



Downslope windstorms in southeast Greenland - Part II: Multi-scale dynamics and spatial evolution of piteraq events

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Abstract. Severe downslope windstorms in southeast Greenland, known locally as piteraq, dominate extreme weather conditions across the region, including the main settlement Tasiilaq. However, their complex multi-scale dynamics remain poorly resolved in numerical weather prediction and global climate models, leading to inaccurate early warnings and biases in regional heat flux and ocean circulation projections. In this study, we utilize polar adapted, high-resolution reanalysis to trace the spatial, kinematic, and thermodynamic evolution of piteraq events from the synoptic scale down to localized turbulent wave breaking. Validation against regional observational networks shows that the reanalysis captures the broad spatial and temporal variability of the wind speed of these events, but it overestimates the peak intensities with a positive mean bias of 6.5 m s^{-1} during the winter cases.

We demonstrate that piteraq events are dynamically triggered by persistent upstream westerlies followed by the crossing of a Canadian cold vortex over Greenland. While approaching Greenland the Canadian cold vortex replaces a statistically significant, anomalous near-surface warming over the ice sheet to a rapid surface cooling. This transition establishes a strong cross-barrier surface pressure gradient (on median $7.3 \text{ hPa } 100 \text{ km}^{-1}$) and a reduction in the Scorer parameter profile that supports wave trapping and resonance. As the stably stratified flow is forced to descend the topography, it accelerates into intense low-level jets that are distinctly localized directly beneath regions of frequent wave breaking and overturning flow aloft. Our analysis further reveals a thermodynamically driven seasonal dichotomy. Winter events are characterized by dense, cold drainage flows that erode low-level coastal stratification and drive strong maritime cold-air outbreaks over the Irminger Sea. In contrast, summer events are characterized by a lack of windward blocking, allowing the upstream westerlies to force moister air masses over the crest. The resulting cloud formation and latent heat release, combined with subsequent adiabatic warming during the leeward descent, manifests as a relatively warm, dry windstorm capable of enhancing local ice melt.

Ultimately, our study defines a distinct, synoptic-scale blueprint for piteraq formation. Identifying these resolvable upstream precursors provides a pathway to train machine learning algorithms for improving the reliability of regional early warning systems, and offers a physical approach to dynamically parameterize sub-grid turbulent heat fluxes for more robust Atlantic Meridional Overturning Circulation (AMOC) projections.



1 Introduction

25 Piteraq events are commonly known as sudden, intense, and cold northwesterly winds in areas near the town of Tasiilaq, south-east Greenland (e.g., Cappelen and Jensen, 2021; Haualand et al., 2026). They are driven by a complex interaction between an upper-air cold vortex often situated near Baffin Island and a deep low-pressure system east of Greenland (Rasmussen, 1989). This synoptic setup forces cold, stable air to accelerate downslope before channeling through coastal fjords (e.g., Heinemann and Klein, 2002; Rasmussen, 1989). The most intense phase of these downslope windstorm events (DWE) is typically associated with pronounced low-level jets (LLJs) (e.g., Heinemann and Klein, 2002; Tollinger et al., 2019).

The disastrous piteraq of 1970 in Tasiilaq prompted the Danish Meteorological Institute (DMI) to establish a specialized piteraq warning service in 1972 (Rasmussen, 1989). Initially, the system relied on 500 hPa wind charts and surface pressure differences between the towns of Nuuk and Tasiilaq. By 1983, the integration of satellite observations allowed forecasters to detect large-scale foam formation and wave structures over the sea suggestive of piteraq events (Rasmussen, 1989). Building on these advances, Kaas and Isaksen (1989) introduced a forecast model with a 75-km grid that captured the broad coastal topography of Greenland, but lacked the resolution to resolve smaller-scale events. Consequently, localized mesoscale DWEs remained difficult to predict, occasionally resulting in unexpected and hazardous conditions for maritime operations in the Irminger Sea (Kaas and Isaksen, 1989).

Beyond their hazard potential, piteraq events are related to the initiation and intensification of Maritime Cold Air Outbreaks (MCAOs) in the Irminger Sea (Gutjahr et al., 2022). As these cold, dry air masses accelerate downslope and exit the fjords, they export coastal sea ice and encounter relatively warm waters, creating strong vertical gradients in temperature and moisture (Oltmanns et al., 2014). The gradients drive substantial sensible and latent surface heat fluxes from the ocean to the atmosphere. The enhanced turbulent fluxes locally deepen the maritime boundary layer over the Irminger Sea, amplifying the low-level baroclinicity between the piteraq outflow and the adjacent undisturbed maritime environment, thereby increasing the vorticity of the flow (Gutjahr et al., 2022). The combined effect can initiate the formation of mesocyclones or intensify existing systems into polar lows (e.g., Scorer 1988; Wos 1992). As a result, the process creates a positive feedback: the deepening low-pressure system strengthens the cross-barrier pressure gradient, which in turn further amplifies the DWE strength. The oceanographic consequences of piteraq events are profound as well. Although piteraq events are only intermittent and occur most frequently in winter, they are estimated to drive approximately 20% of the total loss of winter buoyancy in the upper ocean layers over the Irminger Sea (Oltmanns et al., 2014). The strong extraction of buoyancy leads to densification and subsequent sinking of surface waters (e.g., Gutjahr et al. 2022; Våge et al. 2009). The sinking of dense water masses is known to feed the lower limb of the Atlantic Meridional Overturning Circulation (AMOC), thereby linking the local atmospheric dynamics of southeast Greenland directly to the variability of large-scale global ocean circulation.

Recently, DMI implemented a specialized on-demand warning system involving a sub-kilometer resolution model designed to improve wind warnings. However, the system relies on computationally expensive dynamical downscaling to predict near-surface wind conditions, often without explicitly considering the underlying meteorological diagnostics of DWEs. While such high-resolution forecasting aids immediate coastal safety, the application of physical diagnostics to longer-term climate studies



remains constrained by model resolution and storage capacity. Current global reanalysis products and global climate models are too coarse to resolve the mesoscale dynamics associated with piteraq events (Lofverstrom et al., 2026). This lack of resolution implies that the atmospheric forcing driving sea-ice export and MCAOs in the Irminger Sea is likely underestimated in current climate projections. As argued by Lofverstrom et al. (2026), the primary mechanism limiting downslope windstorms detection is the smoothing of model topography, which alters the gravitational acceleration of the descending air and dampens the topographical funneling effects that are crucial for generating intense LLJs in narrow coastal fjords (DuVivier and Cassano, 2013; Heinemann and Klein, 2002). Furthermore, low spatial resolution fails to resolve sub-grid scale dynamic processes, particularly the breaking of large-amplitude mountain waves (Oltmanns et al., 2015). Because these non-linear wave dynamics are integral to the intensification of DWEs, their absence in coarse models leads to an underestimation of near-surface wind speeds (Lofverstrom et al., 2026), a limitation that threshold-based detection methods cannot overcome.

To avoid these modeling limitations and examine piteraq characteristics at a more localized scale, Haualand et al. (2026) used the network of in-situ observations in the Tasiilaq region to establish a definition for piteraq events, consistent with the criteria suggested by the DMI. The 67-year analysis distinguished seasonal evolutions, revealing that summer events exhibit a distinct "foehn-like" signature characterized by relative warming and a marked drying, with relative humidity dropping by up to 52% in 24 hours. They highlighted that piteraq events do not manifest uniformly across the region but are highly shaped by local topography. The spatial heterogeneity of a piteraq suggests complex flow structures that the available sparse point observations cannot fully resolve. Building upon Haualand et al. (2026), we present here a transition to a multi-dimensional analysis using Copernicus Arctic Regional Reanalysis (CARRA). By leveraging the polar-adapted, high-resolution reanalysis, we aim to move beyond surface-level statistics to investigate the full three-dimensional dynamical mechanisms and flow structures that drive piteraq events.

To identify the multi-dimensional structure of piteraq events within the CARRA dataset, we searched for a suite of meteorological diagnostics rooted in the governing mechanisms of isentropic descent, wave breaking, and hydraulic transitions. At the synoptic scale, identifying the potential for large-amplitude mountain waves typically requires the intersection of three dynamic conditions, as outlined in the global framework by Abatzoglou et al. (2021): (1) a sufficiently strong cross-barrier wind component, (2) high static stability at or above the ridge crest of the mountain, and (3) substantial lee-side subsidence. At the regional scale, existing approaches have relied on diagnostics grounded in the conservation of potential temperature during adiabatic descent (e.g., Plavcan et al. 2014; Steinhoff et al. 2014). This approach is helpful for distinguishing dynamically driven DWE such as foehn winds from radiation-driven katabatic flow. Finally, to assess the potential for wave trapping, lee-wave resonance, and potential rotor formation, we incorporate a theoretical diagnostic derived from hydraulic analogs (Smith, 1989). Specifically, we analyze the Scorer parameter (Scorer, 1949) and identify critical levels (Durrán, 1990) to diagnose the vertical structure required for wave breaking and flow stagnation (e.g., Metz and Durrán 2023; Tollinger et al. 2019). These theoretical metrics provide context to explain the intense, localized accelerations observed in the lee of southeast Greenland.

The primary objective of our study is to characterize the spatial and temporal evolution of piteraq events in southeast Greenland. By extending the analysis of Part I (Haualand et al., 2026) with the high-resolution CARRA dataset (1991–2024), we investigate the multi-scale processes driving piteraq events. The analysis follows the life cycle of these events: we first charac-



terize the synoptic-scale precursor patterns, before transitioning into a detailed investigation of upstream preconditioning and cross-barrier forcing. Specifically, we evaluate key dynamic and thermodynamic conditions such as vertical stability profiles and pressure gradients necessary to sustain mountain wave propagation and isentropic descent. Transitioning to the lee-side, we map the frequency and topographic dependence of LLJs, explicitly linking these acceleration zones to regions of high-frequency wave breaking. Collectively, these diagnostics define a high-resolution piteraq signature. Our study will serve two critical purposes: provide the physical feature set required for future seasonally dependent warning systems and improve the representation of Greenland's mesoscale forcing in the North Atlantic climate system.

2 Data

2.1 Copernicus Arctic Reanalysis (CARRA)

The Copernicus Arctic Reanalysis (CARRA) relies on a non-hydrostatic HARMONIE-AROME dynamical core (Bengtsson et al., 2017) adapted for high-latitude weather forecasting. CARRA is laterally forced by ERA5 (Hersbach et al., 2020) and includes three-hourly data at a 2.5 km horizontal resolution. The high resolution resolves the local- to meso-scale orographic processes, such as foehn winds (e.g., Haacker et al. 2024), gap flows (e.g., Kohnemann and Heinemann 2025, Moore and Imrit 2022), and mountain waves (e.g., Box et al. 2023), often parameterized or poorly resolved in coarser models. Furthermore, the structure of polar lows are better captured in CARRA (e.g., Hallerstig et al. 2024). CARRA likely excludes non-standard, low-elevation wind observations from ice-sheet automatic weather stations during data assimilation. CARRA utilizes the Surface Externalisée (SURFEX, Masson et al. 2013) land surface model, which includes a multi-layer snow scheme and an advanced sea ice model. To better describe the surface radiation coupling, SURFEX is adapted to use satellite-derived albedo, sea surface temperatures, and sea ice concentrations. These improvements resulted in better agreement with in situ observations (e.g., Køltzow et al., 2022), a more detailed representation of the surface energy balance, and a more accurate surface-atmosphere coupling (e.g., Frank et al., 2025; Slättberg et al., 2025), mitigating some of the warm biases over ice surfaces found in ERA5.

We use the CARRA-West domain, covering Greenland and surrounding seas, over the period 1991–2024. All 24 available vertical levels, spanning the surface to 500 hPa, are retained to capture the boundary layer and lower-tropospheric dynamics relevant to DWE development. From the surface level, we used 2 m air temperature (T_{2m}), 2 m relative humidity (RH_{2m}), and the 10 m zonal (u_{10m}) and meridional (v_{10m}) wind components. From the pressure levels, we utilized u , v , geopotential, air temperature (T), relative humidity (RH), and the geometric vertical velocity (w). From the height levels, we used T and RH and derived u and v from the available wind speed and direction at these levels.

2.2 Climate oscillation indices

To evaluate the synoptic-scale atmospheric preconditioning of piteraq events, we explored four climate oscillation indices. A variety of analytical approaches, including principal component analysis (PCA) and k -means clustering, are commonly used



to characterize such spatial atmospheric patterns, with input data typically sourced from reanalysis datasets or station records. For this study, the daily indices were acquired from the following sources:

125 The North Atlantic Oscillation (NAO) index was retrieved from the NCAR/UCAR Climate Data Guide (Hurrell et al., 2003). It is derived by applying PCA to sea-level pressure (SLP) measurements, with the index calculated using the leading principal component of SLP anomalies within the Atlantic domain (20°N–80°N, 90°W–40°E). PCA-based approaches generally provide a more comprehensive representation of spatial NAO patterns compared to traditional indices based on specific weather stations. However, it should be noted that the dynamic nature of PCA-based indices makes them subject to ongoing adjustments
130 as new data is integrated.

The Greenland Blocking Index (GBI) is derived from the 500 hPa geopotential height over the Greenland region (60°N–80°N, 80°W–20°W). The daily records were retrieved from the NOAA PSL/ESRL database (Hanna et al., 2016). Both the NAO and GBI time series originate from the NCEP/NCAR reanalysis dataset (Kalnay et al., 1996).

The Arctic Oscillation (AO) index was retrieved from the NOAA CPC. The index is constructed by projecting the daily
135 (00Z) 1000 hPa geopotential height anomalies poleward of 20°N onto the leading PCA mode pattern of the AO.

The Pacific/North American (PNA) index was also retrieved from the NOAA CPC. The calculation procedure is based on the Rotated Principal Component Analysis (RPCA) methodology utilized by Barnston and Livezey (1987). To obtain the underlying teleconnection pattern, the RPCA technique is applied to monthly standardized 500 hPa height anomalies, which isolates the primary teleconnection patterns across all months and allows the continuous daily time series to be constructed.

140 3 Methods

The CARRA dataset uses a Lambert conformal conic projection, with the wind components following the local grid orientation. Because the projected grid differs from the geographic north, we performed the grid rotation angle approach from MET Norway to adjust the wind components with the geographic north, as suggested in the CARRA Full System Documentation (Schyberg et al., 2020). We follow the definition of the meteorological season for winter (December, January, and February) and the
145 definition of the extended summer (June, July, August, and September) as in Haualand et al. (2026), hereafter just summer.

3.1 Interpolation of CARRA

To study the vertical atmospheric structure and validate the CARRA reanalysis with in situ data during piteraq events, the reanalysis data required both horizontal and vertical interpolation. Horizontally, we interpolated the CARRA grid to the coordinates of in situ observations using Delaunay triangulation with cubic interpolation via SciPy (Virtanen et al., 2020), consistent
150 with recent studies (Posch et al., 2024; Silva et al., 2024; van der Schot et al., 2024). For the automatic weather station (AWS) data at Tasiilaq (Jensen, 2025b), the nearest nine grid points ranged between a few 100 meters to 3.5 km distance. Because the interpolated CARRA orography at Tasiilaq is 58 m a.s.l. (only 5 m higher than the AWS), no lapse rate or wind profile corrections were applied. As geopotential height varies temporally and spatially, we vertically interpolated the 3D variables onto a fixed vertical grid with 10 m spacing.



155 3.2 Wind speed validation and topographic bias

To validate the reanalysis, the near-surface wind speed ($U = \sqrt{u^2 + v^2}$) was compared against the observational dataset in the region around Tasiilaq for the piteraq events detected by Haualand et al. (2026) between 1991 and 2024. The following observational datasets, the Programme for Monitoring of the Greenland Ice Sheet (PROMICE; Fausto et al., 2026), the Greenland Climate Network (GC-Net; Vandecrux et al., 2023), and the DMI (Jensen, 2025b), are only used to validate the reanalysis
160 output. Further description of the station data and references to applied methodologies is detailed in Part I of our study. Figure A1 shows the location and the availability of wind observations in the Tasiilaq region.

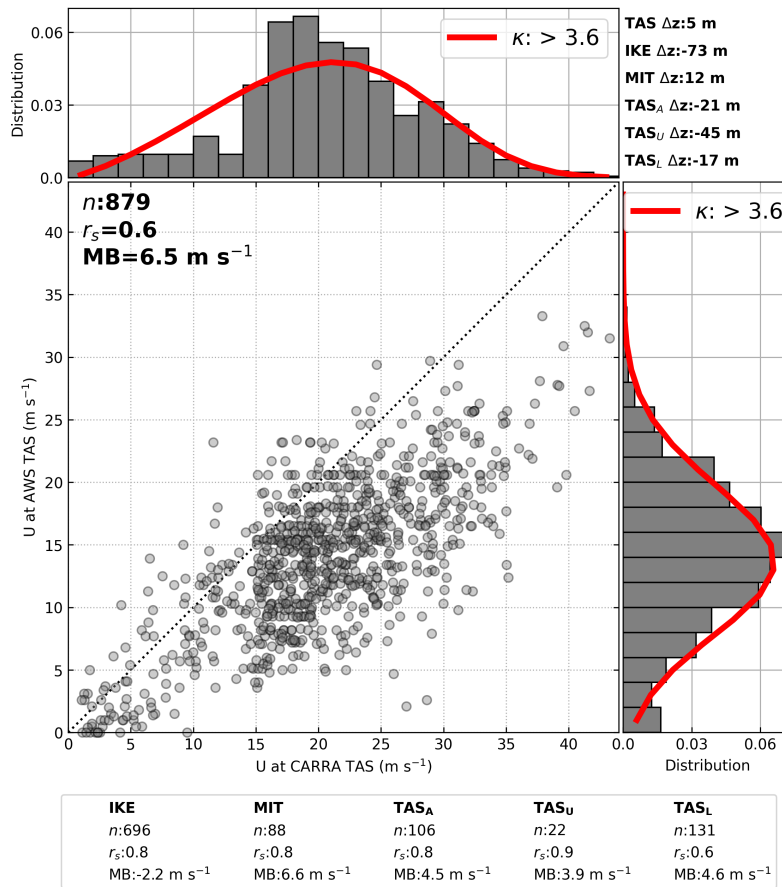


Figure 1. Observed vs. CARRA near-surface wind speed (U) during piteraq events detected between 1991 and 2024 at Tasiilaq (TAS), Ikermitt (IKE), Mittivakkat Gletsjer (MIT), and the Tasiilaq ice sheet transect: lower ablation zone (TAS_L), upper ablation zone (TAS_U), and near the equilibrium line (TAS_A). Sample size (n), Spearman correlation coefficient (r_s), mean bias (MB), and the height difference between the model and the station sensor ($\Delta z = z_{\text{CARRA}} - z_{\text{AWS}}$) are calculated and displayed.

As shown in Figure 1, CARRA captures the broader wind speed variability during piteraq events. However, CARRA exhibits a positive mean bias (MB) of 6.5 m s^{-1} , overestimating observed wind speeds, particularly during the most severe cases. While



periods of calm winds are observed during piteraq events at the AWS at TAS, these are less frequent in CARRA, as evidenced
165 by the sudden increase in the histogram at 15 m s^{-1} U threshold described in (Haualand et al., 2026). Winds below this
threshold are only present when separate events are merged due to an inter-event period of less than 48 h.

The U distribution was evaluated using maximum likelihood estimation to retrieve the shape parameter (κ) of the Weibull
distribution for both modeled and observed values, following Gorter et al. (2014). Both distributions display $\kappa > 3.6$, indicating
that they are left-skewed (negative skewness) with observations only partly concentrated around the arithmetic mean (Wilks,
170 2011). CARRA overestimates U at most stations in the region around Tasiilaq, with MB ranging from 3.9 m s^{-1} at TAS_U to 6.6
 m s^{-1} at the Mittivakkat Gletsjer (MIT). The highest MBs occur within the complex terrain of Ammassalik Island. While this
overestimation is broadly driven by smoothed surface roughness lengths and unresolved complex coastlines, the misrepresen-
tation of the local valley networks is likely the main culprit. At a 2.5 km horizontal resolution, narrow features such as Flower
Valley (Blomsterdalen), a northwest–southeast valley pointing directly toward Tasiilaq, is not resolved. Instead, the CARRA
175 topography merges the Flower Valley with the larger valley system situated north of the Mittivakkat Gletsjer. Such smoothing
results in an artificially wide, unblocked topographic channel in the model. By removing the sub-grid frictional barriers and
flow-blocking features present in reality, the model valley allows the simulated downslope winds to propagate toward Tasiilaq
with unrealistic efficiency, yielding the pronounced positive wind bias observed during peak piteraq intensities. Conversely,
CARRA representation of piteraq winds improves in less topographically complex areas, such as the offshore Ikermiit station
180 and the upper part of the Greenland Ice Sheet ablation zone, though shorter observational periods at the latter limit statistical
confidence. Furthermore, the seasonal-varying height of the anemometer at the ice-sheet weather stations typically below 5 m
a.g.l. (e.g., Fausto et al. 2026; Vandecrux et al. 2023) can exacerbate the bias during extreme wind events.

Consequently, applying the piteraq wind speed thresholds from Haualand et al. (2026) directly to the CARRA data results in
a higher number of detected events compared to the AWS record (CARRA precision = 37.6%). This discrepancy is not a model
185 artifact, as several detected CARRA events coincide with observational gaps in the AWS time series or represent intense events
that fell just below the station criteria. Validating this, we found numerous dates in the CARRA events that, while absent from
the TAS record, are corroborated by media reports detailing local wind damage (Haualand et al., 2026). Furthermore, several of
these CARRA events align with independent literature documenting extreme regional forcing during piteraq events, including
the formation of cold-air outbreaks in the Irminger Sea (Papritz, 2017), the development of severe polar lows (Kristjánsson
190 et al., 2011), and the triggering of extreme submarine melting at the Helheimgletsjer (Wheel, 2019).

Despite successfully capturing unrecorded events, we rely on AWS observations for baseline event detection as CARRA still
generates a substantial number of true false positives. The low precision is likely linked to the misrepresentation of micro-scale
topographic barriers in the Ammassalik Island at the 2.5 km grid scale, which allows even moderate, synoptically forced flows
to propagate to Tasiilaq as piteraq winds. In reality, such moderate flows are likely blocked or dissipated by the complex sub-
195 grid terrain. Consequently, the micro-scale topographic smoothing not only leads to a positive wind speed bias during actual
piteraq events, but also triggers localized false alarms under weaker synoptic forcing. The structural limitation further motivates
the multi-scale analysis, to identify upstream precursors helping to reliably distinguish true piteraq events from topographically
amplified model artifacts.



3.3 Iterative streamline integration

200 Streamlines were started in Tasiilaq and traced backward and forward following the air flow at 10 m a.g.l. For each piteraq event detected, we used the wind field at the time of the strongest wind occurring in Tasiilaq. There, we identified the wind flow direction and moved upwind to a new location 2.5 km away from Tasiilaq. At the new location, the algorithm identifies the four surrounding grid points and uses bilinear interpolation to calculate the new wind vector. The process iteratively repeats, tracing a streamline through the CARRA domain until the wind speed drops below 2 m s^{-1} or it reaches the reanalysis domain
205 boundary. The process was also applied forward (downwind). The final streamline is produced by merging the forward and backward components.

To assess the spatial uncertainty of our approach, the algorithm was applied to an ensemble of starting points offset from Tasiilaq by -0.05° to 0.05° longitude. The resulting ensemble showed high consistency, yielding a set of nearly parallel streamlines in the vicinity of Tasiilaq. Because the sensitivity test confirmed a robust flow path, the single primary streamline originating
210 directly from Tasiilaq was retained.

The density of streamlines for the strongest wind of each piteraq event in Tasiilaq is shown in Figure 2a. The density of streamlines suggests a preferential downslope corridor that spans approximately 100 km near the ice sheet ridge (along blue line in Figure 2b, around segment B) narrowing to around 50 km (over segment C) before reaching Tasiilaq. Downstream Tasiilaq, the streamlines curve northeast and diverge. Two distinct source regions feed the corridor close to the ice sheet ridge:
215 air drawn from low elevations in West Greenland and air descending from the high interior plateau.

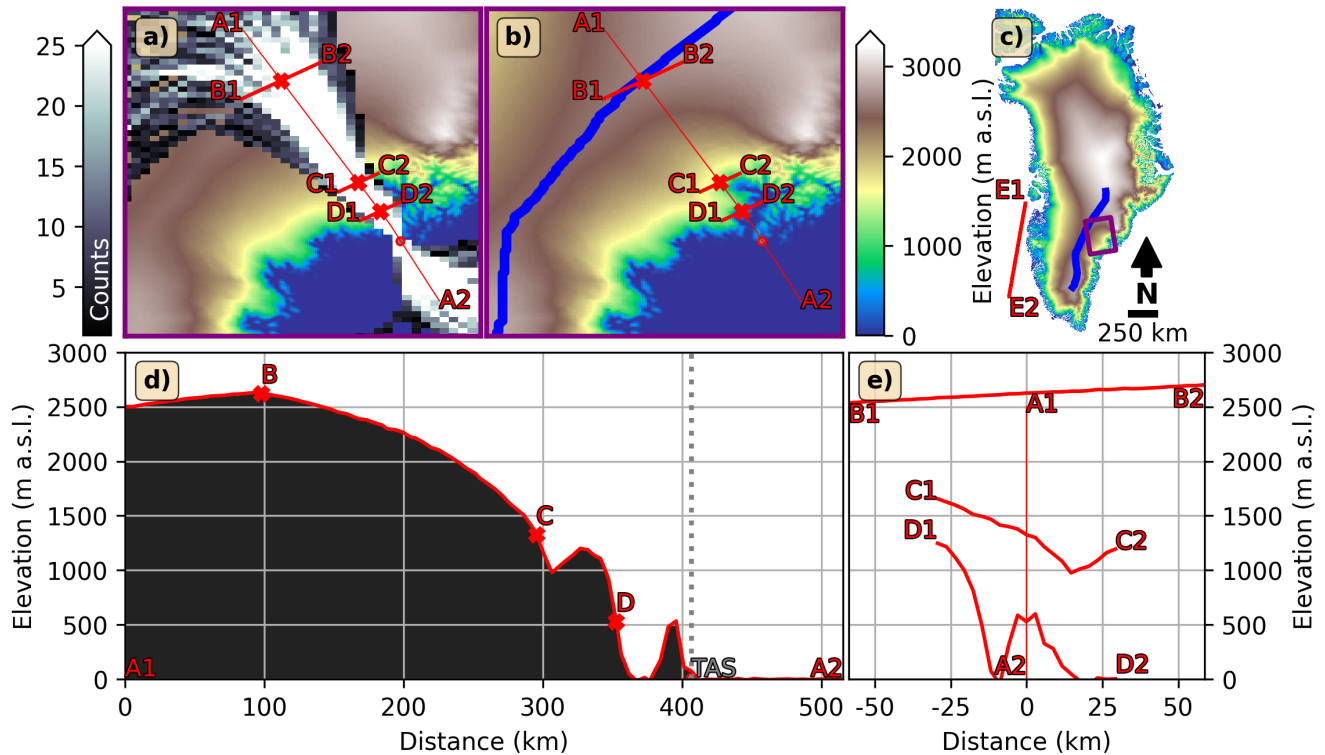


Figure 2. Topographic maps and elevation profiles of Tasiilaq piteraq events (1991–2024). Regional domain showing streamline density (a) and main segments (b) corresponding to the purple rectangle in the Greenland map (c). Segment A traces the primary streamline density, intersected by perpendicular segments B, C, and D. Over the Davis Strait, segment E (red line, c) is drawn parallel to the South Greenland crest (blue line, b–c). Topographic profiles along segment A (d) and segments B–D (e). The location of Tasiilaq (TAS) is marked by red dots in a–b and labeled in d.

To investigate the atmospheric structure during piteraq events, we define segment A along the axis of the highest streamline density passing through Tasiilaq (Fig. 2a). This 500-km transect extends from the high-elevation western ice sheet (A1) to the Irminger Sea (A2), over an elevation difference of approximately 2700 m (Fig. 2d). To examine the influence of surface geometry on the flow, we introduced three perpendicular cross-sections. Segment B is located at the highest, relatively flat part of the ice sheet from segment A (Fig. 2e). As the flow descends to roughly 1000 m a.s.l., it encounters a surface depression formed by a branch of the Helheimgletsjer, with an adjacent branch captured by segment C (Fig. 2e). Beyond the Helheimgletsjer branches, segment A traverses its steepest and most complex terrain as it drops toward Sermilik Fjord. This complexity is illustrated in segment D, which captures the narrow Qeertartivattaap Kangertiva (Johan Petersen Fjord) to the left and the broader Sermilik Fjord to the right (Fig. 2e). Finally, to characterize upstream synoptic conditions, segment E extends 850 km along the Davis Strait (Fig. 2c). Segment E is positioned 100 km parallel to the Greenland Ice Sheet crest, to ensure the flow remains representative of the atmospheric conditions prior to topographic deceleration or blocking.



3.4 Meteorological diagnostics

To characterize the multi-dimensional structure of piteraq events and identify the governing mechanisms of isentropic descent and wave breaking, several dynamic and thermodynamic variables and indicators were derived from CARRA state variables.

230 3.4.1 Thermodynamics and stability

Static stability was quantified using the Brunt–Väisälä frequency (N^2), which defines the frequency of oscillation for an air parcel displaced vertically in a statically stable atmosphere:

$$N^2 = \frac{g}{\theta} \frac{\partial \theta}{\partial z},$$

where g is the gravitational acceleration (9.81 m s^{-2}). Negative values of N^2 (where $\partial \theta / \partial z < 0$) indicate absolute instability and the potential for turbulent overturning.

235 To identify potential zones of turbulence generated by wind shear, we followed Stull (1988) to calculate the gradient Richardson number (Ri_g) between vertical levels:

$$Ri_g = \frac{N^2}{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2}.$$

In line with common meteorological practice, regions where $Ri_g < 0.25$ are identified as favorable for instability and turbulence.

3.4.2 Flow regimes and wave dynamics

240 To characterize the upstream flow regime, the upstream Froude Number (Fr) was calculated along segment E, parallel to and approximately 100 km upstream of the Greenland Ice Sheet ridge crest:

$$Fr = \frac{U}{NH},$$

where U is the undisturbed cross-barrier scalar wind speed, N is the upstream Brunt–Väisälä frequency, and H is the effective Greenland Ice Sheet height. In line with common meteorological practice, flows where $Fr \geq 1$ are considered supercritical.

To assess the potential for trapped lee waves and resonance, we applied the Scorer parameter (l^2) as defined by:

$$l^2(z) = \frac{N^2}{U(z)^2} - \frac{1}{U(z)} \frac{d^2 U(z)}{dz^2}.$$

245 Severe downslope windstorms generally rely on a specific thermodynamic profile: a highly stable layer near the mountain crest that caps the low-level flow (Durrán, 1986). This specific vertical structure, known as "Scorer-parameter layering", is characterized by a high-stability layer (large l^2) at low levels and a lower-stability layer (small l^2) aloft. This trapping mechanism facilitates the downstream propagation of wave energy rather than vertical dissipation (Scorer, 1949).



Intense downslope windstorms often feature complex non-linear flow responses, including upstream stagnation and lee-side
250 wave breaking (e.g., Kozmar and Grisogono, 2020). Because the zonal flow driving the piteraq events (Fig. 2) is strictly positive
in the cross-barrier direction ($u > 0 \text{ m s}^{-1}$), regions of potential flow stagnation and overturning are identified by flow reversal.
Consequently, atmospheric regions where $u < 0 \text{ m s}^{-1}$ (easterly flow) are classified as zones of upstream blocking (when
located windward of the crest) or turbulent wave-breaking (when located in the lee-side).

3.4.3 Cross-barrier forcing

255 The cross-barrier thermal forcing was defined as the difference in potential temperature between the crest and the lee-side
coastal Tasiilaq.

To quantify the pressure gradient force driving the flow, we calculated the mean sea-level pressure (mslp) difference between
the ice sheet crest (point B) and Tasiilaq (Fig. 2d). Since point B is at approximately 2700 m a.s.l., the surface pressure (p_s)
was reduced to mean sea level using the hypsometric equation:

$$mslp_c = p_s \cdot \exp\left(\frac{g \cdot z_s}{R_d \cdot \bar{T}}\right),$$

260 where z_s is the elevation of the crest, R_d is the specific gas constant for dry air ($287.05 \text{ J (kg K)}^{-1}$), and $\bar{T}$ is the mean
temperature of the fictitious air column, estimated using the T_{2m} and the standard lapse rate ($\Gamma=6.5 \text{ K km}^{-1}$).

3.4.4 Detecting low-level jets and cloudiness

A low-level jet (LLJ) associated with downslope winds is a shallow, terrain-following wind maximum constrained to the lowest
troposphere. The typical LLJ altitude core ranges between 100 and 1000 m a.g.l, though it can elevate up to 2000 m a.g.l or
265 descend to within the lowest tens of meters during highly stratified katabatic flows (Gohm et al., 2008; Heinemann, 2025).
Here, an LLJ is defined when U maxima (nose) stronger than 12 m s^{-1} occur in the lowest 2000 m of the atmosphere with
at least 2 m s^{-1} higher than the wind speed recorded at the surface and with a wind speed lower than at least 20% of the U
maxima above it.

To differentiate between clear-air and cloudy conditions, we assessed whether the RH is higher than 95% and the specific
270 humidity (q) higher than 0.05 g kg^{-1} across vertical levels. q was derived from the saturated vapor pressure using the Magnus
formula (Alduchov and Eskridge, 1996).

4 Results and discussion

The results are based on the piteraq events detected by CARRA that co-exist ($\pm 48 \text{ h}$) with the piteraq events detected by the
AWS in Tasiilaq in part I of this study (Hualand et al., 2026). This means that the studied events typically start earlier (56.2%)
275 and end later (68.8%) than those recorded by the AWS (Fig. A2). Since one of the study goals is to understand the spatial
structure and the associated mechanisms before and during piteraq events in Tasiilaq, we opted to use the CARRA timings



of the piteraq detection to ensure consistency between piteraq detection and atmospheric state. For the period 1991–2024, CARRA identified 91.4% of AWS-detected events, and the two detections yield closely comparable seasonal distributions (Table 1).

Table 1. Seasonal distribution (percentage) of simultaneous piteraq events identified in the AWS and CARRA at Tasiilaq between 1991 and 2024.

	DJF	MAM	JJA	SON
AWS	35.2	25.7	2.9	36.2
CARRA	35.4	28.1	3.1	33.3

280 While piteraq events occur year-round, we will primarily focus on winter occurrences, as they are typically more hazardous (Haualand et al., 2026). The less frequent events observed during summer are addressed in Sect. 4.4.

4.1 Synoptic drivers

To determine how synoptic-scale background conditions precondition piteraq events, we investigated the 15-day evolution of the climate oscillation indices detailed in Sect. 2.2: namely the North Atlantic Oscillation (NAO), the Arctic Oscillation (AO),
285 the Pacific-North American (PNA) pattern, and the Greenland Blocking Index (GBI). For winter cases (DJF), we observed a gradual increase in the mean NAO and AO, alongside a decrease in the GBI (Fig. A3). However, the interdecile range for all indices exhibited similarly high variability throughout the entire 15-day period.

The relationship between piteraq events and climate oscillations is highly complex. As argued by Li (2003), the wintertime NAO and AO are largely inseparable, sharing common storm tracks, with the topographic forcing of the Greenland Ice Sheet
290 playing an important role in characterizing their winter patterns. While intense westerly winds at the southern tip of Greenland are a typical characteristic of positive NAO phases, Jung and Rhines (2007) noted that the correlation between the NAO and zonal pressure drag is relatively weak and highly episodic. Furthermore, Oltmanns (2015) demonstrated that while severe downslope winds in southeast Greenland are influenced by the NAO, their occurrence actually correlates strongly ($r=-0.7$) with regional blocking patterns. This large spread suggests that while a general tendency of increasing NAO and AO exists, these
295 climate oscillations are not robust standalone predictors for piteraq events.

In the following, we examined the immediate synoptic-scale configuration and evolution prior to the start of the piteraq events. Therefore, we made seasonal composites for air temperature, wind, and geopotential height at 500 hPa, as well as mean sea-level pressure, for intervals of 48–24 h, 24–12 h, and 12–3 h before the start of the events, as well as during the events (Fig. 3).

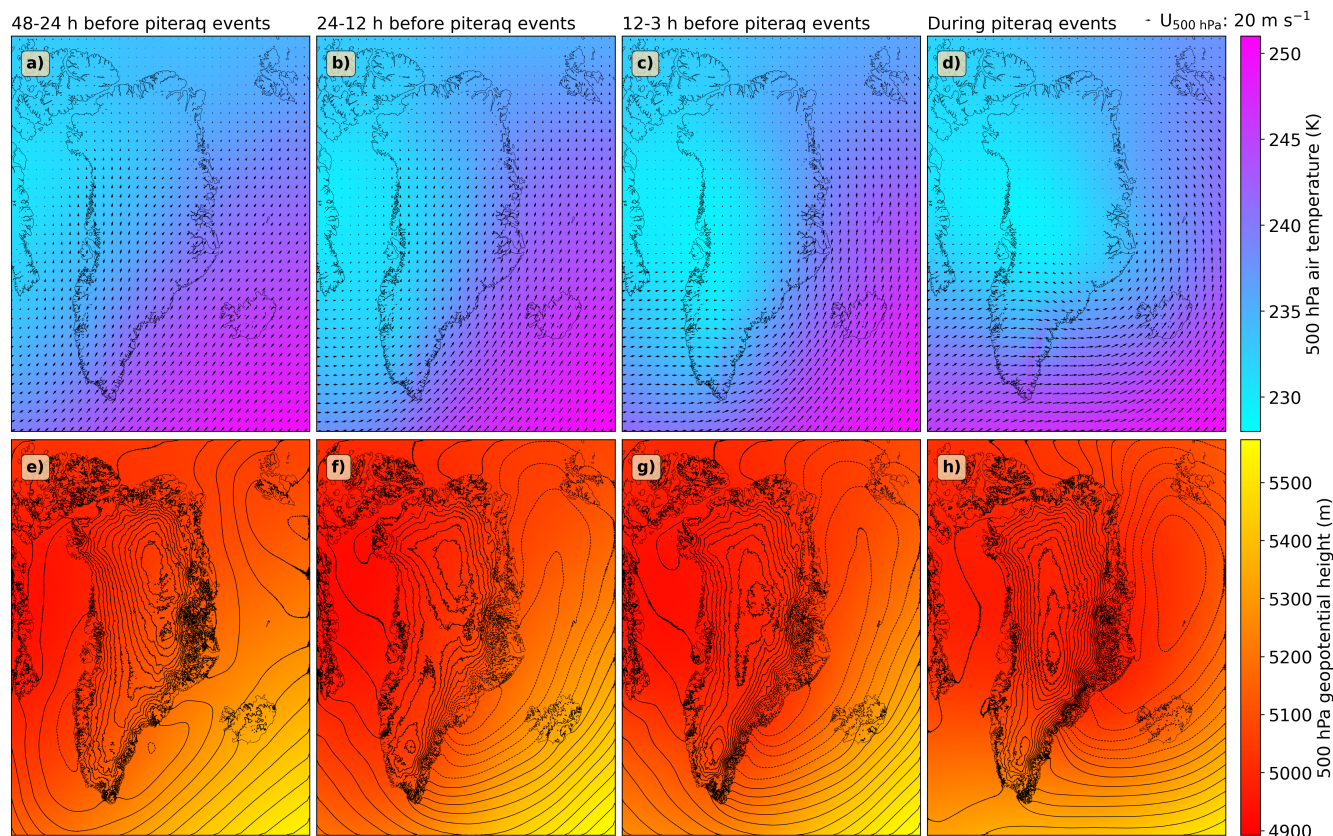


Figure 3. Synoptic evolution of winter (DJF) piteraq events averaged over the 1991–2024 period. (a–d) Top row: 500 hPa air temperature (shading) and wind fields (vectors). (e–h) Bottom row: 500 hPa geopotential height (shading) and mean sea-level pressure (contours, with values below 1000 hPa dashed). Columns represent time windows leading up to and during the event: 48–24 h prior (a, e), 24–12 h prior (b, f), 12–3 h prior (c, g), and during the event (d, h).

Two days before the piteraq onset in Tasiilaq, the composite analysis showed a pronounced southwesterly flow at 500 hPa crossing Greenland (Fig. 3a). Near the surface, the climatological anomaly analysis for the 1991–2024 period revealed an initial positive 2-meter temperature anomaly of approximately 5 K in the interior of Greenland, driven by anomalous southwesterly winds (Fig. A4e). Consequently, the typical background katabatic flow is dampened by two compounding factors: the mechanical opposition of the southwesterly winds, and the thermodynamic erosion of the cold, dense surface layer. Simultaneously, anomalously cold air is positioned over the Canadian Arctic while warm air is located in the North Atlantic. The observed reduction in surface pressure over Greenland suggests that the advancing synoptic system, characterized by strong southwesterly flow, displaces the Greenland block northward, while a relatively strong Atlantic Ridge steers the mesoscale surface low in southeast Greenland toward the ice sheet (Fig. 3e).

As the onset of the piteraq event approaches (24-to-12 h before onset, Fig. 3b), the cold air mass over the Canadian Arctic is advected eastward, particularly over the Davis Strait (Fig. A4f). The deep, cold flow is then blocked by Greenland’s topography,



which increases surface pressure across West Greenland and generates a strong tip jet in South Greenland (Fig. A5b). The impacts of a strong tip jet in conjunction with the downslope windstorm are noticeable over the Irminger Sea, with near gale winds stretching more than 800 km from South Greenland to south of the Tasiilaq region.

Between 12 and 3 hours prior to a piteraq event (Fig. 3c), the blocked cold air eventually overtops the ice-sheet ridge and flows downslope. As the anomalously cold air descends, it creates a shallow and highly stable atmospheric boundary layer (Fig. A4g). The negatively buoyant air drives a drainage flow, converting high amounts of potential energy into kinetic energy. It is crucial to emphasize that the dynamically forced drainage flow should not be confused with classical katabatic winds, which are actually dampened during the earlier preconditioning phase. During the brief window (12–3 h before the event), the cold anomaly progressively expands into the interior of Greenland (Fig. A4h). By the onset of the piteraq, the region experiences its most extreme near-surface temperature, with anomalies reducing by -12 K compared to the climatological mean over the ice sheet, and -5 K over the Irminger Sea near Tasiilaq. Furthermore, the 500 hPa air temperature anomaly clearly identifies the crossing of an anomalously cold vortex, which is the primary driver for the negative near-surface temperature anomalies. Ultimately, the combined evolution of the composite flow and its statistically significant anomalies, specifically the advection of the cold Canadian air mass over an anomalously warm ice sheet, demonstrates that a preexisting cold-air pool over the Greenland Ice Sheet is not necessary to trigger a piteraq event. During the piteraq events, the mesoscale surface low is amplified by the strong, cold, and dry winds in the Irminger Sea. The spatially diffuse low-pressure signal spanning the Denmark Strait to the Fram Strait suggests that mesoscale surface lows are not confined to a fixed location, pointing to considerable case-to-case variability in the precise position and number of formed mesoscale surface lows. Nevertheless, these mesocyclones are ultimately steered towards the northeast by the North Atlantic Ridge.

The observed advection of the Canadian cold vortex aligns well with established dynamic theories regarding upper-level synoptic forcing. The synoptic development of piteraq events is frequently triggered by superposition of an upper-level potential vorticity anomaly, associated with an upper-level trough or the aforementioned cold vortex, onto a stationary topographic Rossby wave over Greenland (Oltmanns, 2015). The upper-level interaction facilitates the rapid deepening of surface cyclones in southeast Greenland through baroclinic instability. As the upper-level disturbance crosses southern Greenland from the west, it induces lee cyclogenesis or intensifies existing low-pressure systems off the east coast (Gutjahr et al., 2022; Oltmanns, 2015; Scorer, 1988). Consequently, the most severe events often involve secondary cyclogenesis in the Denmark Strait (Cappelen and Jensen, 2021; Rasmussen, 1989), which explains the common presence of mesoscale lows between Iceland and southeast Greenland prior to piteraq events (Gutjahr et al., 2022; Klein and Heinemann, 2002; Oltmanns et al., 2014). Consistent with the synoptic composites in Figure 3, Haualand et al. (2026) reported pronounced southwesterly flow in Tasiilaq radiosonde profiles during this phase. Finally, the temporal evolution observed in our analysis strongly aligns with descriptions by Rasmussen (1989), Hasholt et al. (2004), and Cappelen and Jensen (2021), confirming that severe piteraq winds consistently occur on the trailing flank of the mesoscale low as the system is steered northeastward by the North Atlantic upper-level ridge.



4.2 Upstream preconditioning and cross-barrier forcing

Having shown that piteraq events are triggered by the synoptic-scale advection of a Canadian cold vortex, we quantify the atmospheric stability and flow dynamics of the topographically unperturbed air masses in West Greenland during the upstream preconditioning phase (12 to 3 h before piteraq onset). Specifically, we evaluated the frequency of westerlies ($270^\circ \pm 45^\circ$) and the occurrence of supercritical flow across segment E, located approximately 100 km upstream of the Greenland Ice Sheet crest, during the winter months (DJF) from 1991 to 2024 (Figure 4).

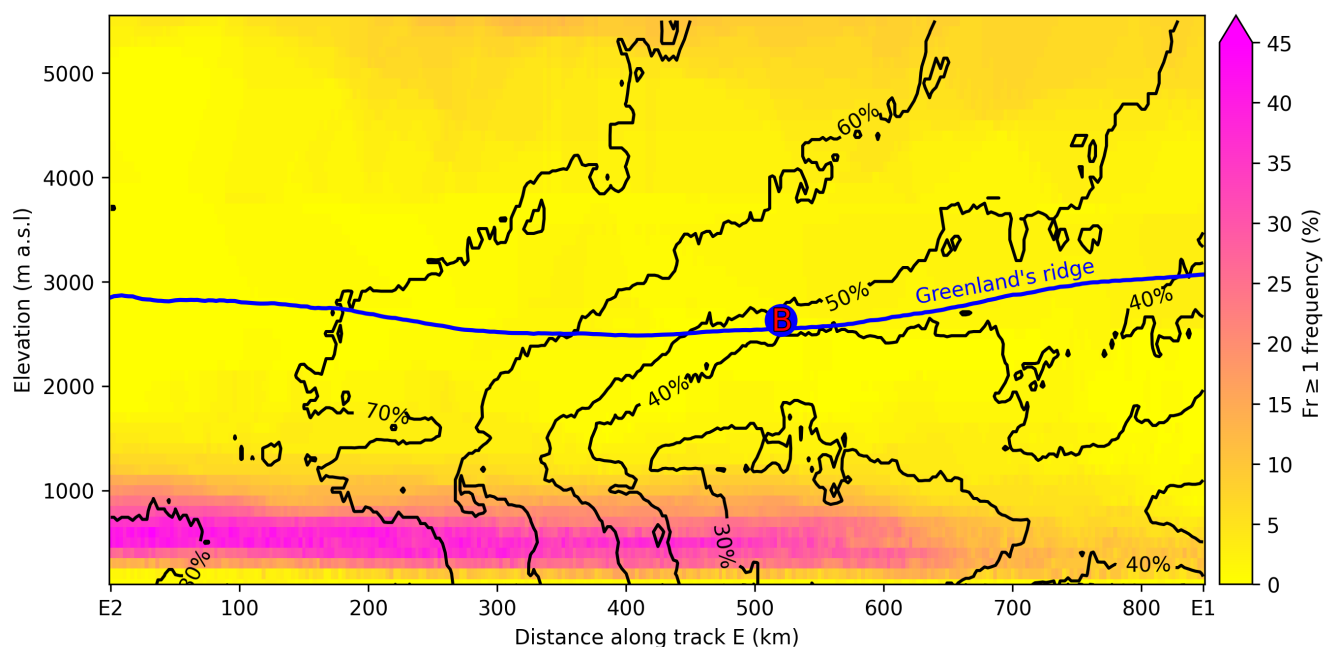


Figure 4. Frequency of westerlies ($270^\circ \pm 45^\circ$, contours) and supercritical flow ($Fr \geq 1$, color-filled contours) through segment E, spanning from south (0 km) to north (820 km) as defined in Figure 2c. Data represent the 12-to-3 h preconditioning phase prior to winter (DJF) piteraq events from 1991 to 2024. The elevation of the Greenland ridge is indicated by the solid blue line. Point B denotes the location of the segment A.

Assessing the flow approximately 100 km upstream across segment E revealed the vertical stratification of the approaching air mass. Near the surface, we observed a low frequency of supercritical flow ($Fr \geq 1$) coinciding with a negative Brunt–Väisälä frequency (N^2). This indicates a turbulent, well-mixed boundary layer where traditional Froude number scaling is not applicable. However, the layer immediately above (200–1000 m a.s.l.) exhibited the highest frequency of supercritical flow (approx. 45%, pink shading in Figure 4), covering the southernmost part of segment E from 0 to 400 km, up to the Maniitsoq Ice Cap located at the 500 km mark. Furthermore, the N^2 composite (not shown) identified the strongest static stable layer at approximately 2000 m a.s.l., suggesting that the approaching supercritical flow is capped by a strong temperature inversion.



While the vertical cross-wind frequency (contours in Figure 4) remains relatively uniform up to 5000 m, it exhibits a strong horizontal latitudinal gradient. Frequencies peak at 70–80% in the southern latitudes of segment E, driven by lower-tropospheric westerlies, but drop to approximately 50% below the crest at higher latitudes (north of the Maniitsoq Ice Cap). This reduction is potentially a result of upstream terrain blocking combined with high N^2 , which likely decouples the upper-level flow from the surface, allowing it to drift over the blocked cold air and the ice sheet.

Overall, our results from upstream conditions demonstrated that the supercritical flow directed toward Greenland frequently hits the terrain as a westerly wind which is forced to lift orographically. As the supercritical flow decelerates against the topography, it piles up, directly contributing to the increased surface pressure on the windward side (as shown in Sect. 4.1) and preconditioning for a severe cross-mountain pressure gradient.

After the piteraq onset, the synoptic forcing intensifies (Fig. A6). The majority of winds above the Greenland crest are westerlies, reaching 80% at around 5000 m a.s.l. The considerable spatial footprint highlights that piteraq events are not localized, topographically isolated features, but rather the result of a synoptic-scale weather event impacting the entire southern half of Greenland. Concurrently, the frequency of supercritical flow in the lower troposphere remains elevated (>40%), while the frequency of westerlies increases with elevation, suggesting that the system is later driven by an upper-level jet.

To further characterize the evolution of the upstream flow regime (12-to-3 h before the onset), we investigate the difference in both cross-barrier potential temperature (θ) and mean sea-level pressure difference (mslp) between the Greenland Ice Sheet crest (point B) and Tasiilaq, alongside the vertical profile of the Scorer parameter (l^2) at the crest (Figure 5).

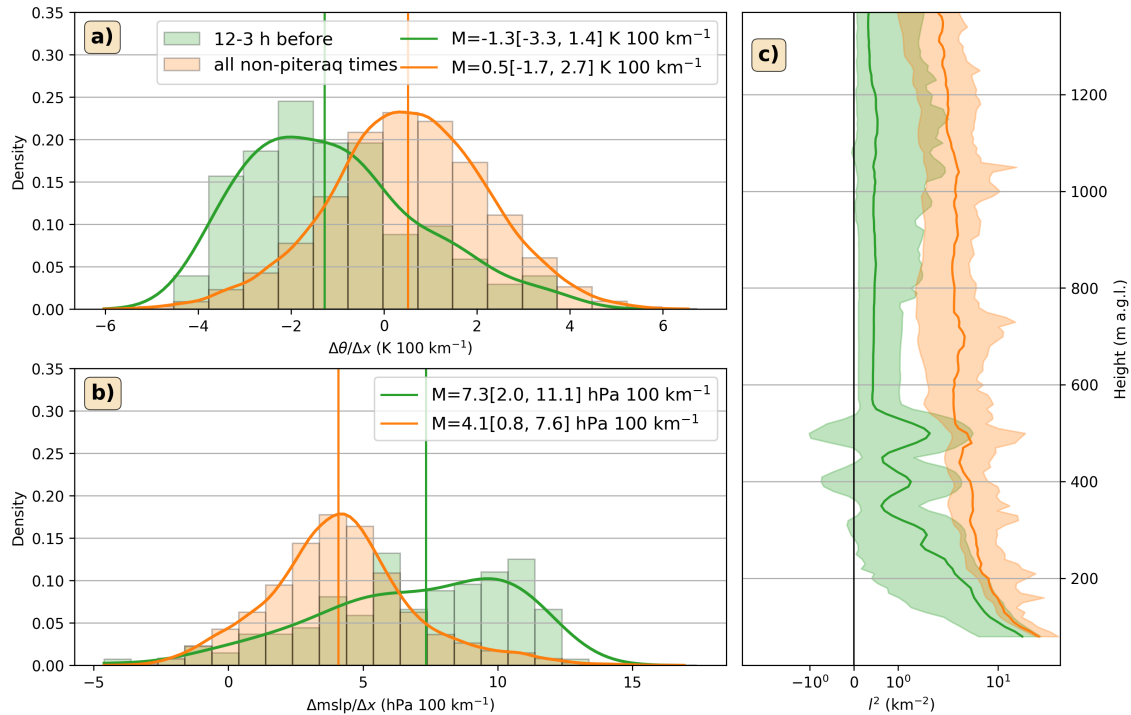


Figure 5. Comparison of atmospheric conditions prior to piteraq events (green; 3–12 h before onset) and the 1991–2024 DJF climatology (orange). Panels (a) and (b) show the difference in potential temperature and mean sea-level pressure, respectively, between the Greenland Ice Sheet crest (point B in Figure 2) and Tasiilaq. The median and interdecile range are shown for both variables. Panel (c) displays the vertical profile of the Scorer parameter (I^2) on a logarithmic scale (x-axis) on the Greenland Ice Sheet crest. Solid lines represent median values, while shaded regions indicate the interdecile range.

Our results showed that the atmospheric conditions at the crest deviate from climatology in the 12-to-3 h window before the piteraq event. Thermodynamically, the crest becomes anomalously cold, with θ dropping to 262.8 K compared to the climatological baseline of 269.5 K (not shown). The cross-barrier potential temperature gradient ($\Delta\theta/\Delta x$) between the ridge and the fjord reverses, shifting from a stable climatological median state of +0.5 K 100 km⁻¹ to a negative value of -1.3 K 100 km⁻¹. This negative $\Delta\theta/\Delta x$ indicates the presence of the dense air mass necessary to drive the gravitational drainage flow.

Simultaneously, the mean sea-level pressure gradient ($\Delta\text{mslp}/\Delta x$) nearly doubles, averaging 7.3 hPa 100 km⁻¹ prior to the event compared to the 4.1 hPa 100 km⁻¹ climatological median. Driven by this enhanced pressure forcing, the near-surface winds accelerate at point B into the northwesterly regime, with a zonal wind of 7.3 m s⁻¹ and a meridional wind of -6.1 m s⁻¹, compared to the corresponding climatological values, 2.2 and 5.3 m s⁻¹ (not shown).

Beyond the thermodynamic and pressure forcing, the vertical profile of the Scorer parameter (I^2) revealed the wave dynamics associated with piteraq events (Fig. 5c). Under typical DJF conditions between 1991 and 2024, the I^2 profile exhibited in both profiles an exponential decrease in the first 200 m. Particularly for the DJF climatology (orange profile in Figure 5c),



385 the near-surface structure suggests stable stratification, while the nearly constant values extending into the mid-troposphere allow mountain waves to propagate vertically. In contrast, the before onset conditions (green profile in Figure 5c) show a different vertical structure as the profile adopts a more complex shape, dropping to approximately one order of magnitude. Most notably, the 10th percentile of l^2 occasionally becomes negative in the lowest 600 m a.g.l. It should be noted that some of the l^2 irregularity, including the negative extremes, may be a consequence of sensitivity to the smoothing window applied
390 to the vertical wind profile, combined with interpolation between the model levels. Nevertheless, the overall reduction in l^2 suggests that intense wind shear likely overcomes the atmospheric stability. This pronounced vertical reduction in l^2 creates an environment highly susceptible to wave trapping and resonance.

Strong cross-barrier mean sea-level pressure gradients are universally recognized as the primary dynamical driver of severe downslope windstorms, including the Alpine foehn, the Rocky Mountains Chinook and the Adriatic Bora (Smith, 1989; Durran,
395 1986; Grisogono and Belušić, 2009). Our results suggest that this forcing for piteraq events is primarily established by the Canadian cold vortex approaching West Greenland. A typical cross-barrier pressure gradient for the Alpine foehn and Chinook ranges from 2 to 8 hPa 100 km⁻¹, reaching up to 10 hPa 100 km⁻¹ only in the most severe foehn and Bora cases (Hoinka, 1985; Horvath et al., 2009). Notably, the precursor conditions (interdecile range) of a DJF piteraq already align with this scale. More strikingly, the median $\Delta\text{mslp}/\Delta x$ during the DJF piteraq events is 9.2 [6.4, 12.3] hPa 100 km⁻¹, an intensity
400 aligned with that of severe Alpine foehn and Bora episodes (Fig. A7). It should be noted that while Alpine pressure gradients are traditionally evaluated across the entire mountain range, our piteraq gradient is calculated specifically over the leeward descent. Even with this spatial distinction, this comparison highlights the severe nature of the localized pressure forcing acting across the Greenland topography. As observed in our crest profiles, the rapid decrease in l^2 with height facilitates the wave energy trapping known to strongly amplify such downslope flows (Tollinger et al., 2019; Metz and Durran, 2023; Durran,
405 1990).

4.3 The leeward response

Following the initial descent from the crest, the local terrain introduces a secondary, critical forcing mechanism. The broad downslope topography forms a wide, flat U-shape, where the northern flank of the terrain abruptly rises towards the Schweizerland range (highest region in Figure 2b), which elevates above 3300 m a.s.l.. Consequently, the flow is topographically blocked
410 at its northern flank and is deflected. This unilateral blocking forces the cold drainage flow to laterally converge and accelerate towards the southeast, channeling the momentum along the unobstructed downslope path (Fig. 2a). Ultimately, this specific topographic deflection dictates the common trajectory of the piteraq events and spatially concentrates its kinetic energy just before it reaches the steepest slopes.

4.3.1 Characteristics of the low-level jets

415 To characterize the lee-side atmospheric response during piteraq events, we evaluate the frequency, intensity, and depth of low-level jets (LLJs) propagating from the flat interior ice sheet down to the rugged coastline. Because LLJs represent the concentrated maxima of kinetic energy generated by the downslope acceleration, identifying their structure is relevant to



understanding how the piteraq manifests at the surface. Therefore, we detected and analyzed these jet cores (LLJ nose) along segment A and across segments B, C, and D. Figure 6 shows the resulting spatial distribution of wind speeds at the LLJ core
420 during the winter months (DJF) from 1991 to 2024.

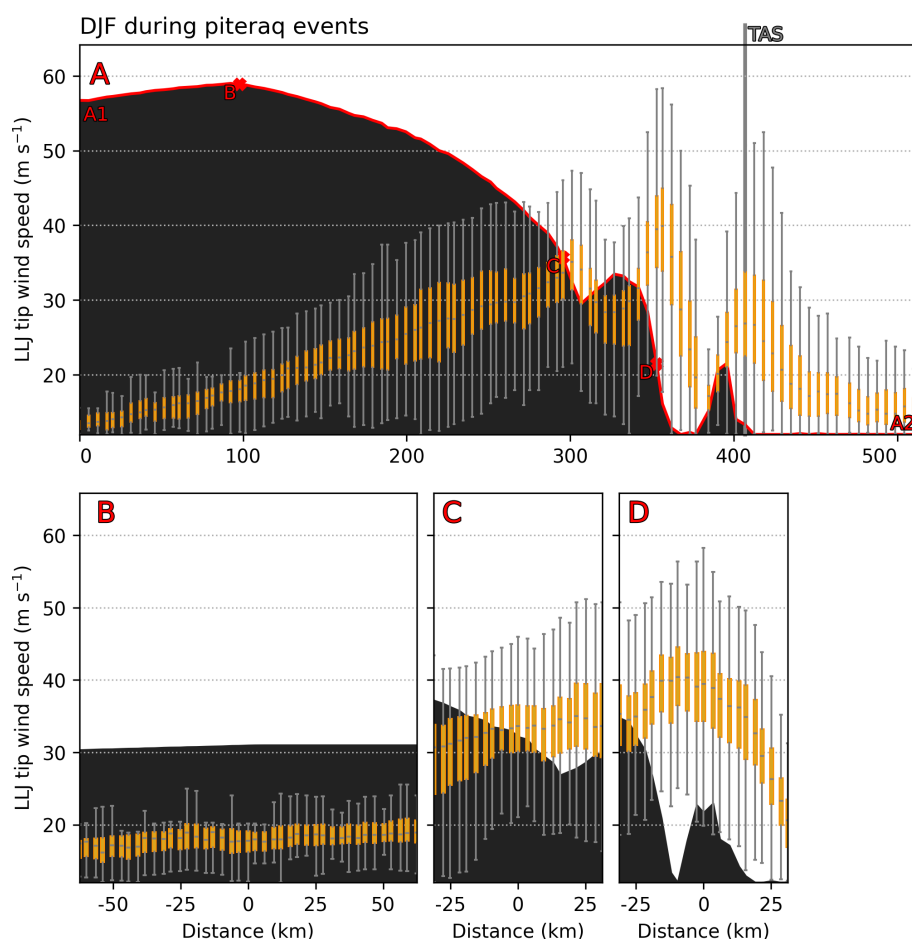


Figure 6. Wind speed at the tip of the low-level jet (LLJ) during the piteraq events for DJF between 1991 and 2024 along the segment A and across segments B, C, and D. The topography of the segments is normalized and displayed in the background. The position of Tasiilaq is marked with a gray vertical line along segment A. The boxplot whiskers extend from the lower inner ($Q_1 - (1.5 \times IQR)$) to the upper inner ($Q_3 + (1.5 \times IQR)$) fences. The gray horizontal line in the center of the boxplot corresponds to the median of the wind speed at the tip of the LLJ.

Near the crest of the Greenland Ice Sheet (Figure 6, segment B), the LLJs exhibited their lowest frequency and intensity, maintaining a consistent magnitude across a 100 km stretch. Progressing downslope, the frequency and mean intensity of the jets showed a strong positive correlation ($r = 0.7$). Moving into segment C, the LLJs enhance further, with the strongest cores concentrated on the northeast flank (0–25 km) due to topographic deflection from the Schweizerland range. Thermodynami-



cally, these piteraq LLJs are frequently coupled with near-surface temperature inversions (TIs), which are present over the ice sheet in approximately 70% of cases (partly seen in Figure 7). Traversing the topographic depression of the Helheimgletsjer tributary (at 300 km in Fig. 6A), the flow is forced to ascend before descending the steepest leeward slope. The adiabatic warming associated with the descent enhances the TI frequency to approximately 75% (not shown) and drives the LLJs to their maximum intensity, increasing median core speeds beyond 40 m s^{-1} . This rapid acceleration coincides with the flow funneling into the narrow, flow-aligned Qeertartivattaap Kangertiva fjord (at 350 km in Fig. 6A and -10 km in Fig. 6D). Similar to the topographically driven gap jets characteristic of Adriatic Bora events (Grisogono and Belušić, 2009; Kozmar and Grisogono, 2020), laterally blocked upstream air accelerates to conserve mass, exiting the gap as a highly concentrated momentum jet.

Over the ice sheet and coastal fjords, the altitude of the piteraq LLJ core (median $\sim 300 \text{ m a.g.l.}$) aligns well with established observations (partly seen in Figures 7C and D), which typically position the jet within the lowest few hundred meters of a highly stratified boundary layer (e.g., Jiang and Doyle, 2005). However, a kinematic and thermodynamic shift occurs once the flow reaches the broader Sermilik Fjord (Figure 6A at 360 km). Here, the LLJ undergoes a sudden deceleration: the median core intensity drops, its elevation increases, and the TI frequency plummets to just 10% (not shown).

Subsequently, the flow is orographically lifted as it reaches Ammassalik Island. On the leeward side of the island (within the merged, flow-oriented valleys leading to Tasilaq), the LLJs re-intensify, recording the highest overall frequency (68%, Fig. A8) and median core wind speeds of 30 m s^{-1} . Concurrently, the TI frequency partially recovers ($\sim 35\%$). This TI increase occurs as the preexisting valley cold pool is strengthened by the descending flow undergoing adiabatic warming aloft. The topographically forced subsidence enhances the negative buoyancy of the descending air mass, which, when channeled through the merged Ammassalik Island valley system, contributes directly to the secondary downslope acceleration of the LLJ. Finally, as this relatively cold, dry flow propagates over the warmer Irminger Sea, the jet core rises to above 500 m a.s.l. and exhibits a spatial oscillation suggestive of trapped lee waves. Large air-sea temperature gradients drive strong turbulent heat fluxes, promoting strong boundary layer convection that erodes the low-level stratification (reducing TI frequency to near zero). This rapid convective destabilization is a classic signature of maritime cold air outbreaks (Gutjahr et al., 2022).

4.3.2 Thermodynamic evolution and wave breaking

To explicitly visualize the evolution of the downslope windstorm, Figure 7 presents a composite of DJF piteraq events (1991–2024). The composite displayed isentropic contours, along-segment flow vectors, and cross-wind speeds across segments A through D. Due to the perpendicular orientations, it is important to note that the along-segment wind components in segment A project as cross-segment winds in segments B, C, and D. Over the Greenland Ice Sheet crest, the cross-wind component is relatively weak (Fig. 7 A and B). Across the initial 100 km descent, the composite revealed a gentle downslope flow that gradually strengthens, dominating the localized wind field while cross-winds remain weak ($< 10 \text{ m s}^{-1}$).

As the flow reaches C and D, the strongest downslope winds consolidate within the lowest 1000 m of the atmospheric column, becoming shallower and more intense toward the coast (Fig. 7 A, C and D). Along segment A, the isentropes showed strong vertical gradients in the regions where the LLJs are most intense (at 300 and 350 km). This high static stability suppresses vertical turbulent mixing, effectively decoupling the near-surface flow from the layers above. Facilitated by this decoupling



and the wave-energy trapping identified in our l^2 profiles, the flow undergoes severe downslope acceleration as it crosses the sudden terrain transition from the relatively smooth ice sheet to the complex coastal topography (Fig. 7 A and D). However, upon reaching the steepest slope immediately upstream of Qeertartivattaap Kangertiva (point D), the descending isentropes sharply ascend, forming a steep wave structure with phase lines that tilt upstream against the flow.

Directly following this wave structure, the vertical gradient of potential temperature weakens over the Sermilik Fjord. The vertical gradient weakening is corroborated by the wind speed in Figure 7D: while the near-surface flow (below 1000 m) continues downslope through the Qeertartivattaap Kangertiva, the vertical wind component aloft (above 1000 m) ascends across the 60-km transect. The localized gap-flow termination and the subsequent wave breaking together explain the concurrent collapse of static stability and jet momentum observed directly over the Sermilik Fjord, as the jet's kinetic energy is dissipated into turbulence and upward vertical motion. Downwind of this wave-breaking region, the lowest isentropes (up to 1500 m a.s.l. in Figure 7A) compress once again and closely follow the local topography on the leeward side of Ammassalik Island. This confirms the previously hypothesized reestablishment of the secondary downslope flow.

Aloft, the strongest cross-wind is observed as a southwesterly flow above 4000 m a.s.l. over the Irminger Sea, reflecting the broader synoptic-scale forcing. Finally, as the near-surface flow propagates over the open Irminger Sea, the isentropes transition to a vertical orientation (at 500 km in Figure 7A). This vertical isentropic structure thermodynamically confirms the presence of the strong heat fluxes and convection characteristic of the maritime cold air outbreak discussed previously.

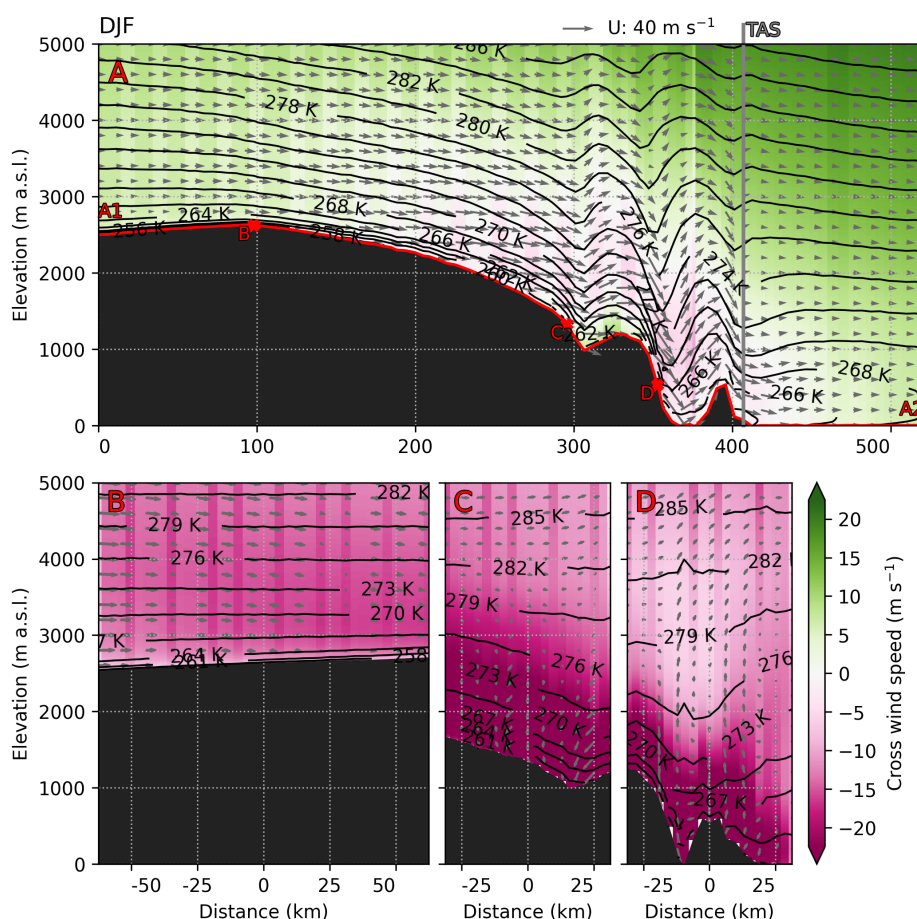


Figure 7. Thermodynamic evolution during DJF piteraq events along segment A and across segments B, C, and D. Contours indicate isentropic lines, while vectors represent flow along segments. Colors denote cross wind speed: purple indicates flow out of the page, and green indicates flow into the page. The position of Tasiilaq is marked with a gray vertical line along segment A.

475 While the isentropic composites suggest broad structural features, such as wave breaking, seasonal composites inherently smooth high-amplitude localized extremes. To rigorously quantify the exact locations of severe shear and overturning flow, we assessed the possibility of the critical Richardson number ($Ri_g \leq 0.25$) and reversed zonal flow ($u < 0 \text{ m s}^{-1}$) for each DJF piteraq event (Fig. 8).

Upstream of the crest, our results revealed that localized, shallow flow blocking occasionally (66%) occurs on the windward
480 side of the Greenland Ice Sheet. This blocking contributes to a strong pressure drag, forcing the stably stratified air to dam up against the topography and laterally split around the obstacle (Hoinka, 1985; Harden and Renfrew, 2012). This lateral deflection physically explains the highly frequent formation of southern tip jets observed prior to and during piteraq events (Fig. 3 and A5).



On the lee-side, while the lowest few hundred meters are persistently turbulent due to LLJ shear (orange shading), elevated
485 wave breaking is highly localized to three distinct geographic regions (blue contours in Fig. 8A). The first is located just
upstream of point C within the ascending section of the Helheimgletsjer tributary depression (at 300 km in Figure 8), where
overturning flow is observed in at least 33% of the piteraq events in DJF. The second and most frequent region coincides
with the steepest leeward slope upstream of the Sermilik Fjord. Here, overturning flow occurs in a majority of piteraq events,
accompanied by a high frequency (>70%) of turbulent flow extending up to 3000 m a.s.l. (Fig. 8A and D). This deep vertical
490 column of turbulence dynamically suggests the common location of the wave breaking deduced from the isentropes. Finally, a
third region of frequent overturning flow is situated on the leeward side of Ammassalik Island, localized zone of wave breaking
as the accelerating flow interacts with the island topography.

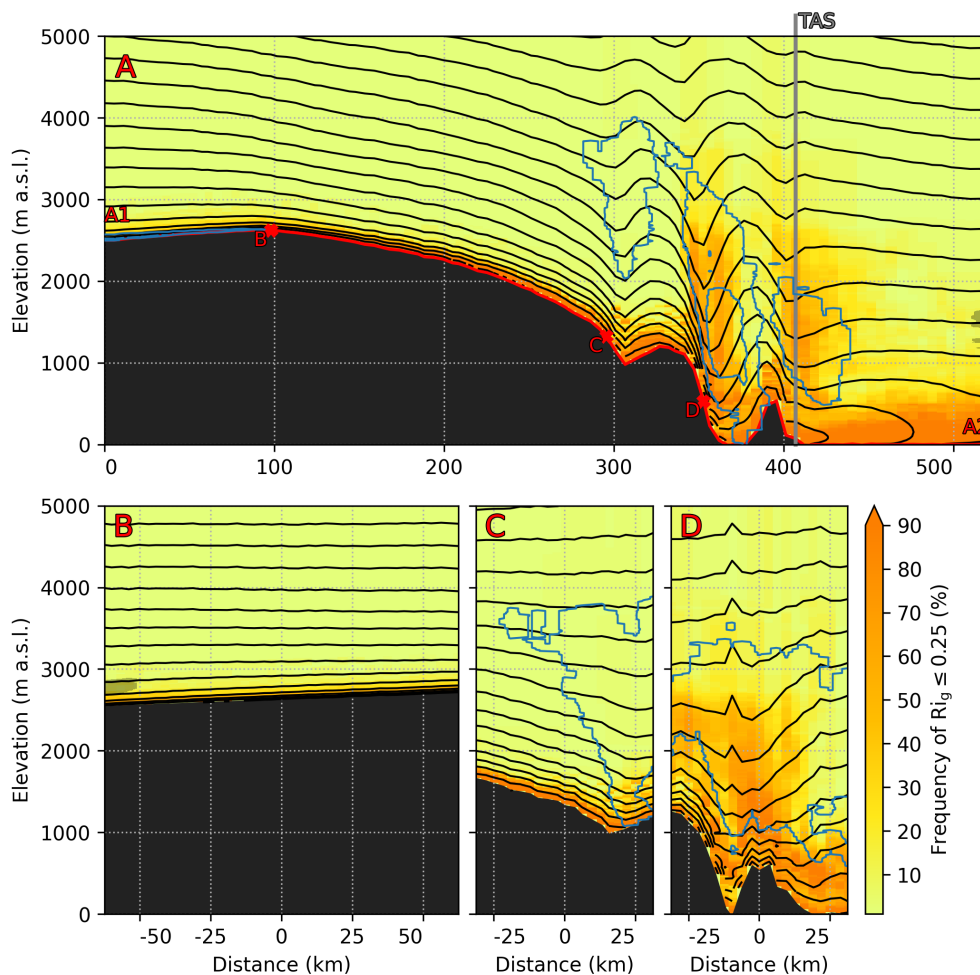


Figure 8. Frequency of $Ri_g \leq 0.25$ (in the colormap), regions with frequency of cloudiness for 33 and 66% (in light and dark gray shading), and frequency of overturning zonal flow for 33, 66 and 99% ($u < 0 \text{ m s}^{-1}$), in blue contour lines) and composite of the potential temperature as in Figure 7 (in black contour lines spaced every 20 K) during piteraq events in DJF along segment A and across segment B, C, and D for the period 1991-2024. The vertical line in segment A marks the location of Tasiilaq.

The strong vertical wind shear between the LLJ and the overlying wave-breaking region frequently reduces the gradient Richardson number (Ri_g) below the critical threshold, leading to Kelvin–Helmholtz instabilities (Grisogono and Belušić, 2009; Tollinger et al., 2019). Furthermore, the specific zones of critical shear and overturning flow identified along the leeward slopes represent regions where wave resonance breaks down into mechanically driven turbulence.

Although our spatial resolution does not explicitly resolve individual microscale wind gusts, high-resolution simulations by Tollinger et al. (2019) demonstrate that such zones of flow reversal are highly conducive to the development of rotor circulations. Consequently, we hypothesize that the localized overturning flow observed in our analysis provides the necessary



500 conditions for rotors, which are potentially responsible for driving the extreme localized surface gusts experienced during piteraq events.

4.4 The role of seasonality

Both winter and extended summer (JJAS) piteraq events are characterized by a distinct synoptic forcing: upper-level winds that warm the air over the interior of the ice sheet, followed by the crossing of a cold Canadian vortex. However, the extended
505 summer revealed a more complex setting, partially influenced by the smaller sample size (Fig. A11).

During JJAS, westerly flow at 500 hPa becomes more pronounced, advecting a relatively cold air mass from higher latitudes north of the Maniitsoq Ice Cap in West Greenland (Fig. A9). Upstream of the crest (segment E), the frequency of westerlies is notably higher in summer (>80%) than in winter, particularly at elevations above the ice sheet crest. The maximum frequency of supercritical flow is centered at approximately 500 m a.s.l., with the highest occurrences localized north of the Maniitsoq Ice
510 Cap. Furthermore, the synoptic timing shifts: while the Greenland tip jet typically precedes winter events by 24–12 hours, it only occurs 12–3 hours prior to summer piteraq events (Fig. A5). However, the primary seasonal distinction is thermodynamic. In contrast to the large pressure and thermal gradients of winter, the summer pressure gradient between the crest and Tasiilaq drops to 3.1 hPa 100 km⁻¹ (on median, compared to 7.3 hPa 100 km⁻¹ in winter), with a distribution that largely overlaps the summer climatology (Fig. A12). This implies that other synoptic conditions can generate similar pressure gradients in JJAS.
515 More importantly, the cross-barrier potential temperature gradient reverses to a positive value (+0.7 K 100 km⁻¹). Unlike the winter drainage flow, this positive gradient indicates that the air mass arriving in Tasiilaq leads to local warming. Although the Scorer parameter profile maintains a similar shape to the winter events, the overall values are lower.

Detected cloud frequencies in JJAS piteraq events provide direct physical evidence for this seasonal dichotomy (Fig. 9). During winter piteraq events, our results showed negligible cloudiness over the Greenland Ice Sheet, consistent with the dry
520 and dense descent air. In contrast, summer events exhibit a high frequency of cloud cover over the ice sheet (at least 66%). Because windward flow blocking is rare in summer, the relatively warmer temperatures allow for the ascent of moister air masses, frequently reaching near-saturation above the crest. This cloudiness is predominantly concentrated within the lowest 1000 m above the crest and establishes itself 24–12 hours prior to the onset of the piteraq event, concurrent with the synoptic advection of the warmer and moister air mass from low elevations. The cloud formation at the crest level releases latent heat,
525 contributing to diabatic warming as the air mass passes over the crest. During the leeward descent, the air mass subsequently warms via adiabatic compression. This dual-heating mechanism, combined with summer warming, drives the distinct near-surface temperature increase observed in Tasiilaq during summer events.

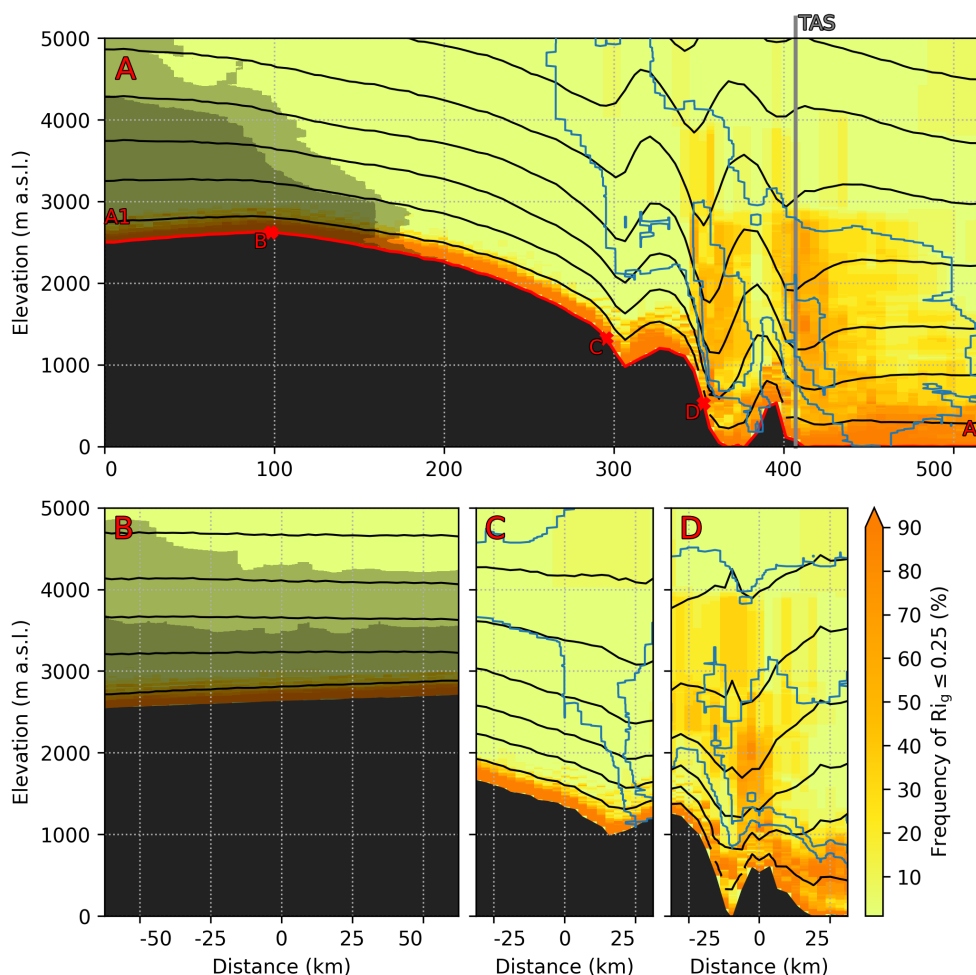


Figure 9. Frequency of $Ri_g \leq 0.25$ (in the colormap), regions with frequency of cloudiness for 33 and 66% (in light and dark gray shading), and frequency of overturning zonal flow for 33, 66 and 99% ($u < 0 \text{ m s}^{-1}$), in blue contour lines) and composite of the potential temperature as in Figure 7 (in black contour lines spaced every 20 K) during piteraq events in JJAS along segment A and across segment B, C, and D for the period 1991–2024. The vertical line in segment A marks the location of Tasiilaq.

LLJs are less frequent during summer events and initially fail to meet jet criteria near the crest (Fig. A10). However, as the flow descends, it still amplifies at the same three topographically forced maxima identified in the winter analysis (Fig. 9A).
 530 Notably, the frequency of leeward turbulence and overturning flow is actually higher in summer (Fig. 9). The leeward shear and wave breaking is likely the mechanical element required for the descending flow to erode the stably stratified, cold-air pool over the Sermilik Fjord and Ammassalik Island.

Ultimately, evaluated through the framework of hydraulic theory, these contrasting thermodynamic, kinematic, and cloud structures reveal a seasonal dichotomy: winter piteraq events are generally characterized as dense, negatively buoyant Bora-like



535 downslope windstorms, whereas summer events commonly manifest as warm, diabatically enhanced foehn-like windstorms in the region around Tasiilaq.

While our findings demonstrated that the vast majority of the analyzed piteraq events are synoptically driven downslope windstorms, we cannot entirely discard the possibility that a minority of the detected events are radiation-driven katabatic storms. If present, katabatic events would most likely occur in autumn and early winter, driven by the rapid radiative cool-
540 ing of the ice sheet interior as the coreless winter is eventually established. Distinguishing synoptically driven events from radiation-driven downslope windstorms is essential because relying solely on single-point in situ surface observations presents an inherent limitation. Because radiation-driven katabatic windstorms likely descend along the same topographical pathways as dynamically forced piteraq events, their localized surface wind signatures in Tasiilaq can appear nearly identical.

5 Conclusions and outlook

545 Our study examines the multi-dimensional structure of piteraq events detected with observations by Haualand et al. (2026) and with high-resolution reanalysis data in southeast Greenland over the period 1991–2024. The multi-scale analysis demonstrates that piteraq events are dynamically forced, topographically channeled downslope windstorms driven by upstream synoptic forcing. The primary synoptic preconditioning relies on the presence and advection of a Canadian cold vortex, which establishes an exceptionally strong cross-barrier surface pressure gradient (with a median of $7.3 \text{ hPa } 100 \text{ km}^{-1}$) and a reduction in the
550 Scorer parameter profile that supports wave trapping and resonance.

As the stably stratified flow crosses the Greenland Ice Sheet, it converts potential energy into kinetic energy, gaining momentum and speed as it descends the steepest leeward slopes. The flow accelerates into a highly supercritical gap jet that occasionally terminates in wave breaking just before encountering the Qeertartivattaap Kangertiva. These localized wave-breaking regions are characterized by intense vertical shear and overturning flow. Furthermore, our spatial thermodynamic
555 analysis reveals a seasonal dichotomy in how piteraq events impact the regional boundary layer. Winter piteraq events are characterized as dense, Bora-like windstorms that erode low-level stratification and drive massive maritime cold-air outbreaks over the Irminger Sea. Conversely, summer events manifest as moist, diabatically enhanced foehn-like windstorms that contribute to localized warming, drying, and an increased potential for snow and ice melt.

The value of the multi-scale spatial analysis presented here overcomes the limitations of single-point event detection. By
560 identifying the distinct synoptic and upstream preconditioning required for dynamically driven piteraq events, specifically the advection of the Canadian cold vortex, windward flow blocking, and Scorer parameter layering, we provide a physical framework that extends beyond local surface metrics. Ultimately, these multi-dimensional atmospheric precursors offer a structured diagnostic footprint to explicitly separate radiatively driven katabatic flows from synoptically driven downslope windstorms.

The diagnostic footprint here presented is particularly valuable given the evolving architecture of atmospheric models. While
565 the physical mechanisms resolved in this study benefit from a non-hydrostatic framework capable of capturing severe vertical accelerations and wave structures, future large-scale reanalysis products may leverage computationally efficient hydrostatic approximations. Because hydrostatic models inherently suppress the vertical momentum transport and mountain wave break-



ing critical to downslope windstorms, they will likely struggle to resolve localized surface accelerations. However, they remain highly capable of simulating the broader synoptic forcing and upstream stratification. By incorporating these resolvable up-
stream variables into advanced machine learning algorithms, future predictive models can statistically bridge the gap between
synoptic preconditioning and micro-scale surface hazards. Leveraging these spatial precursors to bypass localized model lim-
itations will be crucial for improving the lead time, accuracy, and reliability of regional early warning systems for coastal
Greenland communities and maritime traffic.

During winter piteraq events, the topographically channeled low-level jets (LLJs) trigger intense maritime cold-air outbreaks
(MCAOs) over the relatively warm ocean surface. As demonstrated by the vertical isentropic structures, the elevation of the
LLJ, and the widespread regions of turbulence ($Ri_g \leq 0.25$) over the sea, the rapid convective destabilization of the marine
boundary layer indicates the presence of strong upward sensible and latent heat fluxes. By systematically underestimating the
localized intensity of these LLJs and the resulting surface heat extraction, current Global Climate Models (GCMs) likely under-
estimate the buoyancy loss and subsequent deep convective mixing in the Irminger Sea. Furthermore, the seasonal dichotomy
identified in our analysis complicates the ocean-atmosphere feedback. While winter Bora-like events drive ocean densification
and deep convection, the warm, diabatically enhanced foehn-like events observed during summer contribute to higher temper-
atures and likely enhanced melting of the Greenland Ice Sheet. This introduces freshwater fluxes into the coastal fjords and the
Irminger Sea, which is known to increase upper-ocean stratification and actively suppress deep convection.

Determining how the seasonal ratio between these two physically distinct windstorm regimes might evolve under a warming
climate, where regional sea ice reduction and higher atmospheric moisture capacities could alter the frequency of diabatic foehn
warming, remains a critical open question for projecting the net buoyancy forcing acting upon the AMOC. While dynamically
resolving the necessary sub-kilometer topographical funneling and wave breaking remains computationally challenging for
multi-decadal GCMs, our multi-scale analysis offers an alternative approach. Future climate projections can leverage these
resolvable upstream precursors through parameterization schemes to better diagnose sub-grid piteraq occurrences. By dynami-
cally parameterizing the missing momentum and turbulent heat fluxes into the ocean model component when the characteristic
large-scale atmospheric conditions are met, the negative bias in Irminger Sea buoyancy forcing can potentially be reduced,
improving projections of AMOC stability. Furthermore, while the 2.5 km reanalysis provides a robust mesoscale framework,
confirming the hypothesized micro-scale wave breaking and rotor circulations will ultimately require future sub-kilometer
modeling and dedicated observational field campaigns. Ultimately, by redefining piteraq events as complex, multi-scale atmo-
spheric phenomena governed by distinct synoptic and thermodynamic footprints beyond their role as localized surface hazards,
our study establishes a comprehensive physical approach, which provides the foundation for advancing both operational early
warning systems and our broader understanding of extreme atmosphere–ice–ocean interactions in a changing Arctic.

Data availability. Data from weather stations are publicly available from Jensen (2025a) and How et al. (2022). CARRA data (Schyberg
et al., 2020) are publicly available from the Copernicus Climate Change Service (2024). Neither the European Commission nor ECMWF is
responsible for any use that may be made of the Copernicus information or data it contains.



Appendix A: Supporting figures

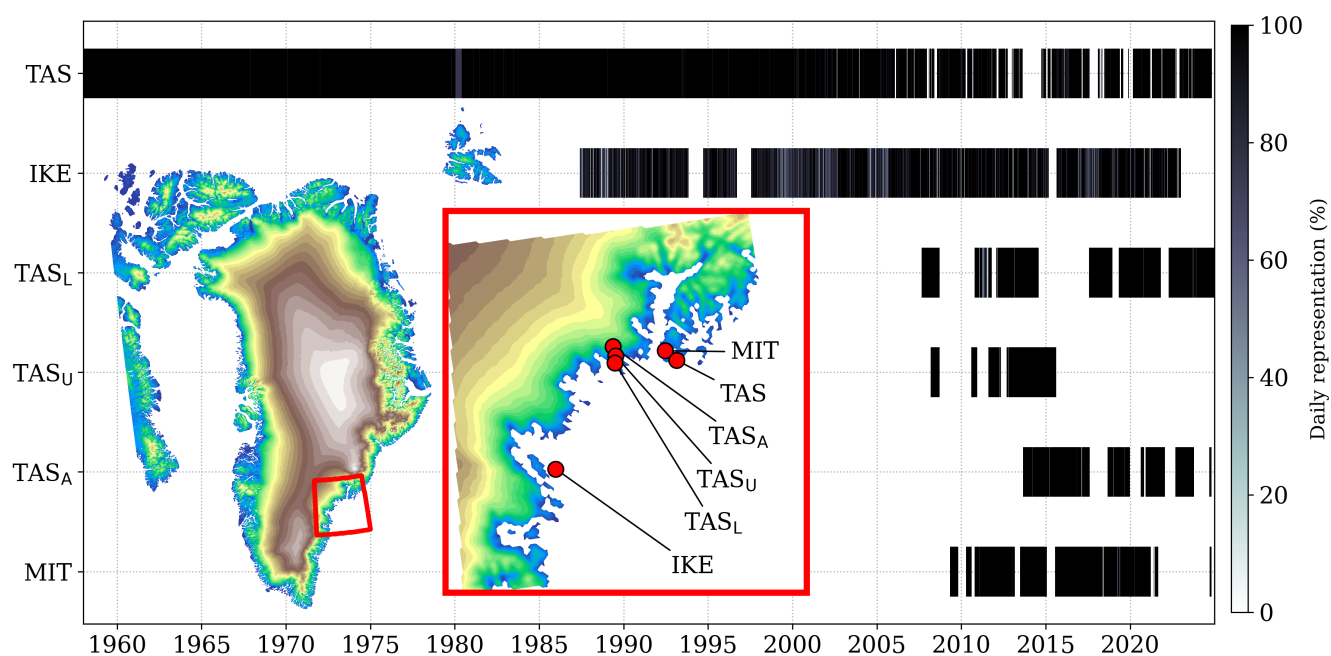


Figure A1. Three-hourly wind observations in the Tasiilaq region (1958–2024). Abbreviations TAS, IKE, MIT, TAS_A, TAS_U, and TAS_L refer to official automatic weather stations (AWS) operated by DMI and PROMICE/GC-NET. The red square on the Greenland map indicates the study area, with station locations shown as red dots in the zoomed inset.

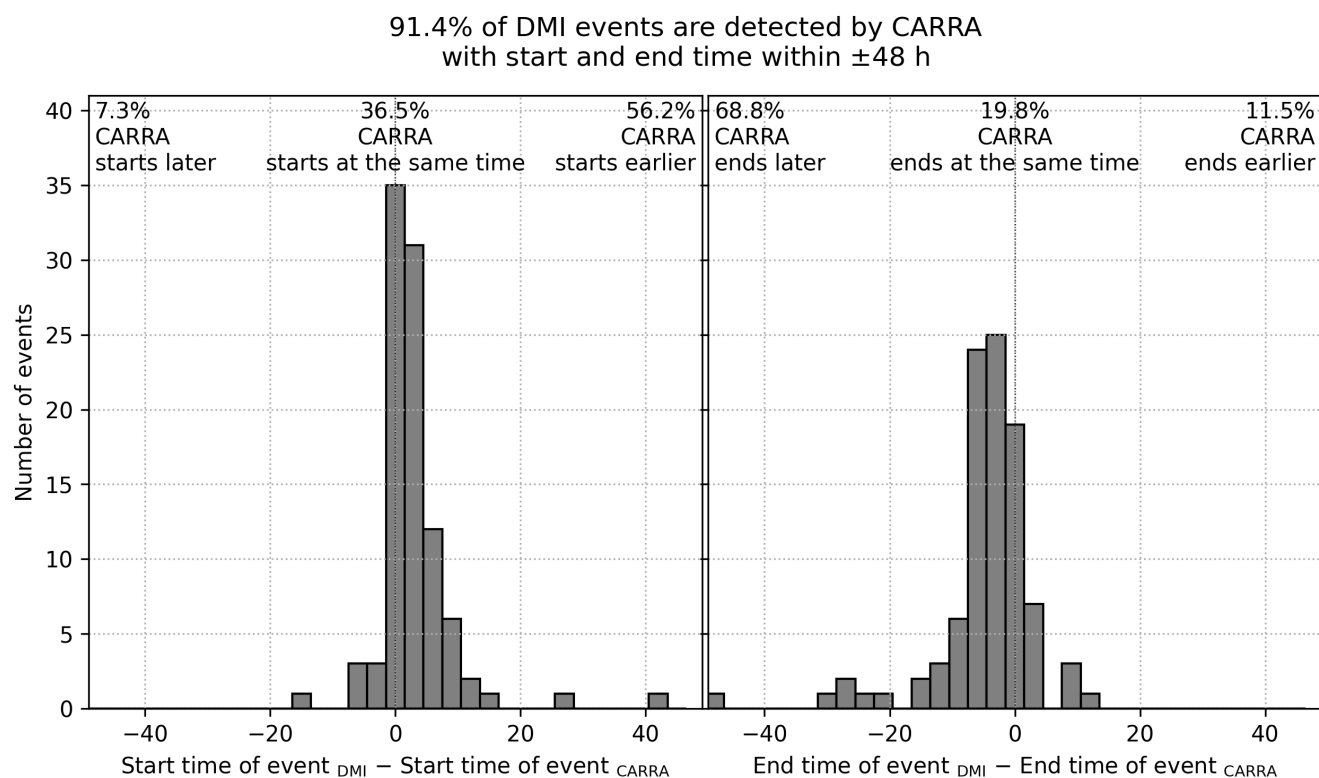


Figure A2. Temporal differences in piteraq event identification between the DMI and CARRA lists. The left panel displays start-time offsets, while the right panel shows end-time offsets. The relative proportions of CARRA events occurring earlier, later, or concurrently with DMI events are indicated for each case.

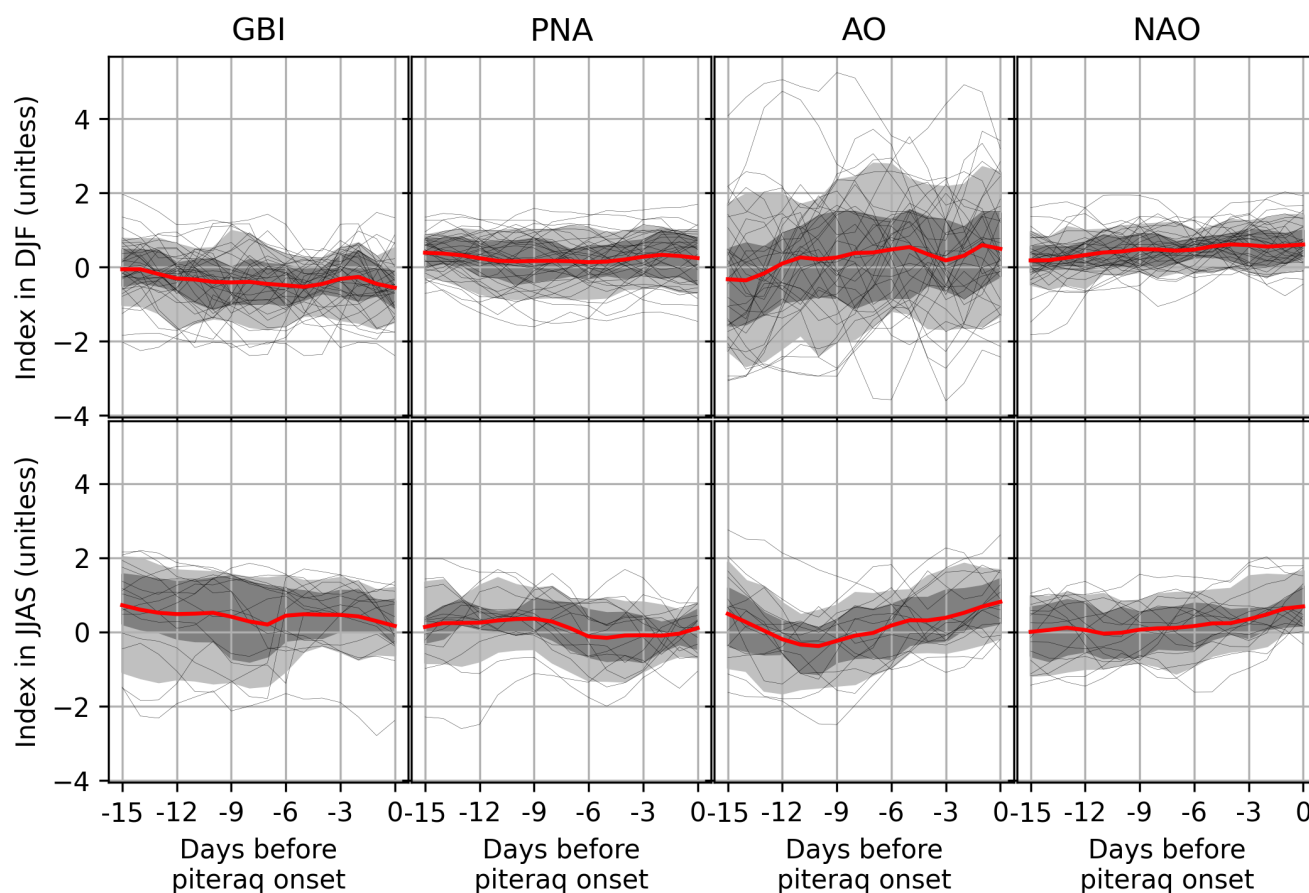


Figure A3. Evolution of climate oscillation indices during the 15 days prior to piteraq events between 1991 and 2024 for winter (DJF) and extended summer (JJAS). Columns from left to right represent the Greenland Blocking Index (GBI), Pacific/North America pattern (PNA), Arctic Oscillation (AO), and North Atlantic Oscillation (NAO). The bold red line denotes the mean evolution, the gray shading indicates the 10th–90th percentile range, and thin lines represent all the piteraq events.

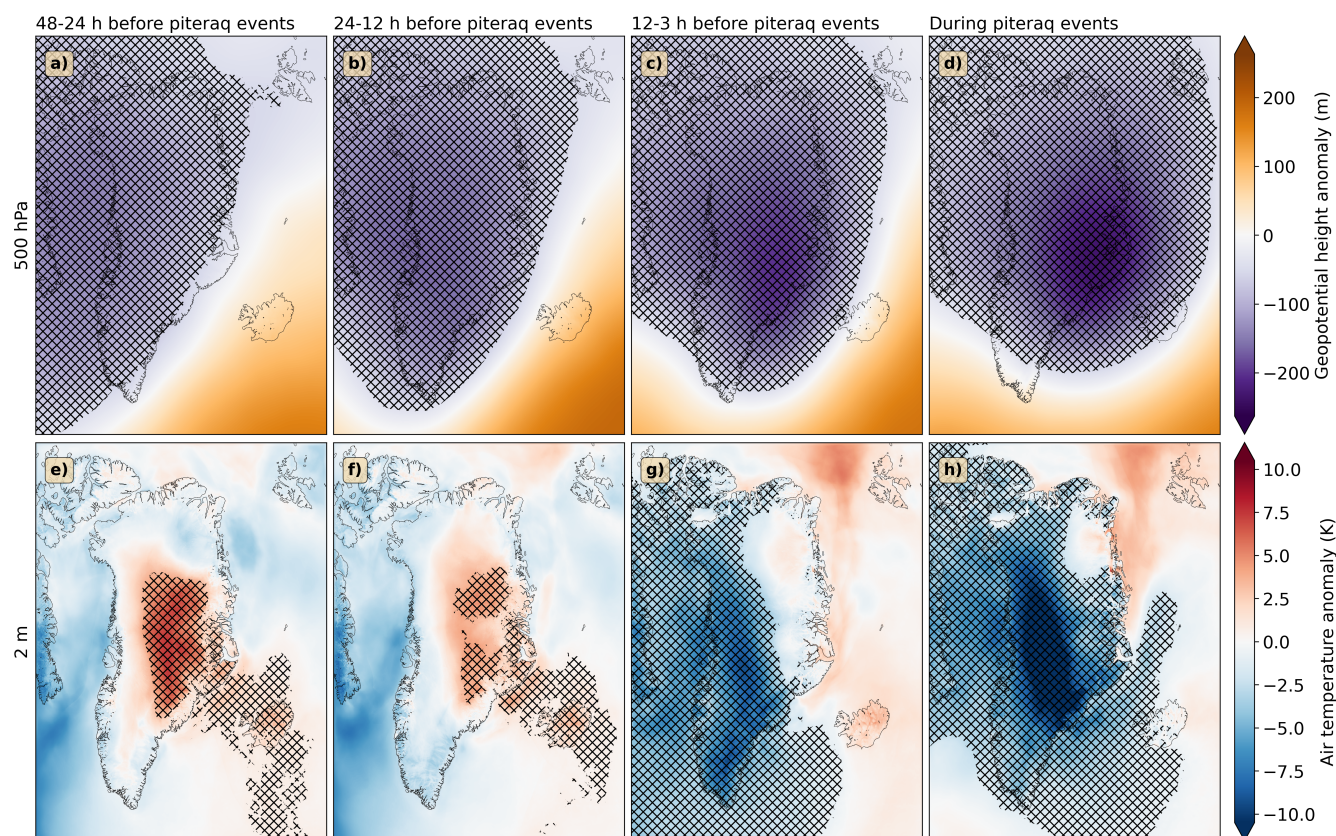


Figure A4. Synoptic evolution of winter (DJF) anomalies prior to and during piteraq events, relative to the 1991–2024 DJF climatology. Columns show time windows leading to the event: 48–24 h, 24–12 h, and 12–3 h prior, followed by the event itself. The top row displays 500 hPa geopotential height anomalies, while the bottom row shows 2-m air temperature anomalies. Stippling indicates regions where anomalies are significantly different from climatology ($p < 0.05$) based on a Wilcoxon rank-sum test.

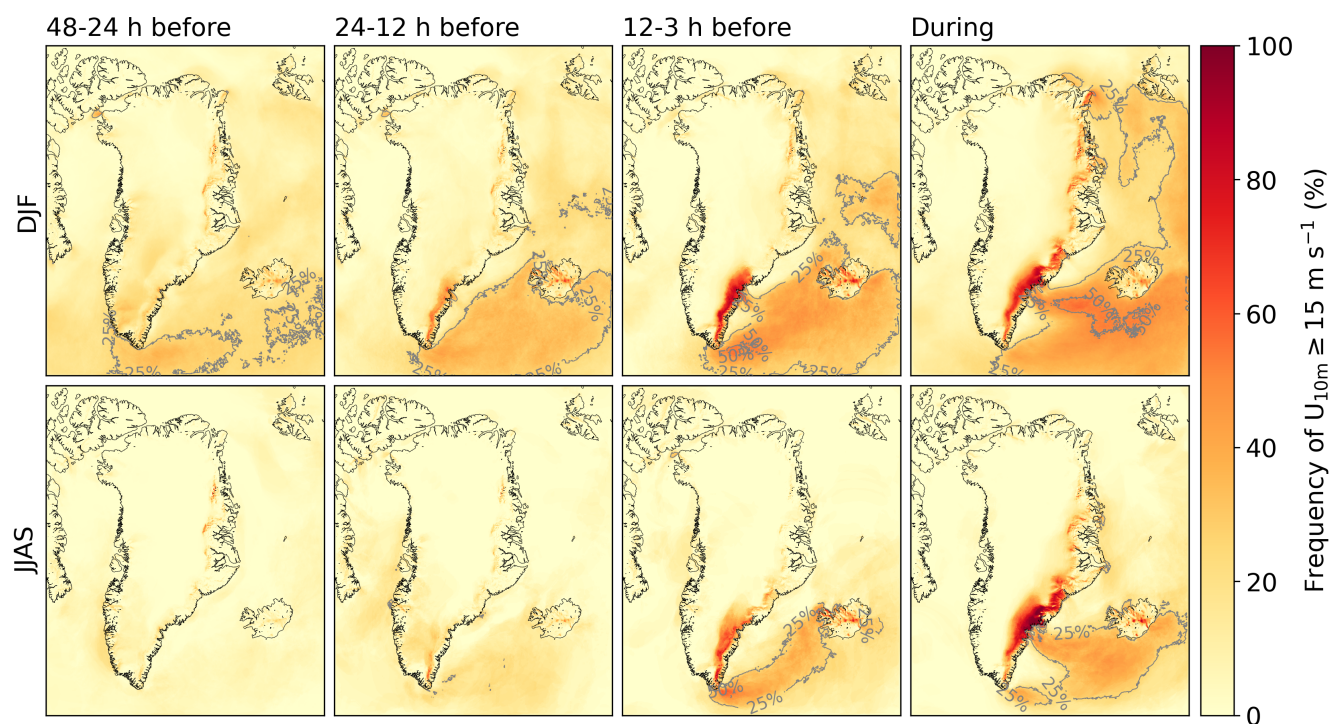


Figure A5. Synoptic evolution of U_{10m} stronger than 15 m s^{-1} prior to and during pitera events. Columns show time windows leading to the event: 48–24 h, 24–12 h, and 12–3 h prior, followed by the event itself. The top row displays winter (DJF) and bottom row displays extended summer (JJAS).

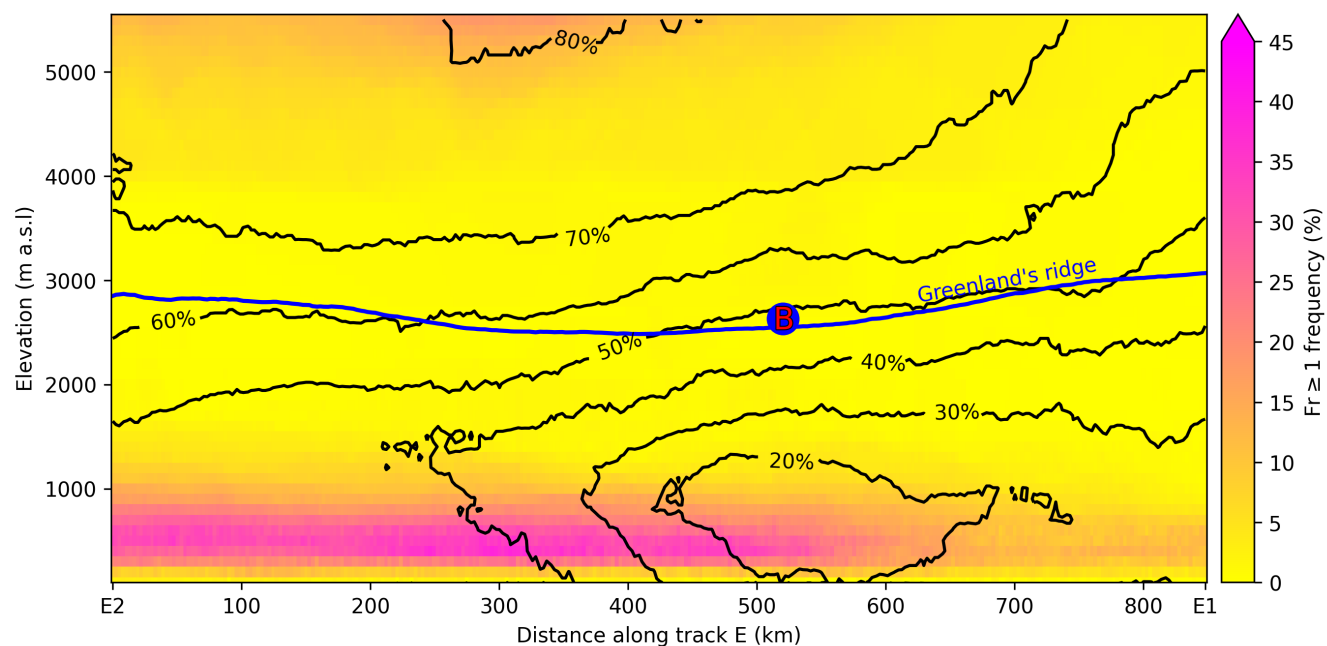


Figure A6. Frequency of westerlies (270 ± 45 degrees, contours) and supercritical flow ($Fr \geq 1$, color-filled contours) through segment E for DJF between 1991–2024 during piteraq events. The Greenland Ice Sheet crest, parallel to segment E, is indicated in blue and point B denotes the location and the highest elevation of segment A.

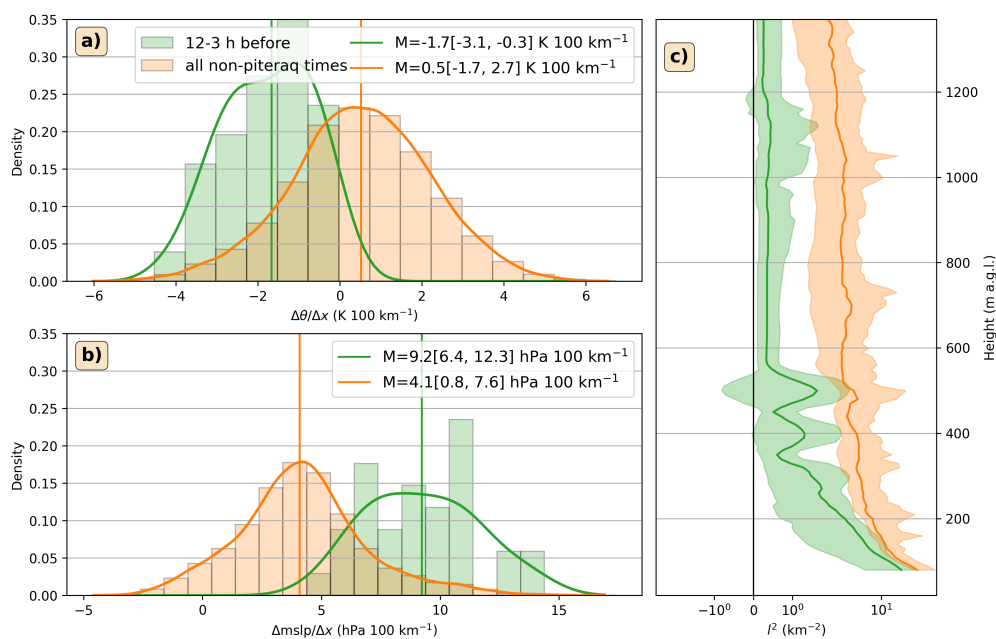


Figure A7. Comparison of atmospheric conditions during piteraq events (green) and the 1991–2024 DJF climatology (orange). Panels (a) and (b) show the difference in potential temperature and mean sea-level pressure, respectively, between the Greenland Ice Sheet crest (point B in Figure 2) and Tasiilaq. Panel (c) displays the vertical profile of the Scorer parameter (l^2) on a logarithmic scale (x-axis) on the Greenland Ice Sheet crest. Solid lines represent mean values, while shaded regions indicate the interdecile range.

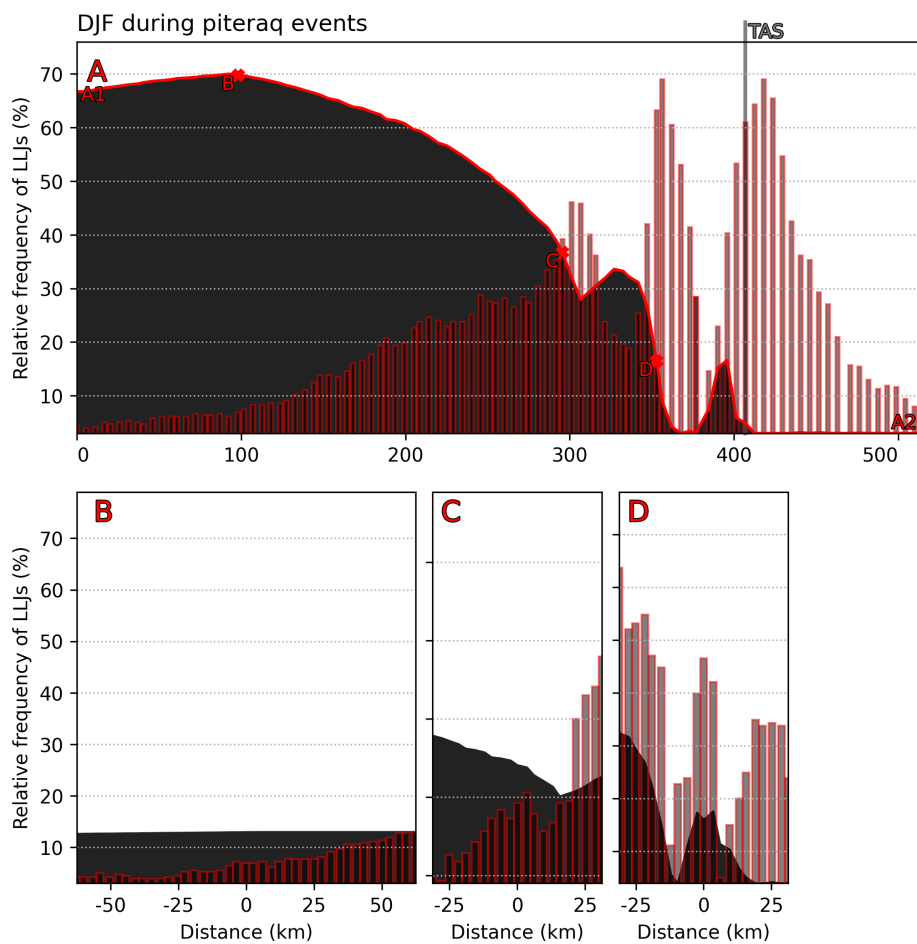


Figure A8. Frequency of low-level jets (LLJs) during the piteraq events for DJF between 1991 and 2024 along the segment A and across segments B, C, and D. The topography of the segments is normalized and displayed in the background. The position of Tasiilaq is marked with a gray vertical line along segment A.

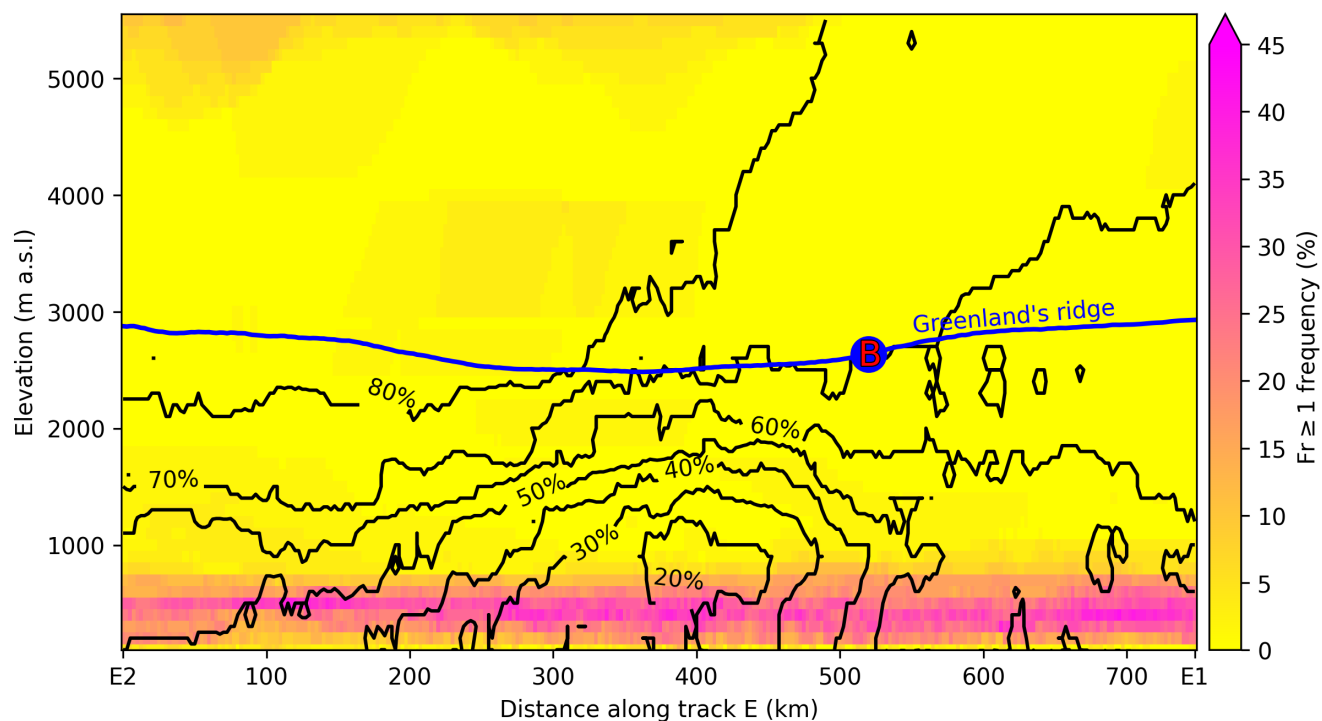


Figure A9. Frequency of westerlies (270 ± 45 degrees, contours) and supercritical flow ($Fr \geq 1$, color-filled contours) through segment E for JJAS between 1991–2024 12–3 hours before piteraq events. The Greenland Ice Sheet crest, parallel to segment E, is indicated in blue and point B denotes the location and the highest elevation of segment A.

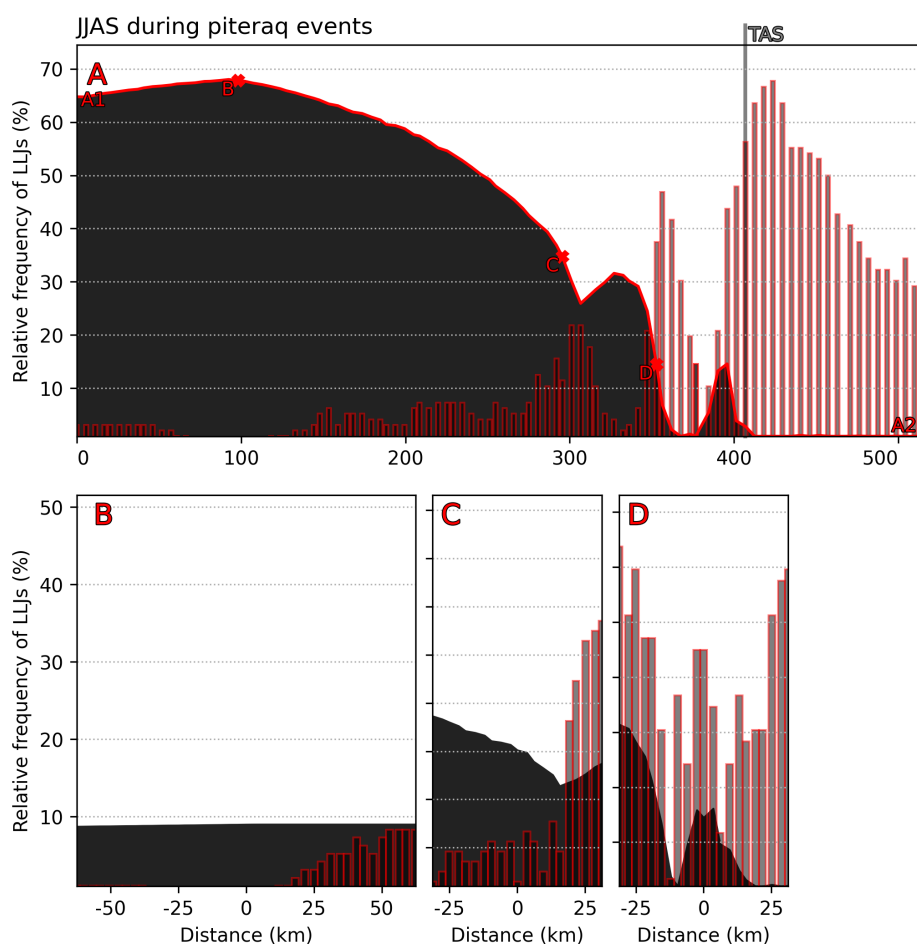


Figure A10. Frequency of low-level jets (LLJs) during the piteraq events for JJAS between 1991 and 2024 along the segment A and across segments B, C, and D. The topography of the segments is normalized and displayed in the background. The position of Tasiilaq is marked with a gray vertical line along segment A.

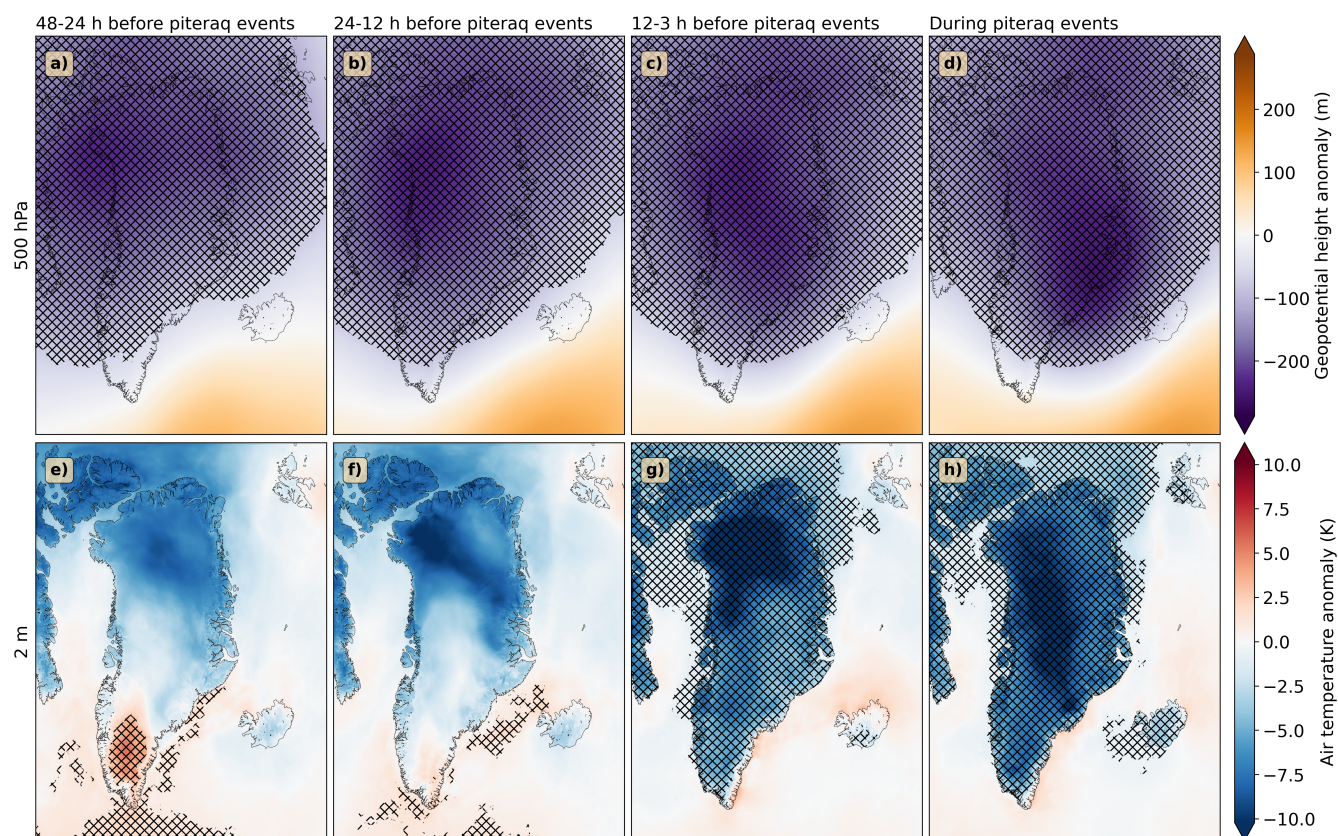


Figure A11. Synoptic evolution of extended summer (JJAS) anomalies prior to and during piteraq events, relative to the 1991–2024 DJF climatology. Columns show time windows leading to the event: 48–24 h, 24–12 h, and 12–3 h prior, followed by the event itself. The top row displays 500 hPa geopotential height anomalies, while the bottom row shows 2-m air temperature anomalies. Stippling indicates regions where anomalies are significantly different from climatology ($p < 0.05$) based on a Wilcoxon rank-sum test.

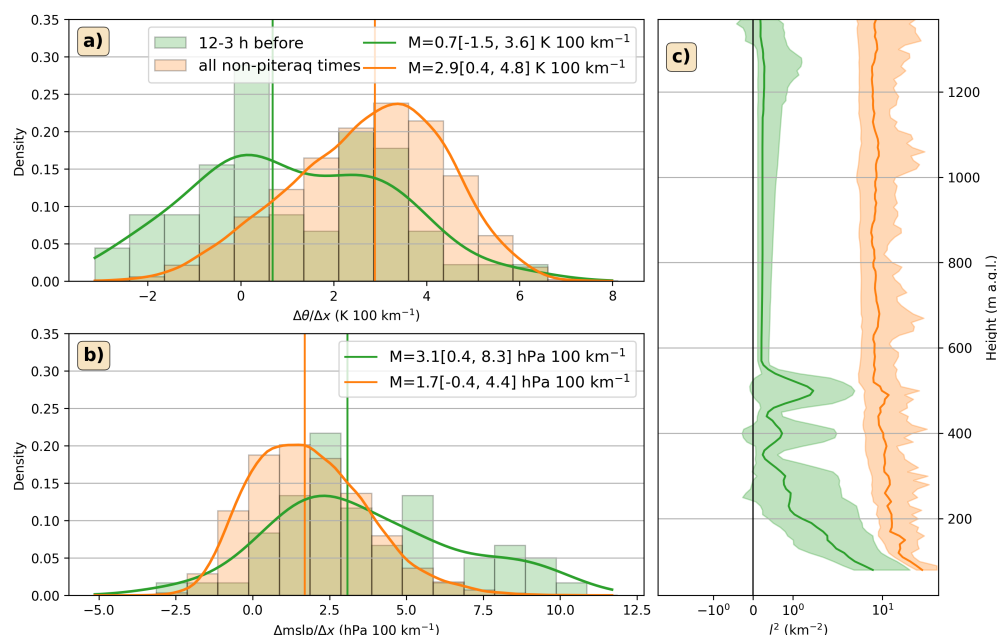


Figure A12. Comparison of atmospheric conditions before piteraq events (green; 3–12 h before onset) and the 1991–2024 JJAS climatology (orange). Panels (a) and (b) show the difference in potential temperature and mean sea-level pressure, respectively, between the Greenland Ice Sheet crest (point B in Figure 2) and Tasiilaq. Panel (c) displays the vertical profile of the Scorer parameter (l^2) on a logarithmic scale (x-axis) on the Greenland crest. Solid lines represent mean values, while shaded regions indicate the interdecile range.

Author contributions. The inspiration for the paper was brought by JCY and JA, the concept and methodology were developed by TS and KFH, the original paper draft was written by TS, and the data were processed and analyzed by TS. The interpretation of results as well as revision and editing of the final paper draft were made by all authors.

605 *Competing interests.* The authors declare that they have no conflict of interest.

Disclaimer. Statement on inclusion in global research: This research investigates piteraq events, severe downslope windstorms that impact the communities of southeast Greenland. The use of the term piteraq, originated from the local East Greenlandic Tunumiisut language, reflects the generational knowledge of such extreme coastal hazards. We are grateful for the local perspectives and lived experiences that helped contextualize the physical impacts of these events. We remain committed to sharing our findings to support operational safety and climate resilience in the region.

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Acknowledgements. All authors would like to acknowledge Kerstin Rasmussen, Simon de Villiers, and Caroline Drost Jensen for the fruitful discussions. All authors would like to acknowledge the joint effort from Florina Roana Schalamon, Jakob Steiner, and Matthew Blaise Swintanek on downloading the Copernicus Arctic Reanalysis. Gemini 3.1 Pro (Google) is acknowledged for improving the manuscript readability and refining the English prose.



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