (see Fig. 2, Table 1). Each glacier is  
 133 partitioned into accumulation and ablation areas. For simplicity, meltwater refreezing is assumed only to  
 134 occur in the accumulation area, where porous snow and firn exist year-round, allowing meltwater percolation  
 135 and retention. All meltwater from the ablation area is assumed to run off, as glacier ice is generally less

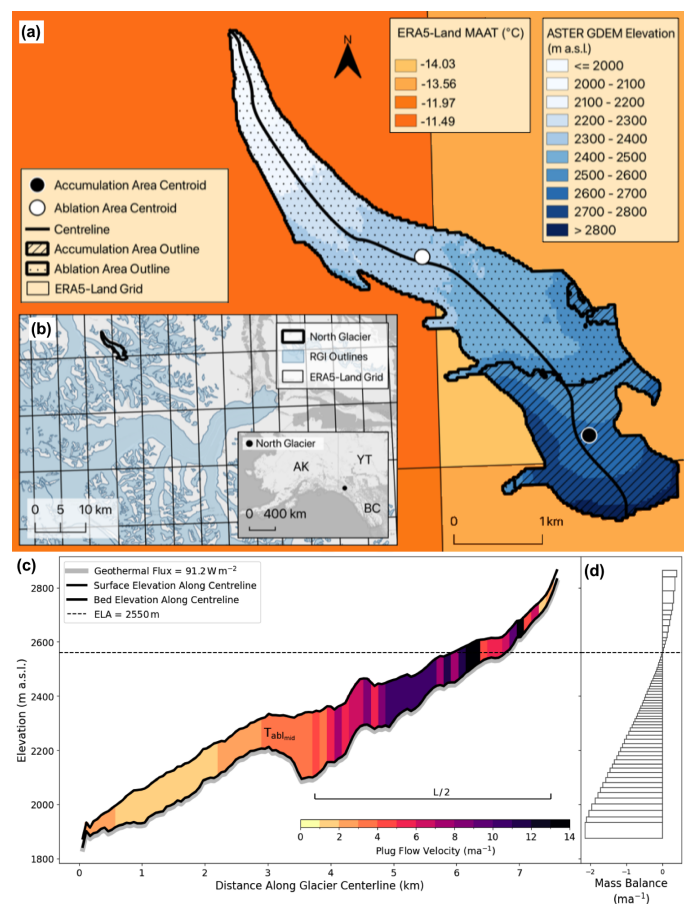


**Fig. 2.** Relationship between data sources, model variables, and final proxies. Arrows centered on symbols (e.g.,  $u_s$ ) apply to individual variables, whereas arrows between symbols (e.g.,  $T^{\text{acc}_e}$  and  $T^{\text{abl}_e}$ ) apply to all variables in a given box and are represented with double arrowheads. Note that any product that applies to accumulation and/or ablation areas (i.e. acc/abl) uses ASTER GDEM, the mass balance output from Rounce and others (2023), and RGI 7.0 outlines to define these areas.

permeable than snow or firn (e.g. Ryser and others, 2013). Conversely, diffusion is assumed only to occur in the ablation area, where the lack of a near-surface heat source permits steeper temperature gradients. Strain heating and advection are considered to be glacier-wide processes.

### Partitioning accumulation and ablation areas

Accumulation and ablation areas for each glacier are defined using the equilibrium line altitude (ELA) derived from the mass-balance model described in Rounce and others (2023). We use ERA5-forced model outputs provided to us directly (unpublished data courtesy of B. Tober and D. Rounce). We average this elevation-binned mass-balance data over all years (2000–19), and only elevation bins with valid data for every year are retained. The ELA is then taken as the mean elevation of the adjacent bins where mass balance changes sign. Accumulation and ablation areas are then defined as the subsets of ASTER GDEM cells within the RGI 7.0 outline containing elevations above and at or below the ELA, respectively. Fig. 3d shows an example mass-balance profile and its corresponding ELA. Accumulation and ablation areas need not be contiguous (Fig. 3a) and have areas  $a^{\text{acc}}$  and  $a^{\text{abl}}$ , respectively.



**Fig. 3.** Visualization of all datasets used in the proxy model, exemplified on North Glacier (Yukon, Canada). (a) Gridded ERA-5 Land mean annual temperatures (orange shading) behind the North Glacier outline, elevations (blue shading), centerline, and accumulation and ablation areas/centroids. (b) Location of North Glacier in the St. Elias Range (inset showing larger region in North America). (c) Average 2014–20 surface velocities along the RGI 7.0 centreline (shown in colour within the centreline profile), with geothermal heat flux (grey) at the glacier bed. (d) Elevation-binned 2000–20 mass-balance model outputs. See Table 1 for citations to the individual datasets.

### Accumulation- and ablation-area surface temperatures

All temperature-related variables in the proxy model come from the ERA5-Land 2 m-temperature product. As shown in Fig. 3a, the coarse grid-cell size of ERA5-Land data ( $0.5^\circ \times 0.5^\circ$ ) does not permit temperature differences within most glaciers to be resolved. To estimate surface temperatures for the accumulation and ablation areas of individual glaciers, we apply a lapse rate from the ERA5-Land cell elevation to the median elevations of the accumulation and ablation areas, respectively:

$$T_i^{\text{acc}} = T_i^{\text{accERA5-Land}} - \Gamma(z^{\text{acc}_{\text{med}}} - z^{\text{accERA5-Land}}), \quad (4)$$

$$T_i^{\text{abl}} = T_i^{\text{ablERA5-Land}} - \Gamma(z^{\text{abl}_{\text{med}}} - z^{\text{ablERA5-Land}}), \quad (5)$$

where  $T_i^{\text{acc}}$  and  $T_i^{\text{abl}}$  are the lapse-rate-adjusted temperatures of the accumulation and ablation areas, respectively, at each three-hourly timestep  $i$ ,  $T_i^{\text{accERA5-Land}}$  and  $T_i^{\text{ablERA5-Land}}$  are the raw ERA5-Land 2 m temperatures for the accumulation and ablation area, respectively,  $z^{\text{accERA5-Land}}$  and  $z^{\text{ablERA5-Land}}$  are the elevations of the ERA5-Land cells that contain the accumulation and ablation area centroids, respectively, which are obtained by dividing the ERA5-Land geopotential height by gravitational acceleration ( $g$ ),  $z^{\text{acc}_{\text{med}}}$  and  $z^{\text{abl}_{\text{med}}}$  are the median elevations of the accumulation and ablation areas, respectively, according to the 30 m ASTER GDEM, and  $\Gamma$  is the applied lapse rate ( $-6.5^\circ\text{C km}^{-1}$ ).

### Meltwater refreezing

The proxy for meltwater refreezing uses a temperature-index melt model and an estimate of the cold content of the snow which limits refreezing. The meltwater-refreezing proxy is calculated as

$$L_f M_f \approx \begin{cases} \frac{a^{\text{acc}}}{a^{\text{total}}} \left( \frac{q_m \wedge q_{\text{cc}}}{t_{\text{year}}} \right), & \text{acc area} \\ 0, & \text{abl area,} \end{cases} \quad (6)$$

where the energy associated with the available meltwater  $q_m$  is limited ( $\wedge$  indicating the minimum of the terms on either side) by the cold content of the snow  $q_{\text{cc}}$ , expressed as a rate using the number of seconds in a year  $t_{\text{year}}$ , and scaled by the ratio of the accumulation area  $a_{\text{acc}}$  to the total area of the glacier  $a_{\text{total}}$  under the simplifying assumption that refreezing occurs only in the accumulation area.

The amount of melt available to potentially refreeze is parameterized following Wilson and Flowers

| Symbol                              | Description                        | Value              | Units                             | Source       | Average | Used In  |
|-------------------------------------|------------------------------------|--------------------|-----------------------------------|--------------|---------|----------|
| Constants                           |                                    |                    |                                   |              |         |          |
| $\rho_i$                            | Ice density                        | 910                | $\text{kg m}^{-3}$                | CP2010       | –       | HA/SH    |
| $\rho_w$                            | Water density                      | 1000               | $\text{kg m}^{-3}$                | CP2010       | –       | MR       |
| $C$                                 | Heat capacity of ice               | $2.11 \times 10^3$ | $\text{J kg}^{-1} \text{K}^{-1}$  | CP2010       | –       | HA       |
| $k$                                 | Thermal conductivity of ice        | 2.1                | $\text{W m}^{-1} \text{K}^{-1}$   | –            | –       | VD       |
| $g$                                 | Gravitational acceleration         | 9.81               | $\text{m s}^{-2}$                 | –            | –       | SH       |
| $L_f$                               | Latent heat of fusion              | $3.34 \times 10^5$ | $\text{J kg}^{-1}$                | –            | –       | MR       |
| $T_m$                               | Surface melting temperature        | 273.15             | K                                 | –            | –       | MR       |
| $\Gamma$                            | Lapse rate                         | –6.5               | $\text{K km}^{-1}$                | –            | –       | MR/HA/VD |
| $\text{DDF}_{\text{snow}}$          | Snow degree-day factor             | 4.5                | $\text{mm d}^{-1} \text{°C}^{-1}$ | H2003        | –       | MR       |
| $D$                                 | Active layer                       | 4                  | m                                 | F1989, S1978 | S       | MR       |
| Model Variables                     |                                    |                    |                                   |              |         |          |
| $u_s$                               | Centreline surface velocity        | –                  | $\text{m s}^{-1}$                 | ITS_LIVE     | S/T     | HA/SH    |
| $L$                                 | Centreline length                  | –                  | m                                 | RGI 7.0      | –       | HA       |
| $\theta_s$                          | Glacier-wide surface slope         | –                  | radians                           | RGI 7.0      | S       | SH       |
| $r$                                 | Meltwater runoff fraction          | 0–1                | –                                 | –            | –       | MR       |
| $n_{\text{DD}}$                     | Number of degree days              | –                  | –                                 | ERA5-Land    | –       | MR       |
| $T^{\text{acc}}_e$                  | Average ice temperature (acc)      | Eqn. 18            | K                                 | ERA5-Land    | T       | HA       |
| $T^{\text{abl}}_e$                  | Average ice temperature (abl)      | Eqn. 17            | K                                 | ERA5-Land    | T       | HA       |
| $h^{\text{acc}}$                    | Ice thickness (acc)                | Eqn. 27            | m                                 | M2022, F2019 | S       | VA       |
| $h^{\text{abl}}$                    | Ice thickness (abl)                | Eqn. 26            | m                                 | M2022, F2019 | S       | HA       |
| $T^{\text{acc}}_{\text{ns scaled}}$ | Near-surface ice temperature (acc) | Eqn. 23            | K                                 | ERA5-Land    | T       | HA/VD    |
| $T^{\text{acc}}_b$                  | Basal ice temperature (acc)        | Eqn. 25            | K                                 | ERA5-Land    | T       | MR/HA    |
| $T^{\text{abl}}_{\text{ns}}$        | Near-surface ice temperature (abl) | Eqn. 19            | K                                 | ERA5-Land    | T       | HA/VD    |
| $T^{\text{abl}}_b$                  | Basal ice temperature (abl)        | Eqn. 24            | K                                 | ERA5-Land    | T       | HA/VD    |
| $q_{\text{cc}}$                     | Snow cold content (acc)            | Eqn. 10            | $\text{J m}^{-3}$                 | ERA5-Land    | T       | MR/HA    |
| $q_m$                               | Potential Refreezing               | Eqn. 7             | $\text{J m}^{-3}$                 | ERA5-Land    | T       | MR/HA    |
| $Q^{\text{acc}}$                    | Geothermal heat flux (acc)         | –                  | $\text{W m}^{-2}$                 | L2019        | –       | HA       |
| $Q^{\text{abl}}$                    | Geothermal heat flux (abl)         | –                  | $\text{W m}^{-2}$                 | L2019        | –       | HA       |

**Table 1.** Constants and variables used in the proxy model. “S” and “T” in column “Average” indicate spatial and temporal averages, respectively. “Used In” indicates the relevant proxy/proxies (MR = meltwater refreezing, SH = strain heating, HA = horizontal advection, and VD = vertical diffusion). “CP2010” is Cuffey and Paterson (2010), “F1989” is Fountain (1989), “S1978” is Schommer (1978), “H2003” is Hock (2003), “M2022” is Millan and others (2022), “F2019” is Farinotti and others (2019), and “L2019” is Lucazeau (2019). Labels “acc” and “abl” indicate accumulation area and ablation area, respectively.



(2013):

$$q_m = L_f (1 - r) \frac{\rho_w}{D} M^{\text{acc}}, \quad (7)$$

with the amount of accumulation-area melt  $M^{\text{acc}}$  in m w.e. converted to mass using the density of water  $\rho_w$ , distributed through the active layer  $D$ , scaled by the fraction of melt that runs off  $r$ , and converted to energy using the latent heat of fusion  $L_f$ . Parameter  $D$  is known to vary between glaciers, with measured values from 1.25 m in South Cascade Glacier, Washington (Fountain, 1989), to 7 m in Aletschgletscher, Switzerland (Schommer, 1978). We adopt an intermediate value of 4 m. The runoff fraction  $r$  is poorly constrained and set to an arbitrary and constant value of 0.2 (see e.g. Wilson and Flowers, 2013).  $M^{\text{acc}}$  is calculated as the average annual melt  $M_j$  for each year from 2000–20:

$$M^{\text{acc}} = \frac{1}{N_t} \sum_{j=1}^{N_t} M_j, \quad (8)$$

where  $N_t = 20$  years of data and  $M_j$  is the melt in the  $j^{\text{th}}$  year estimated using a temperature-index model (e.g. Hock, 2003):

$$M_j = DDF_{\text{snow}} \sum_{i=1}^{N_+} T_i^{\text{acc}} > T_m, \quad (9)$$

where  $DDF_{\text{snow}}$  is the degree-day factor for snow,  $N_+$  is the number of three hourly timesteps  $i$  with air temperatures ( $T_i^{\text{acc}}$ , Equation 4) above the melting point  $T_m$  in the  $j^{\text{th}}$  year of the ERA5-Land temperature data from 2000–20.  $DDF_{\text{snow}}$  is known to vary between glaciers, with Hock (2003) reporting a range of 2.7 mm d<sup>-1</sup> °C<sup>-1</sup> to 11.6 mm d<sup>-1</sup> °C<sup>-1</sup>. We adopt the median value from this dataset (4.5 mm d<sup>-1</sup> °C<sup>-1</sup>) for all glaciers.

The cold content  $q_{\text{cc}}$  is calculated as

$$q_{\text{cc}} = \rho_s C T^-, \quad (10)$$

where  $\rho_s$  is the density of snow,  $C$  is the heat capacity of ice, and  $T^-$  represents the average snow temperature from 2000–20:

$$T^- = \frac{1}{N_t} \sum_{j=1}^{N_t} T_j^-, \quad (11)$$

where  $N_t = 20$  years of data and  $T_j^-$  is the annual average of the minimum of air temperature and the

190 melting point (since snow temperature cannot exceed the melting point) for the  $j^{th}$  year:

$$T_j^- = \frac{1}{N_{3h}} \sum_{i=1}^{N_{3h}} (T_i^{\text{acc}} \wedge T_m), \quad (12)$$

191 with  $N_{3h}$  the number of three-hourly timesteps in a year,  $T_m$  the melting point, and  $T_i^{\text{acc}}$  defined as in  
192 Equation 4.

### 193 *Strain heating due to vertical shear*

194 The strain-heating proxy is calculated as

$$2 \dot{\epsilon}_{xz} \sigma_{xz} \approx u_s \rho_i g \sin(\theta_s), \quad (13)$$

195 where  $u_s$  is the glacier surface speed,  $\rho_i$  is the ice density,  $g$  is gravitational acceleration, and  $\theta_s$  is the  
196 glacier-wide average surface slope. This expression comes from considering the vertical shear-strain rate  
197  $\dot{\epsilon}_{xz}$  defined in terms of the vertical velocity gradient ( $\frac{\partial u}{\partial z}$ ), approximated using ice-surface speed  $u_s$  and  
198 thickness  $h$ :

$$\dot{\epsilon}_{xz} = \frac{1}{2} \left( \frac{\partial u}{\partial z} \right) \approx \frac{1}{2} \left( \frac{u_s}{h} \right), \quad (14)$$

199 and approximating the shear stress  $\sigma_{xz}$  as the basal shear stress (here assumed equal to the driving stress):

$$\sigma_{xz} \approx \rho_i g h \sin \theta_s. \quad (15)$$

200 For each glacier,  $u_s$  is calculated by extracting speeds from the ITS\_LIVE annual velocity mosaic for each  
201 year (2000–20 for Alaska and 2013–22 for Svalbard) using a spatial mask along the longest corresponding  
202 RGI 7.0 centreline, resulting in a profile of speeds similar to what is shown in Fig. 3c. The spatial median  
203 velocity along this centreline (with any NaN values from data gaps excluded) is calculated for each year,  
204 and then the temporal median across all years is used as  $u_s$  in the proxy model. We use the median rather  
205 than the mean to reduce the influence of extreme flow years (i.e. those associated with surges) in this  
206 relatively short dataset. The glacier-wide average surface slope,  $\theta_s$ , is taken directly from the RGI 7.0  
207 attribute table.

208 An alternative way to formulate the strain-heating proxy would have been to estimate strain rate using  
209 Glen's flow law and the driving stress. We explored this approach (see Supplementary Material §11) but

rejected it because it depends both on uncertain ice-thickness estimates and on specifying the flow-law coefficient  $A$ , which is temperature-dependent. In our study areas, where the vast majority of glacier temperatures are unknown, prescribing a value of  $A$  would amount to assuming the thermal structure we aim to infer.

#### Horizontal advection

The horizontal-advection proxy is calculated using the surface speed, as in the strain-heating proxy, and estimating the longitudinal temperature gradient using one temperature for the accumulation area and one for the ablation area:

$$-u \frac{\partial T}{\partial x} \rho C \approx -u_s \left( \frac{T^{\text{abl}_e} - T^{\text{acc}_e}}{L/2} \right) \rho_i C, \quad (16)$$

where  $T^{\text{abl}_e}$  is the englacial temperature of the ablation area,  $T^{\text{acc}_e}$  is the englacial temperature of the accumulation area,  $L$  is the length of the glacier,  $\rho_i$  is ice density, and  $C$  is the heat capacity of ice. For our purposes,  $u$  (on the left hand side of Equation 16) should represent the depth-averaged velocity, which is expected to be 80–100% of the surface speed  $u_s$ , depending on the relative contributions of deformation and sliding (Cuffey and Paterson, 2010). For simplicity, we assume plug flow and approximate  $u = u_s$ , determined as described in § 3.2. Centreline length  $L$  is taken directly from the RGI 7.0 centreline product attributes, where we select the longest of the available centrelines.

The englacial temperatures of the ablation and accumulation areas are estimated as the averages of their respective near-surface and bed temperatures:

$$T^{\text{abl}_e} = \frac{T^{\text{abl}_{\text{ns}}} + T^{\text{abl}_b}}{2} \quad (17)$$

$$T^{\text{acc}_e} = \frac{T^{\text{acc}_{\text{ns}}} + T^{\text{acc}_b}}{2}. \quad (18)$$

The near-surface temperature of the ablation area is estimated as

$$T^{\text{abl}_{\text{ns}}} = \frac{1}{N_t} \sum_{i=1}^{N_t} T_j^{\text{abl}_{\text{ns}}}, \quad (19)$$

where  $N_t = 20$  is the number of years of data from 2000–20 and  $T_j^{\text{abl}_{\text{ns}}}$  is an annual average temperature computed with temperatures above the melting point set to the melting point, because ice temperature

cannot exceed the pressure melting point:

$$T_j^{\text{abl}_{\text{ns}}} = \frac{1}{N_{3\text{h}}} \sum_{i=1}^{N_{3\text{h}}} (T_i^{\text{abl}} \wedge T_{\text{m}}), \quad (20)$$

where  $N_{3\text{h}}$  is the number of three-hourly timesteps in a year,  $T_i^{\text{abl}}$  is defined in Equation 5 and  $T_{\text{m}}$  is the melting temperature. The near-surface temperature in the accumulation area is calculated similarly as

$$T_j^{\text{acc}_{\text{ns}}} = \frac{1}{N_t} \sum_{i=1}^{N_t} T_j^{\text{acc}_{\text{ns}}}, \quad (21)$$

with a key difference being that after  $T_j^{\text{acc}_{\text{ns}}}$  is calculated as

$$T_j^{\text{acc}_{\text{ns}}} = \frac{1}{N_{3\text{h}}} \sum_{i=1}^{N_{3\text{h}}} (T_i^{\text{acc}} \wedge T_{\text{m}}), \quad (22)$$

it is then scaled by the ratio of potential refreezing to cold content:

$$T_{\text{scaled}}^{\text{acc}_{\text{ns}}} = T_j^{\text{acc}_{\text{ns}}} + \left( \frac{q_{\text{m}}}{q_{\text{cc}}} \wedge 1 \right) (T_{\text{m}} - T_j^{\text{acc}_{\text{ns}}} \wedge T_{\text{m}}), \quad (23)$$

where  $T_i^{\text{acc}}$  is defined in Equation 4,  $q_{\text{m}}$  is defined in Equation 7 and  $q_{\text{cc}}$  is the cold content of the snow (Equation 10). Equation 23 increases the near-surface temperature in the accumulation area in the presence of meltwater refreezing. In the limiting case of zero refreezing,  $\left( \frac{q_{\text{m}}}{q_{\text{cc}}} \wedge 1 \right) = 0$  and  $T_{\text{scaled}}^{\text{acc}_{\text{ns}}} = T_j^{\text{acc}_{\text{ns}}}$ . In the limiting case that refreezing entirely eliminates the cold content,  $\left( \frac{q_{\text{m}}}{q_{\text{cc}}} \wedge 1 \right) = 1$  and  $T_{\text{scaled}}^{\text{acc}_{\text{ns}}} = T_{\text{m}}$ .

Bed temperatures for the ablation and accumulation areas are calculated as

$$T^{\text{abl}_{\text{b}}} = T^{\text{abl}_{\text{ns}}} + h^{\text{abl}} \frac{Q^{\text{abl}}}{k} \quad (24)$$

$$T^{\text{acc}_{\text{b}}} = T_{\text{scaled}}^{\text{acc}_{\text{ns}}} + h^{\text{acc}} \frac{Q^{\text{acc}}}{k}, \quad (25)$$

where  $k$  is the thermal conductivity of ice,  $Q^{\text{abl}}$  is the geothermal flux in the ablation area,  $Q^{\text{acc}}$  is the geothermal flux in the accumulation area (both obtained from Lucazeau (2019)),  $h^{\text{abl}}$  is the mean ice thickness in the ablation area, and  $h^{\text{acc}}$  is the mean ice thickness in the accumulation area. In the interest of simplicity, ice thicknesses are computed as the average of values from the widely used global ice-thickness

245 datasets produced by Millan and others (2022) and Farinotti and others (2019):

$$h^{\text{abl}} = \frac{1}{2} \left( \frac{1}{N^{\text{Fabl}}} \sum_{i=1}^{N^{\text{Fabl}}} h_i^{\text{Fabl}} + \frac{1}{N^{\text{Mabl}}} \sum_{i=1}^{N^{\text{Mabl}}} h_i^{\text{Mabl}} \right) \quad (26)$$

$$h^{\text{acc}} = \frac{1}{2} \left( \frac{1}{N^{\text{Facc}}} \sum_{i=1}^{N^{\text{Facc}}} h_i^{\text{Facc}} + \frac{1}{N^{\text{Macc}}} \sum_{i=1}^{N^{\text{Macc}}} h_i^{\text{Macc}} \right), \quad (27)$$

246 where  $N^{\text{Fabl}}$ ,  $N^{\text{Facc}}$ ,  $N^{\text{Mabl}}$ , and  $N^{\text{Macc}}$  are the number of cells in the ablation or accumulation areas (abl  
247 or acc) of the Farinotti and others (2019) (F) or Millan and others (2022) (M) ice-thickness datasets, while  
248  $h_i^{\text{Fabl}}$ ,  $h_i^{\text{Facc}}$ ,  $h_i^{\text{Mabl}}$ , and  $h_i^{\text{Macc}}$  are the ice thicknesses in the  $i^{\text{th}}$  cells of the ablation or accumulation areas  
249 (abl or acc) of the Farinotti and others (2019) (F) or Millan and others (2022) (M) datasets.

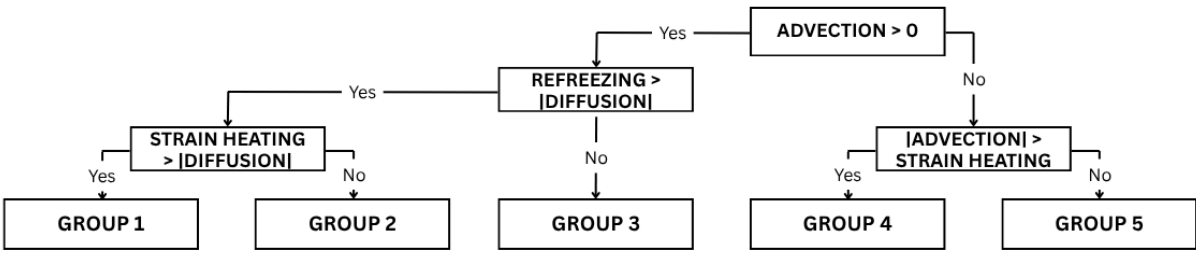
### 250 Vertical diffusion

251 The vertical-diffusion proxy is calculated by approximating the second derivative of temperature within  
252 the ice column using finite differences:

$$k \frac{\partial^2 T}{\partial z^2} \approx \begin{cases} 0, & \text{acc area} \\ \frac{a^{\text{abl}}}{a^{\text{total}}} \left( k \frac{T^{\text{abl}_{\text{ns}}} - 2T^{\text{acc}_e} + T^{\text{abl}_b}}{\left(\frac{h^{\text{abl}}}{2}\right)^2} \right), & \text{abl area,} \end{cases} \quad (28)$$

253 where  $k$  is the thermal conductivity of ice,  $T^{\text{abl}_{\text{ns}}}$  is the near surface temperature of the ablation-area ice  
254 (as defined in Equation 19),  $T^{\text{abl}_b}$  is the basal ice temperature in the ablation area (Equation 24),  $T^{\text{acc}_e}$   
255 is the englacial temperature of the accumulation area (Equation 18), and  $h^{\text{abl}}$  is the ice thickness in the  
256 ablation area (Equation 26). This quantity is scaled by the ratio of the ablation area  $a^{\text{abl}}$  to the total  
257 glacier area  $a^{\text{total}}$  in order to reflect our assumption that diffusion takes place primarily in the ablation  
258 area. This scaling enables a direct comparison between diffusion and the other whole-glacier proxies.

259 We estimate the ice temperature at the mid-depth of the ablation area using the englacial temperature of  
260 the accumulation area  $T^{\text{acc}_e}$ . This approach simplistically aims to account for the influence of ice advected  
261 from upstream on the vertical temperature profile (and hence on  $\frac{\partial^2 T}{\partial z^2}$ ) in the ablation area. In contrast,  
262 the englacial temperature of the ablation area  $T^{\text{abl}_e}$ , used in the advection proxy (Equation 16), is taken  
263 as the mean of surface and bed temperatures. If  $T^{\text{abl}_e}$  were used in the diffusion proxy, the temperature



**Fig. 4.** Decision tree used to sort glaciers into groups based on their thermal proxies.

gradient would be uniform and diffusion therefore zero.

**3.3 Documented thermal structures**

In an attempt to relate the calculated proxies to glacier thermal characteristics, and ultimately to glacier thermal structure, we examine glaciers with well-constrained thermal structures based on ice-penetrating radar measurements and/or in-situ temperature measurements.

Based on a review of the original literature, we assign a thermal-structure confidence rating to each glacier in Alaska and Svalbard that appears in the glenglat database of global glacier temperature measurements (Jacquemart and others, 2025) (see Table 2) as well as to glaciers with only radar data and no temperature measurements (which are currently not included in glenglat). The high, medium, and low confidence ratings are intended to describe bulk glacier thermal structure (temperate, cold, polythermal) based on the extent and nature of the relevant observations. Polythermal glaciers are further divided into cold-layer polythermal (temperate ice overlain by a cold layer, at least in the ablation area, that is thicker than the seasonal cold layer) or warm-core polythermal (extensive cold ice with localized temperate ice near the bed). Note that it is much easier to determine if a glacier is polythermal than if it is cold or temperate, as the latter two require spatially comprehensive data whereas the former may be detected with a single well-placed radar profile or borehole. Among the 18 of 34 glaciers in Alaska and Svalbard whose thermal structures we rate as high-confidence, we find that three of four are cold, one of nine is temperate, 12 of 15 are cold-layer polythermal, and four of six are warm-core polythermal.

**3.4 Glacier classification based on thermal characteristics**

In order to group glaciers of similar thermal characteristics based on the calculated proxies, we employ a manual decision tree (Fig. 4). A manual approach was selected after experimentation with multiple automated methods, including Gaussian Mixture Model clustering and Random Forest classification. The

Gaussian Mixture Model clustering produced poorly separated groupings that were unable to capture the relative proxy contributions that influence thermal structure (see Supplementary Material §3). The Random Forest classification was rejected due to limitations on model training and evaluation given the low number of reliable thermal-structure labels, exacerbated by the imbalance of labels between thermal-structure classes (see Supplementary Material §2).

In contrast to the automated approaches, the use of expert knowledge allowed us to create a simple decision tree that yields interpretable results. This reference tree was developed iteratively and is one of many configurations explored (see Supplementary Material §4). It aims to exploit obvious relative differences in the proxy values as well as the balance between heat sources and sinks. The first branch of the tree partitions positive and negative values of advection (or, equivalently, negative and positive values of diffusion). Where advection is negative, its magnitude is compared to strain heating as a means of evaluating the competition between heat sources and sinks as ice moves down glacier. Where advection is positive, the magnitudes of refreezing (heat source) and diffusion (heat sink) are first compared, followed by the magnitudes of strain heating (source) and diffusion (sink).

The reference tree has favourable characteristics such as a manageable and interpretable number of output groups (5) that each contain a meaningful number of glaciers ( $\mathcal{O}(10^2 - 10^4)$ ). Other trees that shared these attributes produced similar glacier groupings, suggesting the accessible insight is not unique to the reference tree (see Supplementary Material §4, Fig. 6). In contrast, more complex trees (e.g. with more layers) and those with meaningfully different decision criteria often produced numerous and imbalanced groups, some representing fewer than ten glaciers. Proxy sensitivities to model parameters and input data are characterized by perturbing model inputs and examining the resulting glacier classification using the reference tree described above (see Supplementary Material). We summarize the sensitivity-test results by calculating a “stability score” for each glacier which represents the fraction of all sensitivity tests yielding no perturbation to the glacier classification.

## 4 RESULTS

The proxy model was applied to 25,526 glaciers in Alaska (92.8% of RGI-01 glaciers) and 1,554 in Svalbard (92.7% of RGI-07 glaciers). The remaining glaciers had insufficient input data (see Supplementary Material §5). Below we describe the spatial distributions of the individual proxies, the classification results based on glacier thermal characteristics defined by the suite of proxies, and the comparison of model results with

documented glacier thermal structures.

#### 4.1 Proxy distributions across Alaska and Svalbard

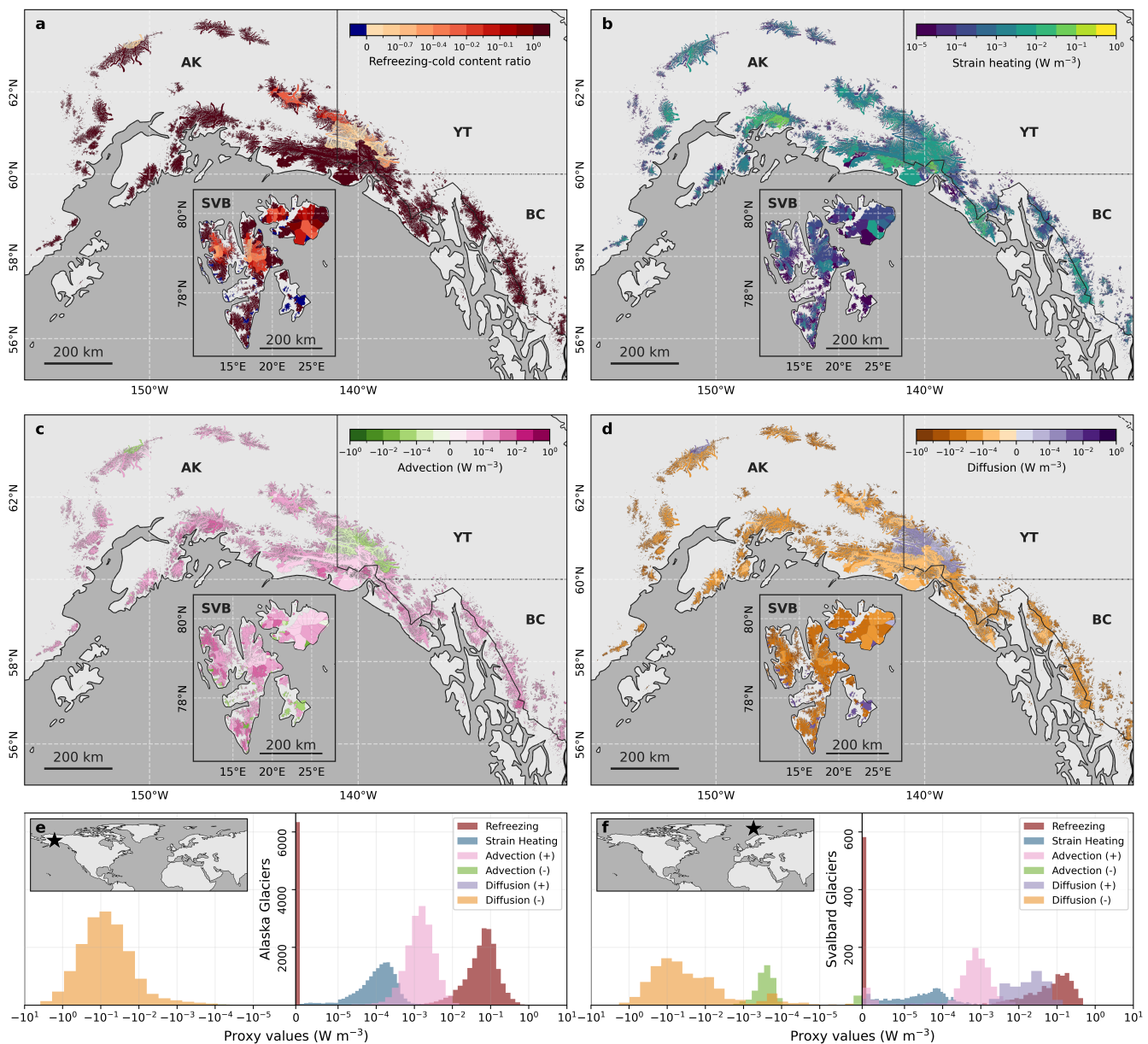
Clear spatial patterns appear in the map of refreezing–cold-content ratio (Fig. 5a). In Alaska, the refreezing–cold-content ratio is one across 81% of the glacierized area, suggesting that the amount of available melt is equal to the cold-content, leading to temperate accumulation areas. The ratio is zero for glaciers with no accumulation area which represent 2% of the glacierized area in Alaska, and between zero and one for the remaining 17% of glacierized area. These glaciers generally have high-elevation accumulation areas. Compared to Alaska, a smaller fraction of the glacierized area in Svalbard is characterized by a refreezing–cold-content ratio of one (46%) and a larger fraction has no refreezing (10%) (Fig. 5a inset). Glaciers representing the remaining 45% of glacierized area are found primary in northern Spitzbergen and across Nordaustlandet, and have high-elevation accumulation areas.

The strain-heating proxy (Fig. 5b) exhibits a heterogeneous distribution across Alaska and Svalbard, largely reflecting the distribution of glacier size and its expected association with speed. Exceptions include surge-type glaciers whose speeds, and therefore the associated values of strain heating, could be over- or underestimated depending on whether the glaciers were in the active or quiescent phases of the surge cycle during the observation period (see Supplementary Material §9).

Across both Alaska and Svalbard, most glacierized area (79%) is characterized by positive values of the advection proxy (Fig. 5c), indicating warmer ice in the accumulation area than in the ablation area. Note that most glaciers (66%) have average ablation area thicknesses greater than average accumulation area thicknesses, meaning that even temperate glaciers will have slightly colder ablation areas than accumulation areas due to the melting point depression. Glaciers with negative values of the advection proxy (colder ice in the accumulation than ablation area) in Alaska are either small ( $\lesssim 5$  km) or found in continental settings that tap into high-elevation accumulation areas where the refreezing–cold-content ratio is less than one. The spatial pattern of the diffusion proxy (Fig. 5d) mirrors that of advection, with the opposite sign as dictated by the model formulation (Equation 16). Most of the domain is therefore characterized by negative values of diffusion, representing conductive cooling of ice in the ablation area. Diffusion values are strongly controlled by ice thickness, with larger glaciers generally having lower values.

Histograms of the values of the four proxies across Alaska and Svalbard (Figures 5e and 5f) show that refreezing and diffusion are the dominant processes for most glaciers. Refreezing dominates strain heating



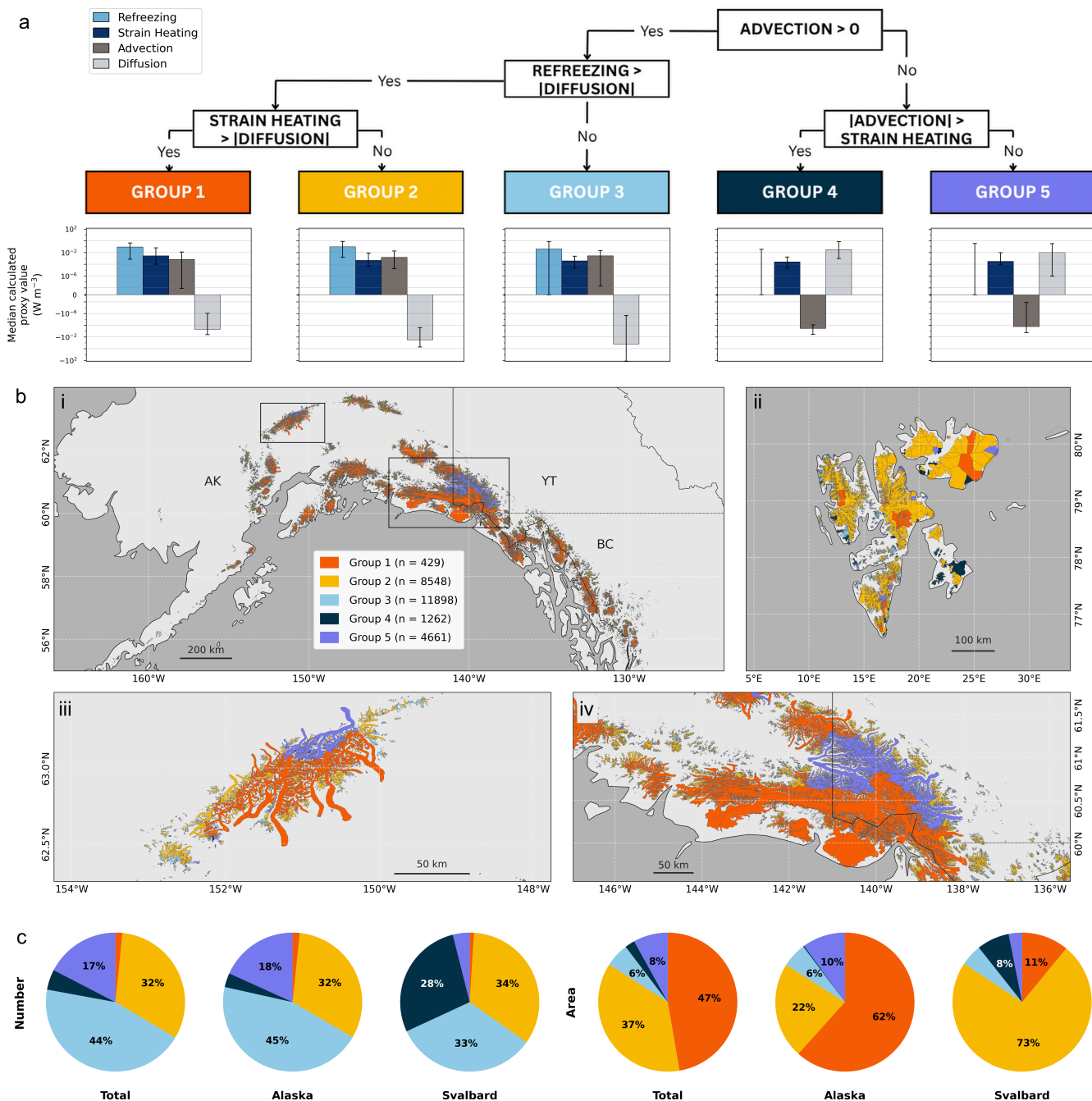


**Fig. 5.** Maps and histograms of proxies for Alaska (AK = Alaska, YT = Yukon Territory, BC = British Columbia) and Svalbard (SVB). (a) Refreezing-cold-content ratio. (b) Strain heating. (c) Advection. (d) Diffusion. (e) Logarithmic proxy histograms for all glaciers in Alaska with inset map (black star indicates location). The histograms of negative values of advection and positive values of diffusion are not visible at this scale. (f) Logarithmic proxy histograms for all glaciers in Svalbard with inset map (black star indicates location).

as a heat source, with values spanning similar orders of magnitude in both regions but having a 66% higher median in Svalbard. The refreezing proxy is zero for glaciers without an accumulation area (ELA above the glacier, based on 2000–20 mass-balance estimates from Rounce and others (2023)), representing 22% of all glaciers in Alaska and 32% in Svalbard (Fig. 5e and 5f). Diffusion is largely a cooling mechanism (negative values) and dominates advection in magnitude for most glaciers. The mean value of the diffusion proxy in Alaska is 2.6 times higher than in Svalbard, reflecting a greater proportion of thin glaciers. The mean magnitude of the advection proxy in Alaska is 2.5 times greater than that in Svalbard, primarily due to higher ice-flow velocities in Alaska.

## 4.2 Classification results

Based on the decision tree in Fig. 6a, glaciers in Alaska and Svalbard are classified into five groups. Group 1 comprises <2% of all glaciers, yet it accounts for 47% of glacierized area across Alaska and Svalbard (Fig. 6b-v, b-vi). The glaciers in this group occupy 62% of glacierized area in Alaska, but only 11% in Svalbard. They have the highest median velocity ( $46 \text{ m a}^{-1}$ ), thickness (151 m), and length (18 km), and lowest median elevation (1300 m) and slope ( $12^\circ$ ) of all groups (see Supplementary Material §7). Group 2 glaciers are both numerous and extensive, representing 32% of all glaciers and 22% and 73% of the glacierized area in Alaska and Svalbard, respectively (Fig. 6b-v, b-vi). On the basis of median values, Group 2 glaciers in Alaska are shorter (1.74 km), thinner (39 m), and slower ( $4 \text{ m a}^{-1}$ ) than Group 1 glaciers, but higher (1560 m) and steeper ( $22^\circ$ ) than glaciers in Groups 3–5. In Svalbard, Group 2 glaciers represent most of the large glaciers at low elevations (yellow in Fig. 6b-ii). Group 3 glaciers are the most abundant, comprising 44% (over 11,000) of all glaciers but only 6% of the glacierized area; they are predominantly very small (0.79 km median length) in both regions. Group 4 glaciers are abundant in Svalbard, accounting for 28% by number (c.f. 3% in Alaska), and have a median length of only 0.8 km. By median values, they are also the slowest ( $3 \text{ m a}^{-1}$ ) and thinnest (16 m) glaciers among the five groups. The majority of them have no accumulation area. Group 5 glaciers are more common in Alaska than in Svalbard (18% vs 3%). They account for 10% of the glacierized area in Alaska, primarily concentrated on the continental side of the St. Elias Mountains in Yukon, Canada (purple in Fig. 6b-iv), and on the north side of Denali in the Alaska Range (purple in Fig. 6b-iii). The remaining Group 5 glaciers in Alaska are comparatively small and scattered throughout the region.



**Fig. 6.** Decision-tree classification of glaciers and distribution of resulting groups across Alaska and Svalbard. (a) Decision tree and bar charts (see legend in upper left) of the median proxy values for all glaciers in each group, with whiskers showing the maximum and minimum values of each proxy in each group. (b) Map of glaciers in Alaska and Svalbard coloured by decision-tree group along with pie charts showing the number and area of glaciers in each group: (i) Map of all Alaska glaciers (AK = Alaska, YT = Yukon, BC = British Columbia) with legend showing the number of glaciers in each group across both Alaska and Svalbard, (ii) Map of glaciers in each group across Svalbard, (iii) Denali massif in Alaska, (iv) St. Elias sub-region of Alaska. (c) Pie charts showing the number of glaciers by group for the total sample set and for Alaska and Svalbard separately and glacierized area by group for the total sample set and for Alaska and Svalbard separately.

## 5 DISCUSSION AND INTERPRETATION

Our proxy model provides a highly simplified approach to assessing glacier thermal structures. The number of glaciers with documented thermal structure precludes a thorough quantitative evaluation of model accuracy, but we can nevertheless appraise the limitations of the modeling framework, uncertainties in both the input datasets and chosen parameter values (explored via sensitivity tests where possible), and the classification approach. Against the background of all limitations, we then use the available thermal-structure observations from Alaska and Svalbard to propose a cautious physical interpretation of the classification results.

### 5.1 Structural limitations of the model

The proxy model's simplified design necessitates assumptions that may not uniformly hold, such as restricting refreezing to the accumulation area or neglecting lateral shear strain heating. The lumped nature of the model and its application to a single centreline per glacier also fail to account for effects of complex geometry and multiple tributaries, which can create significant thermal complexity (e.g. Eisen and others, 2009). The restriction of refreezing to the accumulation area and diffusion to the ablation area is a further simplification that neglects potentially significant contributors to glacier thermal characteristics. For example, refreezing can be important in glaciers with heavily crevassed ablation areas (Colgan and others, 2016). In this case, the proxy model would underestimate refreezing and misrepresent accumulation–ablation-area temperature gradients, thereby incorrectly estimating both advection and diffusion (Equations 16 and 28).

Restricting refreezing to the accumulation area where there is snow present year-round is a conservative estimate of the amount of refreezing, but is expected to be reasonable for the majority of glaciers where significant refreezing does not occur in the ablation area (e.g. van Pelt and others, 2016; Østby and others, 2013). The crude approach to estimating refreezing in the proxy model using a single degree-day factor, active layer depth, snow and ice densities, and prescribed runoff fraction (Equations 7 and 9) is a further structural limitation. In reality, glacier thermal structure depends on the details of the surface and near-surface energy balance as well as percolation (e.g. Gilbert and others, 2014b,c) and other processes that are difficult to parameterize in a simple model and impractical to specify for a population of nearly 30,000 individual glaciers.

In pursuit of a parsimonious model applicable to the wide range of glaciers across Alaska and Svalbard,

we neglect horizontal diffusion and vertical advection in the proxy model, similar to simplifications that have been made in some thermomechanical ice-flow models (Wilson and others, 2013). While horizontal diffusion is arguably negligible on the basis of scaling, vertical advection may not be negligible depending on glacier characteristics (see Supplementary Material §1). The consequences of neglecting vertical advection in the model remain unexplored, in part due to its unclear influence on glacier-wide thermal characteristics and the challenge of devising an appropriate proxy representation. Finally, strain heating may be underestimated in the proxy model by neglecting contributions from lateral shear, which can lead to temperate ice near the valley walls (e.g. Wilson and others, 2013). Glaciers most affected would be those with high velocities that occupy narrow valleys which is not a common combination. Varying the strain heating proxy by  $\pm 30\%$  via changing the velocity inputs produced a change in the final classification of  $<2\%$  of glaciers representing  $<2\%$  of the glacierized area of Alaska and Svalbard (see Supplementary Material §8, Figures 46 and 59).

## 5.2 Data and parameter uncertainty

In addition to the structural limitations, uncertainty in the proxy model arises from several factors, namely temporal inconsistencies among datasets and simplifying assumptions in their application, the use of static glacier geometries, simplified delineation of accumulation and ablation areas, and the adoption of spatially and temporally uniform parameter values.

A fundamental challenge in this study is the mismatch between the long (decadal to centennial) timescales of glacier thermodynamic adjustment and the short (annual to decadal) time span of available data. There is an unavoidable inconsistency in aiming to predict present glacier thermal characteristics with data that do not span the timescales of thermal adjustment. Transience and temporal mismatch are through-going themes of the uncertainties discussed below.

We use the RGI 7.0 database (RGI Consortium, 2023) to define glacier outlines and centrelines. These products represent glaciers in the year 2000. Thus, over the 2000–20 study period, glacier area and length will be systematically overestimated as retreat has been widespread in both Alaska and Svalbard (Hugonnet and others, 2021). Overestimation of glacier lengths lead to an underestimation of advection magnitudes. However, the impact of centreline length estimation on final results should be minimal, since varying the centreline length used in advection from  $L/4$  to  $L$  results in  $<10\%$  of glaciers changing groups and  $<2\%$  change in glacierized area (see Supplementary Material §8, Figures 56 and 57).

Because RGI 7.0 represents glaciers in the year 2000, it likely that ablation areas are overestimated due

to terminus retreat. Any error in the estimation of accumulation or ablation areas impacts the refreezing and diffusion proxies, which are scaled by area fractions. Note also that the mass-balance (and ice-thickness) datasets were produced only for glaciers included in the RGI 6.0, meaning that the 413 glaciers newly added to RGI 7.0 lack sufficient data to be included in the proxy model (see Supplementary Material §5). We estimate ice thickness for the proxy model by averaging the Millan and others (2022) and Farinotti and others (2019) datasets. The underlying data used by Millan and others (2022) and Farinotti and others (2019) to estimate ice thicknesses differ, including in their time stamps, but broadly represent 2010–20. Considering the general decline in ice thickness during our 2000–20 study period, models using data from 2010–20 would tend to underestimate ice thickness relative to 2000–20 in the absence of model inaccuracy. Because observed surface velocities are used in ice-thickness models, ice thickness could be overestimated for glaciers that experienced anomalously high flow rates (e.g. surges) during the observation period. Overestimated ice thicknesses lead to underestimated diffusion magnitudes and basal temperatures, and vice versa, in the proxy model. It is difficult to determine if a relevant and systematic bias arises from these challenges. Varying input ice thicknesses by  $\pm 30\%$  results in 12% of glaciers switching classes, representing 7% of glacierized area (see Supplementary Material §8, Figures 42 and 43).

We use annual velocity mosaics from ITS\_LIVE to extract glacier flow speeds. ITS\_LIVE covers the full 2000–20 study period in Alaska but only 2013–20 in Svalbard. The Millan and others (2022) velocity dataset would have been a possible alternative, but represents the relatively short recent interval of 2017–18. By using velocities only from the longest RGI 7.0 centreline for each glacier, we likely overestimate glacier-wide flow speeds since velocities along the glacier centreline are typically higher than near the margins, and we might expect the longest tributary to flow the fastest. Overestimated velocities would lead to overestimated advection magnitudes. A tendency to overestimate strain heating should be minimized or countered by the proxy formulation, which underestimates the velocity gradient near the bed. We observed that varying the input velocity data by  $\pm 30\%$  has minimal effect, producing a change in the final classification of  $<0.5\%$  of glaciers representing  $<2\%$  of glacierized area (see Supplementary Material §8, Figures 46 and 59).

The air temperatures required by the proxy model are sourced from raw ERA5-Land data. For simplicity, we used the raw values at the native 9 km posting rather than attempting downscaling and/or bias correction. Because we apply a lapse rate to the ERA5-Land temperatures to adjust them to the median elevation of the accumulation and ablation areas, the model is sensitive to ERA5-Land temperature



inaccuracies, lapse rate values, and ASTER GDEM elevations.

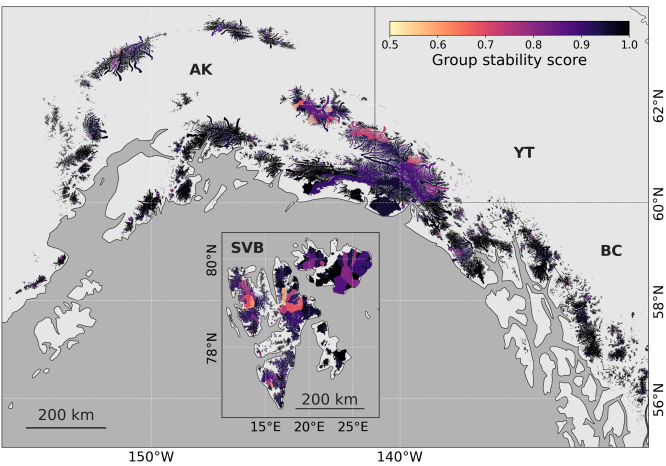
To explore this sensitivity, we vary both the raw ERA5-Land input temperatures by  $\pm 5^\circ\text{C}$  and the lapse rate from  $-2$  to  $-10^\circ\text{C km}^{-1}$ . ERA5-Land temperature variations lead to the greatest number of glaciers switching groups, with the extreme  $-5^\circ\text{C}$  case resulting in 12% of glaciers representing 50% of glacierized area switching groups. A more modest range of  $\pm 1^\circ\text{C}$  resulted in  $< 5\%$  of glaciers representing 10% of glacierized area switching groups. Varying the lapse rate results in  $< 15\%$  of glaciers representing  $< 3\%$  of glacierized area switching groups (see Supplementary Material §8, Figures 28 and 29).

Similar to the air temperatures, we use the native global  $0.5^\circ$  geothermal flux data from Lucazeau (2019) without attempts at downscaling given the sparsity of heat-flux measurements in glacierized regions. Because the geothermal heat flux is used to calculate basal temperatures from surface temperatures and ice thicknesses, we expect uncertainties in surface temperature and ice thickness to dominate. This is confirmed by sensitivity tests, where varying the geothermal heat flux by  $\pm 30\%$  results in  $< 3\%$  of glaciers representing  $< 3\%$  of glacierized area switching groups (see Supplementary Material §8, Figures 38 and 39).

For practical reasons, we adopt uniform values of quantities that are known to vary in space and time, such as the air-temperature lapse rate ( $\Gamma$ ), runoff fraction ( $r$ ), firn-aquifer thickness ( $D$ ), ice density ( $\rho_i$ ), and snow density ( $\rho_s$ ). These choices greatly simplify the proxy calculations, but also introduce inaccuracies, particularly where spatial and temporal variability is significant. Runoff and firn aquifer thickness are both known to be highly variable (e.g. Schommer, 1978; Fountain, 1989). Sensitivities are explored for each of the above variables in Supplementary Material §8. Runoff fraction is the least constrained, and has a significant impact on model results when varied from the reference value of 0.2 up to a value of 0.9:  $< 5\%$  of glaciers but 30% of glacierized area switch groups. All other variables result in  $< 10\%$  of glaciers and glacierized area switching groups. Across all 99 sensitivity tests, 56% of glaciers remain in the same output group regardless of the varied values and 91% remain in the same group for  $> 90\%$  of sensitivity runs (Fig. 7).

### 5.3 Classification approach

We experimented with many decision trees to group glaciers by various thermal characteristics (see Supplementary Material §4) prior to deciding on the reference tree (Fig. 6). All seven alternative trees with five output groups yielded similar results, with each group retaining an average  $> 95\%$  of the same glaciers as in the reference tree. The classification is most sensitive to whether cold content is included in the de-



**Fig. 7.** Map of stability score for each glacier. Stability score is calculated as the fraction of all 99 sensitivity tests that result in no perturbation to glacier classification determined using reference parameter values and the reference decision tree (4).

cision tree (i.e. using cold content > refreezing as a split). When cold content is included with five groups prescribed, 70% of glacierized area and 32% of glaciers in Alaska and Svalbard are unhelpfully classified into the same group (see Supplementary Material §6, Fig. 12). When strain heating is excluded, 84% of glacierized area and 33% of glaciers in Alaska and Svalbard are classified into the same group (see Supplementary Material §6, Fig. 14). The reference tree was constructed with physically meaningful comparisons in mind (e.g. comparing heat sources and sinks), and its final form guided by some knowledge of general geographical patterns of thermal structure along with the desire to avoid groups with negligibly small numbers of glaciers. A more intuitive reference tree might have included a comparison between advection and refreezing on one side of the tree; however, because the majority of glaciers with negative values of the advection proxy (right hand side of Fig. 6a) have no accumulation area and thus no refreezing, this approach yields seven unbalanced groups with 18–11,989 glaciers each.

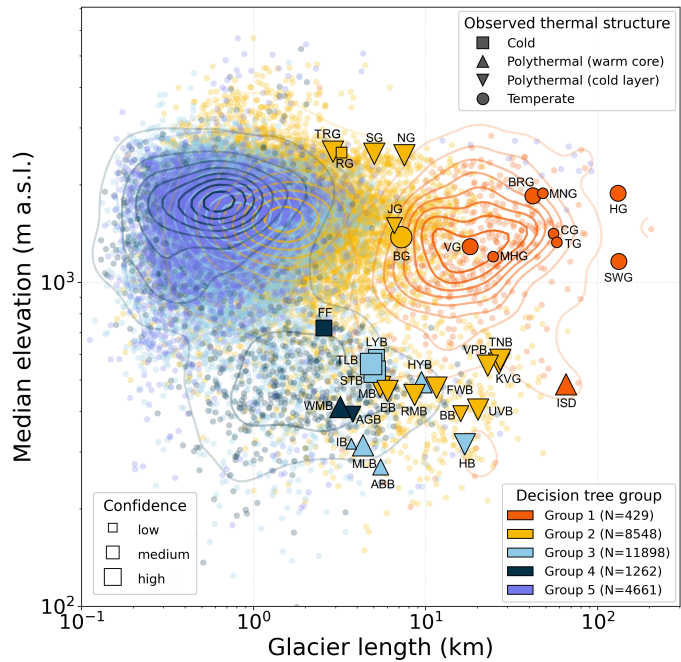
**5.4 Interpretation of groups in the context of known and expected thermal structures**

Below we cautiously associate each decision-tree group with the likely thermal structures it would include, based on the relative contributions of each proxy as well as an examination of the glaciers with known thermal structures. While there may be alternative pathways to the development of particular glacier thermal structures in reality, below we focus exclusively on the drivers of thermal characteristics included in our model.



**Table 2.** Classification results for glaciers with known thermal structures. Codes correspond to labels in Fig. 8. Documented thermal structures are temperate (T), cold-layer polythermal (P (CL)), warm-core polythermal (P (WC)), and cold (C). Confidence level is high (H), medium (M), low (L). No glaciers with known thermal structures landed in Group 5. Citations are indicated with superscripts: <sup>1</sup>Brown and others (2009), <sup>2</sup>Harrison and others (1975), <sup>3</sup>Bindschadler and others (1976), <sup>4</sup>Kienholz and others (2020), <sup>5</sup>Gulley (2009), <sup>6</sup>Motyka and Truffer (2007), <sup>7</sup>Rasmussen and others (2011), <sup>8</sup>Pelto (2011), <sup>9</sup>Jarvis and Clarke (1975), <sup>10</sup>Wilson and others (2013), <sup>11</sup>Sevestre and others (2015), <sup>12</sup>Hodgkins and others (1999), <sup>13</sup>Delf and others (2022), <sup>14</sup>Björnsson and others (1996), <sup>15</sup>Ødegård and others (1997), <sup>16</sup>Bacchin and Navarro (2026), <sup>17</sup>Ødegård and others (1992), <sup>18</sup>Jania and others (1996), <sup>19</sup>Moore and others (1999), <sup>20</sup>Murray and others (2000), <sup>21</sup>Gerbi and others (2021), <sup>22</sup>Navarro and others (2005), <sup>23</sup>Małecki and others (2013), <sup>24</sup>Watts and others (1997), <sup>25</sup>Karušs and others (2022), <sup>26</sup>Sobota (2011), <sup>27</sup>Bælum and Benn (2011), <sup>28</sup>Lovell and others (2015), <sup>29</sup>Classen (1970).

| Code | Glacier name                      | Documented structure | Conf. Level | Group |
|------|-----------------------------------|----------------------|-------------|-------|
| BG   | Bench Glacier <sup>1</sup>        | T                    | H           | 2     |
| BRG  | Black Rapids Glacier <sup>2</sup> | T                    | M           | 1     |
| VG   | Variegated Glacier <sup>3</sup>   | T                    | M           | 1     |
| SWG  | Seward Glacier                    | T                    | M           | 1     |
| MHG  | Mendenhall Glacier <sup>4</sup>   | T                    | L           | 1     |
| MNG  | Matanuska Glacier <sup>5</sup>    | T                    | L           | 1     |
| HG   | Hubbard Glacier <sup>6</sup>      | T                    | L           | 1     |
| CG   | Columbia Glacier <sup>7</sup>     | T                    | L           | 1     |
| TG   | Taku Glacier <sup>8</sup>         | T                    | L           | 1     |
| TRG  | Trapridge Glacier <sup>9</sup>    | P (CL)               | H           | 2     |
| SG   | South Glacier <sup>10</sup>       | P (CL)               | H           | 2     |
| NG   | North Glacier <sup>10</sup>       | P (CL)               | H           | 2     |
| TNB  | Tunabreen <sup>11,12</sup>        | P (CL)               | H           | 2     |
| VPB  | Von Postbreen <sup>11,13</sup>    | P (CL)               | H           | 2     |
| KVG  | Kongsvegan Glacier <sup>11</sup>  | P (CL)               | H           | 2     |
| UVB  | Uversbreen <sup>14</sup>          | P (CL)               | H           | 2     |
| FWB  | Finsterwalderbreen <sup>15</sup>  | P (CL)               | H           | 2     |
| RMB  | Ragna Mariebreen <sup>16</sup>    | P (CL)               | H           | 2     |
| MB   | Mettebreen <sup>16</sup>          | P (CL)               | H           | 2     |
| EB   | Erikbreen <sup>17</sup>           | P (CL)               | H           | 2     |
| HB   | Hansbreen <sup>18,19</sup>        | P (CL)               | H           | 3     |
| BB   | Bakinbreen <sup>20</sup>          | P (CL)               | M           | 2     |
| JG   | Jarvis Glacier <sup>21</sup>      | P (CL)               | M           | 2     |
| AGB  | Aldegondnebreen <sup>22</sup>     | P (CL)               | M           | 4     |
| HYB  | Horbyebreen <sup>23</sup>         | P (WC)               | H           | 3     |
| MLB  | Midtre Lovenbreen <sup>11</sup>   | P (WC)               | H           | 3     |
| ISD  | Isdomen <sup>24</sup>             | P (WC)               | H           | 1     |
| WMB  | Waldermalderbreen <sup>25</sup>   | P (WC)               | H           | 4     |
| ABB  | Austre Brogenbreen <sup>14</sup>  | P (WC)               | M           | 3     |
| IB   | Irenebreen <sup>26</sup>          | P (WC)               | L           | 3     |
| LYB  | Longyearbreen <sup>11</sup>       | C                    | H           | 4     |
| TLB  | Tellbreen <sup>27,28,11</sup>     | C                    | H           | 4     |
| STB  | Scott Turnerbreen <sup>12</sup>   | C                    | H           | 4     |
| RG   | Rusty Glacier <sup>29</sup>       | C                    | L           | 2     |



**Fig. 8.** Median elevation versus length for all glaciers in Alaska and Svalbard compared to glaciers with known thermal structures. Colour indicates the decision-tree group (orange = Group 1, yellow = Group 2, light blue = Group 3, dark blue = Group 4, purple = Group 5), shape indicates documented thermal structure (square = cold, triangle = warm-core polythermal, upsidedown triangle = cold-layer polythermal, circle = temperate), and symbol size indicates qualitative confidence in documented thermal structure. Codes are given in Table 2.

*Group interpretation based on relative proxy contributions*

Group 1 glaciers have strong heat sources, with the magnitudes of meltwater refreezing and strain heating exceeding that of diffusion (Fig. 6a). If the proxy values are realistic, these heat sources would produce temperate ice throughout the glacier. Glaciers in Group 1 are large and fast, and exist at relatively low elevations—characteristics consistent with some of the best-known temperate glaciers. This group covers most of the glacierized area in Alaska, but represents only a small number and area of glaciers in Svalbard. The dearth of Group 1 glaciers in Svalbard aligns well with expectations, as no fully temperate glaciers have been documented there to date. Eight out of nine documented temperate glaciers in Alaska land in Group 1 (Fig. 8), corroborating the association between Group 1 and temperate glaciers.

Group 2 glaciers are characterized by lower values of strain heating and higher values of diffusion than Group 1 glaciers, leading to less imbalance between heating and cooling terms (Fig. 6a). With less heating and more cooling through diffusion, we expect Group 2 glaciers to be largely polythermal. Group 2 glaciers comprise most of the glacierized area in Svalbard (6) and capture smaller, higher-elevation glaciers than those in Group 1 in Alaska. This result is consistent with the many documented and assumed polythermal

glaciers in Svalbard. Of the documented cold-layer polythermal glaciers, 12 of 14 fall within Group 2 (Fig. 8).

Group 3 glaciers have similar heat-source values to Group 2 glaciers, but diffusion is the largest-magnitude term, suggesting strong conductive cooling (Fig. 6a). We expect these proxies to be consistent with polythermal glaciers, but with more cold-ice content than Group 2 glaciers (e.g. warm-core polythermal). Of the documented warm-core polythermal glaciers, 4 of 6 fall in Group 3 (Fig. 8). Interestingly, 3 of 5 documented cold glaciers also fall into this group. Most Group 4 glaciers (97%) have no accumulation area and thus no refreezing, making strain heating the only heat source. With strain-heating magnitudes smaller than those of advection and diffusion (Fig. 6a), we would expect Group 4 glaciers to be largely cold. Three glaciers with different documented thermal structures (cold, warm-core polythermal, and cold-layer polythermal) land in this group. Further explanation is provided below (§5.4).

Group 5 glaciers experience high rates of refreezing and strain heating, with advection of opposite sign compared to glaciers in Groups 1–3, indicating colder accumulation than ablation areas. Among these glaciers, there are two distinct subgroups: those with and those without accumulation areas. The glaciers without accumulation areas are small and have higher magnitudes of strain heating than advection, suggesting they may be polythermal but dominated by cold ice because the effects of advection exceed those of strain heating. Those with accumulation areas tend to be large (>20 km long) and have cold content that exceeds the magnitude of refreezing. These glaciers have high-elevation accumulation areas and are most abundant in Alaska, concentrated on the northeast flank of Denali (the highest peak in North America) and northeast of Mount Logan (the highest peak in Canada). With significant heat sources, but high-elevation accumulation areas, we would expect polythermal glaciers with some cold ice in the accumulation area and temperate ice in the ablation area. While no glaciers with known thermal structures end up in this group (Fig. 8), there is some evidence that this structure describes Kaskawulsh Glacier, one of the large glaciers on the northeast side of the St. Elias Mountains. Firn cores and firn modelling have suggested that the accumulation area of Kaskawulsh Glacier has recently undergone a transition from cold to temperate in the 2000–19 period (Ochwat and others, 2021). Since our proxy model uses input data from the 2000–20 period, the average thermal state in the accumulation area would likely have been cold. While this particular thermal structure has not been directly documented in Alaska, it has been identified in the Alps (e.g. Suter and others, 2001).

548 *Glaciers with the same documented thermal structures that fall into different groups*

549 Glaciers deemed to be fully cold on the basis of observations end up in Groups 2, 3, and 4, with three  
 550 of five of these glaciers in Group 3 (Fig. 8). Rusty Glacier lands in Group 2, which includes most of the  
 551 documented cold-layer polythermal glaciers. Low confidence, however, is assigned to the determination of  
 552 Rusty Glacier's thermal structure which was based on measurements from a small number of boreholes  
 553 drilled in 1976, only three of which reached the glacier bed. Rusty Glacier is adjacent and similar in size to  
 554 Trapridge Glacier, which has been documented as cold-layer polythermal (Jarvis and Clarke, 1975). Three  
 555 cold glaciers in Svalbard (Longyearbreen, Tellbreen, Scott Turnerbreen) all end up in Group 3 which we  
 556 preferentially associate with warm-core polythermal glaciers. The thermal structures of these glaciers have  
 557 been documented through multiple lines of evidence, including ice-penetrating radar surveys and borehole  
 558 temperature measurements. The reason these cold glaciers are sorted into Group 3 is primarily because  
 559 they still have accumulation areas with nonzero values of refreezing and positive values of advection, so  
 560 they do not end up in Group 4 where most cold glaciers are expected to land.

561 Glaciers documented to have warm-core polythermal structures are the least well-sorted and end up in  
 562 Groups 1–4. Five of these seven glaciers end up in Group 3 which matches what we would expect based  
 563 on the proxy representations of heat sources and sinks. Waldermalderbreen ends up in Group 4 rather  
 564 than Group 3 largely because it has no accumulation area. More problematically, Isdomen ends up in  
 565 Group 1, which we have associated with temperate glaciers. Isdomen is part of the Austfonna icecap and  
 566 underwent a surge that is imprinted in the 2013–20 velocity data used in this study. The high velocities  
 567 associated with the surge lead to large overestimates of strain heating in the proxy formulation, a known  
 568 limitation of the model. If we instead assign Isdomen the velocity of the neighbouring glacier, it would  
 569 have more sensibly landed in Group 2 (cold-layer polythermal). The significant ice thickness of Isdomen  
 570 also contributes to its classification as Group 1 via the low magnitude of diffusion.

571 Glaciers with documented cold-layer polythermal structures end up in Groups 2, 3, and 4, with 13 of  
 572 15 in Group 2 as expected. Austegondenbreen and Hansbreen are the exceptions, with Austegondenbreen  
 573 in Group 4 because it has no accumulation area and Hansbreen in Group 3 because its magnitude of  
 574 diffusion exceeds that of refreezing. Hansbreen's low refreezing is largely driven by its small accumulation  
 575 area; without the area-weighting of refreezing, Hansbreen would have ended up in Group 2. Hansbreen's  
 576 thermal structure was documented in the late 1990s (Jania and others, 1996) and may have evolved since  
 577 this time. While we would normally expect the loss of accumulation area to produce a net cooling,

Hansbreen has been observed to be warming in the ablation area, due to a combination of loss of cold surface-ice (nearly 100 m in one location from 2007–21), increased snowfall insulating the ice from diffusive cooling, and increased refreezing in crevasses, which is not captured in our proxy model (Kachniarz and others, 2025).

Bench Glacier in Alaska is the only known temperate glacier that does not fall into Group 1. Instead it lands in Group 2, which we have associated with cold-layer polythermal glaciers. The relative low velocities and thin ice of Bench Glacier lead to low values of strain heating and high values diffusion compared to other glaciers in Group 1. The fact that Bench Glacier proxy values align more closely with cold-layer polythermal glaciers than temperate glaciers highlights the complexity of thermal structure and the limitations of this model. There may be other factors, such as refreezing in ablation-area crevasses or significant lateral shearing in this relatively narrow (1 km wide, 7 km long) glacier, that contribute to its temperate status. It is worth noting that a radar-transparent layer was observed in the ablation area of Bench Glacier, but has been confirmed to be temperate with borehole analysis (Brown and others, 2009).

While the proxy model cannot be thoroughly evaluated on the basis of 34 glaciers, we have nonetheless been able to make general associations between the five decision-tree groups and established thermal structures. Groups 1–4 represent a spectrum from warm to cold thermal structures. Groups 1 and 2 align strongly with documented temperate and cold-layer polythermal glaciers, respectively. Group 3 contains primarily warm-core polythermal and cold glaciers, the blend of which attests to the difficulty of distinguishing between mostly versus entirely cold thermal regimes. Group 3 status nevertheless provides a good general indication of an abundance of cold ice. Groups 4 is the least tidy, with most glaciers lacking an accumulation area and likely being far from equilibrium. These glaciers likely present the greatest violation of our model assumptions. Group 5 is interesting in that it captures none of the 34 glaciers with known thermal structures. Most Group 5 glaciers are small and lack an accumulation area, similar to Group 4 glaciers. However, a prominent subset of Group 5 is comprised of large continent glaciers with high-elevation accumulation areas. We hypothesize that Group 5 glaciers have cold ice at the highest elevations and temperate ice in the ablation areas. It is logistically challenging to determine the thermal structures of large glaciers tapping into high elevation mountain regions which might explain why Group 5 glaciers remain under-represented in the literature.

6 SUMMARY AND OUTLOOK

This study aimed to provide an efficient method for estimating glacier thermal characteristics at a regional scale. To this end, proxies for the major heat sources and transport mechanisms within a glacier were identified, and a decision tree used to classify glaciers based on their thermal characteristics. When applied to 26,789 glaciers in Alaska and Svalbard, five groups that emerged were cautiously associated with different glacier thermal structures, including temperate, cold and several types of polythermal. The group associated with temperate glaciers is the smallest in number (429 of 26,798 total glaciers) yet represents 62% of the glacierized area in Alaska. Polythermal glaciers dominate Svalbard, with one group representing 73% of its glacierized area and 34% of its glacier count. The most populous group of 11,898 is comprised of very small, likely polythermal glaciers that represent  $\leq 6\%$  of glacierized area across Alaska and Svalbard. Of the 34 glaciers in Alaska and Svalbard whose thermal regimes have been well documented in the literature, 25 of these land precisely in the groups expected. Those that do not are preferentially on the cold end of the spectrum, where small glacier size, lower-confidence observations and disequilibrium complicate model–data comparison. One group has no representation from the 34 glaciers, yet includes several large and well studied outlets of the St. Elias Mountains and Denali Range which we hypothesize are polythermal, with cold ice at higher elevations and temperate ice below.

This work was partially inspired but also hindered by the dearth of data pertaining to glacier thermal structure. A notable exception is the global englacial temperature database glenglat (Jacquemart and others, 2025). There remains a challenge, however, in characterizing glacier thermal structure from borehole data alone. Appropriately validated ice-penetrating radar data may be useful here, including reprocessing of data collected for ice-depth determination (Mannerfelt, Erik and 12 others, 2026). There is a tendency in the literature to classify glaciers as cold or temperate based on their environments or characteristics, without supporting data, thereby risking oversimplification or even misrepresentation of glacier thermal structure. Thermomechanically coupled modelling remains the best tool available to estimate glacier thermal structure in the absence of extensive field campaigns. Development and validation of statistical emulators (e.g. Jouvét and others, 2024; Bonnet and others, 2025) will allow these models to be applied at regional to global scales, providing an informative comparison with the proxy model presented here. Efficient and effective means of determining glacier thermal structure are required to answer some of the most interesting questions related to thermal evolution and glaciological hazards.

## 7 SUPPLEMENTARY MATERIAL

The Supplementary Material for this article has been submitted for review [doi upon publication].

## 8 DATA AVAILABILITY

Glacier outlines, centrelines, and slopes were obtained from RGI 7.0: <https://nsidc.org/data/nsidc-0770/versions/7>. Ice thickness data were obtained from the Millan et al. (2022) product: <https://www.sedoo.fr/theia-publication-products/?uuid=55acbdd5-3982-4eac-89b2-46703557938c> and Farinotti et al. (2019) product: <https://doi-org.proxy.lib.sfu.ca/10.3929/ethz-b-000315707>. Glacier elevations were obtained from ASTER GDEM products: [https://search.earthdata.nasa.gov/search/granules?p=C1711961296-LPCLLOUD&pg\[0\]\[v\]=f&pg\[0\]\[gsk\]=--start\\_date&fi=ASTER](https://search.earthdata.nasa.gov/search/granules?p=C1711961296-LPCLLOUD&pg[0][v]=f&pg[0][gsk]=--start_date&fi=ASTER). Temperature data were obtained from ERA5-Land 2m-temperature product: <https://cds.climate.copernicus.eu/datasets/reanalysis-era5-land?tab=download>. Geothermal heat flux data were obtained from the Lucazeau (2019) product: <https://agupubs-onlinelibrary-wiley-com.proxy.lib.sfu.ca/doi/full/10.1029/2019GC008389>. Data pre-processing, proxy calculation, and output group analysis code will be made public on github upon manuscript publication. Model inputs and outputs will be made available on Zenodo upon manuscript publication.

## 9 ACKNOWLEDGEMENTS

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## 10 AUTHOR CONTRIBUTIONS

GF conceived of the study and DMG/GF co-developed the details. DMG wrote and ran the model code, performed the analysis of model output in consultation with GF and MJ, and generated the figures. DMG led the manuscript preparation, with contributions from GF. All authors contributed to various aspects of the interpretation and edited the manuscript.

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