



Characteristics of wintertime multilayer clouds during air mass transformations at Ny-Ålesund, Svalbard

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Abstract.

Multilayer clouds (MLCs) are vertically stacked cloud systems that occur frequently in the Arctic and strongly influence the region's surface energy balance. Yet the air mass conditions driving their occurrence and vertical structure remain poorly understood. Here, we investigate 19 years of wintertime MLC observations at Ny-Ålesund, Svalbard, an Atlantic-sector Arctic gateway subject to frequent warm air intrusions (WAIs) and cold air outbreaks (CAOs). We use radiosonde profiles to identify potential cloud layers and verify them using collocated cloud radar. Air mass regimes are classified from equivalent potential temperature profiles using rotated principal component analysis. The radiosonde profiles show that thermodynamic conditions favorable for MLCs are common, with 54.4 % of winter profiles containing multiple potential cloud layers. Radar verification reduces this to 27.3 %, as some potential layers lack hydrometeors, particularly in the upper troposphere. Across air mass regimes, WAIs are the cloudiest regime and favor vertically complex cloud structures, with 43.8 % of radar-verified WAI profiles containing multiple layers compared with 21 % during CAOs. Idealized seeder-feeder calculations indicate that falling small ice particles are more likely to reach the lower cloud during CAOs, whereas for larger particles, vertical layer separation is the main constraint regardless of synoptic conditions. Radiatively, MLCs modulate surface longwave cooling most effectively during CAOs, while during WAIs single- and multilayer clouds produce similar effects. Thus, the air mass regime, rather than the number of cloud layers, modulates how MLCs affect the Arctic winter surface energy budget, providing an observational basis for evaluating MLCs in models.

1 Introduction

The Arctic has warmed much faster than the global average in recent decades, a phenomenon known as Arctic amplification (AA) (Serreze and Barry, 2011; Rantanen et al., 2022). This amplified warming is particularly pronounced in winter, when the absence of solar radiation makes the surface energy budget strongly controlled by longwave radiative exchange (Pithan and Mauritsen, 2014; Bintanja et al., 2011). During the polar night, clouds play a central role by enhancing downwelling longwave radiation, thereby reducing surface radiative cooling and influencing near-surface temperature, boundary-layer stability, and sea-ice growth (Shupe and Intrieri, 2004; Morrison et al., 2012). However, the magnitude of this cloud radiative effect depends



not only on cloud presence, but also on cloud lifetime, phase, optical depth, and vertical structure (Shupe and Intrieri, 2004;

25 L'Ecuyer et al., 2019; Hang et al., 2019; Wang et al., 2021).

This complex dependence on cloud properties makes Arctic clouds a persistent source of uncertainty in climate models. Mixed-phase clouds are especially difficult to represent as they depend on the balance between supercooled liquid water, ice formation, sedimentation, turbulence, and radiative processes (Morrison et al., 2012; Kay et al., 2016; Hofer et al., 2024). Multilayer clouds (MLCs) are vertically separated cloud layers within the same atmospheric column, adding another level
30 of complexity. Unlike single-layer clouds (SLCs), MLCs can interact through both microphysical and radiative pathways. Microphysically, ice crystals or frozen hydrometeors from an upper cloud layer may sediment into a lower cloud through the seeder–feeder mechanism (Bergeron, 1965). If these particles survive sublimation in the subsaturated gap between the cloud layers, they can seed ice to the lower layer and influence cloud phase, precipitation production, and lifetime through an increase in cloud ice and promotion of precipitation formation, while also impacting the Wegener–Bergeron–Findeisen
35 process (Wegener, 1911; Bergeron, 1928; Findeisen, 1938; Korolev, 2007; Wallentin et al., 2026b). Radiatively, an overlying cloud layer can increase the downwelling longwave radiation received by a lower cloud, thereby reducing the net longwave cooling at the lower cloud top (Luo et al., 2008; Christensen et al., 2013; Adebiyi et al., 2020; Chen et al., 2020; Wallentin et al., 2026b). Because cloud-top radiative cooling helps drive turbulence and vertical mixing, this shielding effect can suppress the development of the lower cloud (Verlinde et al., 2013).

40 MLCs are frequent in the Arctic, although reported occurrence frequencies vary with region, season, observing platform, and even detection method. Observations from satellites (Liu et al., 2012; Schirmacher et al., 2023; Dietel et al., 2024), radiosondes (Silber and Shupe, 2022), and ground-based remote sensing (Nomokonova et al., 2019) consistently show that MLC structures make up a substantial fraction of Arctic cloud occurrence. Vassel et al. (2019) developed a framework that uses both radiosondes and cloud radar to classify Arctic MLCs and their possible seeder–feeder interactions, showing that MLCs occur
45 more frequently in the Arctic than the global mean. Achtert et al. (2026) later adopted a similar framework to examine MLC variability across the Arctic region using observations from expeditions and long-term weather stations. More recent studies have further emphasized that MLCs are common and radiatively important (L'Ecuyer et al., 2019; Hang et al., 2019), while their detailed vertical structure and microphysical interactions remain challenging to represent in models (Wallentin et al., 2025, 2026a).

50 Cloud formation and structure in the Atlantic sector of the Arctic depend heavily on the origin and evolution of incoming air masses (Gierens et al., 2020; Silber and Shupe, 2022; Lackner et al., 2023). As air travels between the mid-latitudes and the Arctic, it undergoes rapid thermodynamic changes driven by surface exchanges, radiation, and turbulence (Pithan et al., 2018; Wendisch et al., 2025). For instance, the poleward transport of warm, moist lower-latitude air during warm air intrusions (WAIs) perturbs the cold, stably stratified Arctic winter atmosphere. These intrusions increase temperature, humidity, cloudiness,
55 and surface energy input, thereby reducing radiative cooling and impacting sea ice (Woods and Caballero, 2016; You et al., 2022). In contrast, cold air outbreaks (CAOs) transport cold, dry Arctic air over warmer ocean surfaces, where the air mass is rapidly modified by turbulent heat and moisture fluxes (Pithan et al., 2018; Afargan-Gerstman et al., 2020). This modification drives boundary-layer growth and supports initial cloud formation, which can later develop into convective clouds. As a result,



these distinct air mass transformations create contrasting thermodynamic environments that strongly dictate cloud formation, structure, and radiative effects.

Although it is widely hypothesized that MLC formation is strongly linked to WAIs (Tan et al., 2023), this relationship remains poorly quantified. For instance, while Silber and Shupe (2022) used back trajectories to show that high-Arctic MLCs during the 2019/20 MOSAiC expedition were typically associated with WAIs, which was further corroborated in a recent study of Dürlich et al. (2026), most studies continue to examine air mass transformations and Arctic MLCs separately. Arctic MLC studies generally document cloud occurrence, phase, and vertical structure (Nomokonova et al., 2019; Vassel et al., 2019; Achtert et al., 2026), whereas air mass transformation studies focus on transport pathways, thermodynamic modifications, and surface energy impacts (Woods and Caballero, 2016; Pithan et al., 2018; You et al., 2022). Consequently, it remains unclear whether WAIs and CAOs systematically promote different MLC frequencies, vertical structures, and seeding potentials. Bridging this gap is critical because while air mass transformation sets the thermodynamic background, the vertical arrangement and microphysical coupling of cloud layers ultimately determine how effectively surface longwave radiation is modified. Therefore, assessing MLCs within the context of air mass transformations is crucial for determining whether the radiative effects of WAIs and CAOs are controlled primarily by the air mass itself, or by the cloud structures that develop within it.

To quantify the link between Arctic air mass transformation and MLCs, we use 19 years of wintertime (December–February; DJF) observations (2006/07–2024/25) from Ny-Ålesund, Svalbard—a site heavily influenced by both WAIs and CAOs (You et al., 2022; Dahlke et al., 2022; Walbröl et al., 2024; Wendisch et al., 2025). Under typical wintertime conditions, the site is characterized by a stably stratified lower troposphere, with temperature inversions below 2 km detected in the majority of radiosonde soundings and a capping inversion near 1 km present in roughly half of all profiles. Near-surface air temperature averages around -10°C , decreasing to about -35°C at 5 km and roughly -60°C in the upper troposphere, and the ground remains snow-covered throughout the wintertime (Maturilli and Kayser, 2017; Maturilli et al., 2019). The adjacent fjord and coastal waters are strongly influenced by the West Spitsbergen Current, and their winter ice state varies considerably from year to year, ranging from largely ice-covered conditions to open-water winters in which Atlantic water reaches Svalbard, with a general shift toward more frequent ice-free winters in recent decades (Tverberg et al., 2019; De Rovere et al., 2022). We combine long-term radiosonde profiles with collocated cloud radar observations to identify potential and radar-verified MLCs. By associating these cloud structures with contrasting air mass regimes, this study examines whether WAIs and CAOs systematically promote distinct MLC frequencies, vertical structures, and seeder-feeder potentials. In doing so, we examine how air-mass transformations and the resulting cloud structures interact to shape the surface radiative effects in the Arctic.

2 Data and Methods

2.1 Observations

We used wintertime (DJF) observations from 2006/07 to 2024/25 at the AWIPEV research base (Alfred Wegener Institute and French Polar Institute Paul-Émile Victor) in Ny-Ålesund, Svalbard, an Atlantic-sector Arctic gateway influenced by oceanic inflow of warm Atlantic water through the West Spitsbergen Current, warm and moist air intrusions (WaMAIs), and marine



cold-air outbreaks (CAOs) (Frank et al., 2025; You et al., 2022; Dahlke et al., 2022). We derived the cloud structure from radiosonde profiles, complemented by cloud radar observations from 2017 onward, and evaluated its radiative impact with surface longwave radiation measurements.

95 Although radiosondes have been launched routinely at 12:00 UTC since 1992, with additional launches during campaigns, we found that residual dry biases of the pre-2006 sensors persist even after homogenization (Maturilli and Kayser, 2016) and affect cloud layer identification with the MLC classification algorithm of Vassel et al. (2019), see Appendix A1. We therefore restricted the analysis to GRUAN-processed RS92 and RS41 profiles available from May 2006 onward (Sommer et al., 2012, 2022), and considered DJF observations from 2006/07 to 2024/25. The GRUAN product provides temperature,
100 relative humidity (RH), pressure, and horizontal wind at 1 s resolution, corresponding to approximately 5 m vertical sampling at typical ascent rates. GRUAN processing applies traceable corrections for radiation effects and sensor-related biases and provides profile-dependent uncertainty estimates, improving consistency between the RS92 and RS41 records (Dirksen et al., 2014; von Rohden et al., 2022; Sommer et al., 2023). Manufacturer-specified relative humidity uncertainties are approximately 5 % for RS92 and 4 % for RS41.

105 Radar reflectivity was obtained from a zenith-pointing 94 GHz Doppler cloud radar operated by the University of Cologne since June 2016 within the Arctic Amplification: Climate Relevant Atmospheric and Surface Processes, and Feedback Mechanisms (AC)³ project (Wendisch et al., 2017); its calibration and limitations are described in Küchler et al. (2017). We used the Cloudnet target classification product, archived from February 2017 (Ebell, 2026), for DJF 2017–2024/25. It has a temporal resolution of 30 s and a vertical resolution of 20 m, covering altitudes from 223 m to 10 km (Chellini et al., 2023).

110 To assess the radiative effects of the identified cloud structures, we used 1 min surface longwave radiation measurements from the Baseline Surface Radiation Network (BSRN) station at Ny-Ålesund (Driemel et al., 2018; Maturilli, 2020). For each radiosonde launch, downwelling ($L_{\downarrow}$) and upwelling longwave ($L_{\uparrow}$) radiation were averaged from launch time to 30 min after launch. The net longwave (L_{net}) radiation was calculated as the difference between downwelling and upwelling longwave radiation (positive downward). Using alternative averaging windows of 15 and 60 min produced no significant differences.

115 2.2 MLC detection algorithm

The cloud layer structure in each radiosonde profile was identified based on the MLC detection algorithm of Vassel et al. (2019). We implemented the method and incorporated the refinements introduced by Achtert et al. (2026). We applied the algorithm to the full wintertime radiosonde record and verified the detected layers using collocated radar reflectivity measurements when available.

120 For each sounding, the radiosonde humidity profile was first analyzed to identify potential cloud layers, which follows the general radiosonde–radar framework of Vassel et al. (2019). Previous studies adopted slightly different thresholds to account for radiosonde uncertainty. For example, for relative humidity with respect to liquid, $\text{RH} \approx 95\text{--}96\%$ has been used to identify supercooled liquid layers (Silber et al., 2020), while for relative humidity with respect to ice, $\text{RH}_i = 100\%$ has been used to indicate ice-saturated, ice-containing layers (Vassel et al., 2019). Here, a layer was treated as supersaturated when the air

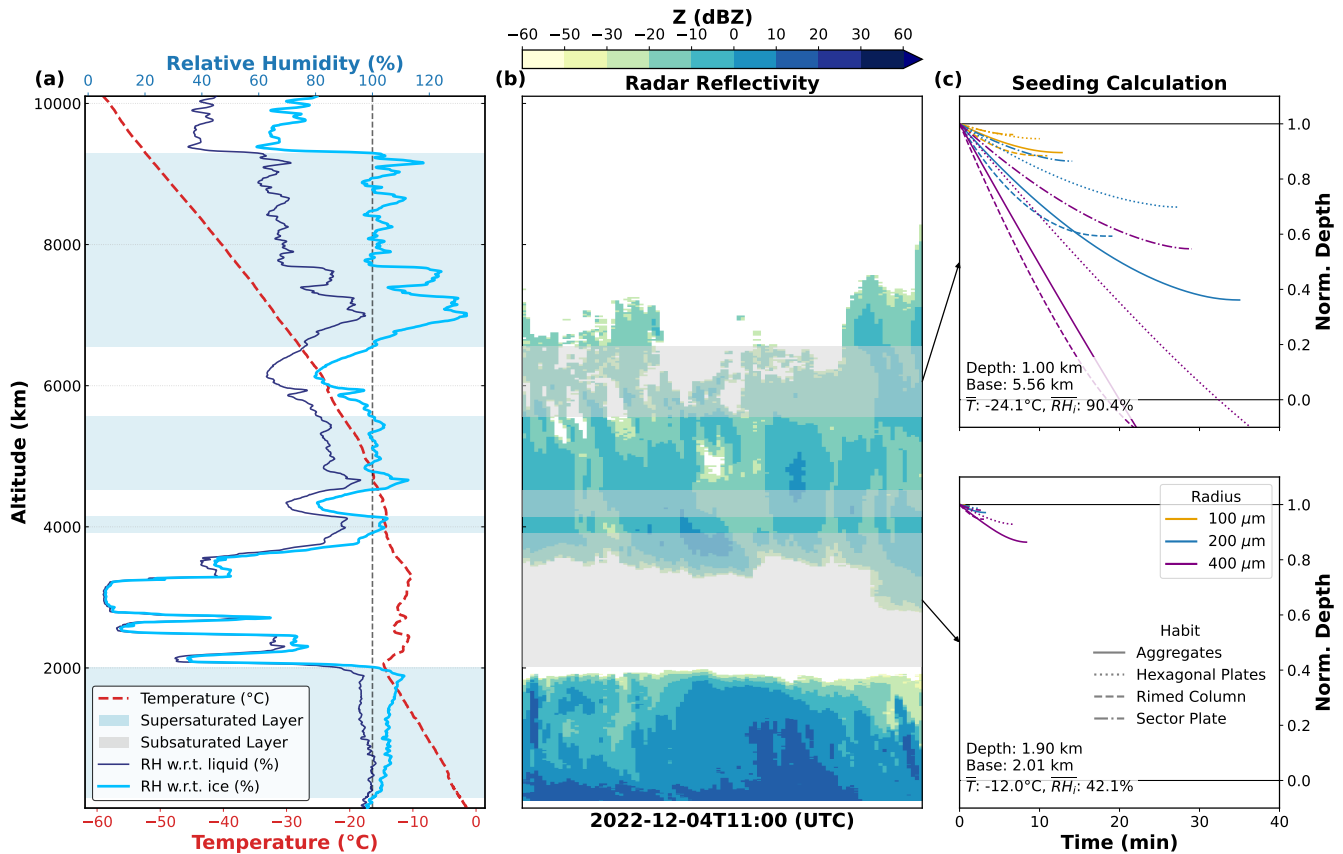


Figure 1. (a) Radiosonde profile from Ny-Ålesund (Svalbard) from the 11:00 UTC launch on 4 December 2022 showing temperature (red) and relative humidity with respect to liquid water (blue) and ice (light-blue) as a function of altitude. Blue shaded regions indicate the supersaturated layers as classified by the MLC algorithm. (b) Collocated time–height radar reflectivity Z (dBZ) from 30 min before the radiosonde launch to 30 min after the ascent reached 10 km altitude. Grey shaded regions indicate subsaturated layers. (c) Idealized seeding calculation for the upper and lower subsaturated layers (indicated by the arrows). The curves show the fall distance of ice crystals with time in normalized depth for three initial particle radii (line colors) and four ice habits (line styles). An ice crystal is classified as *seeding* (upper panel) if it reaches the normalized layer depth 0 before the particle fully sublimates; otherwise it is *non-seeding* (lower panel). Each inset lists the properties of the corresponding subsaturated layer used in the calculation: the layer depth, and the layer averages of temperature $\bar{T}$ and relative humidity with respect to ice $\overline{RH_i}$ (overbars denote vertical averages across the layer).

reached saturation ($\geq 100\%$) with respect to either ice or liquid, while all other parts of the profile were treated as a subsaturated layer.

Figure 1a illustrates the layer segmentation for 4 December 2022. The two upper supersaturated layers show humidity fluctuations around the 100 % threshold. Such fluctuations can reflect measurement noise and small-scale variability, potentially causing artificial splitting into thin layers. To limit this effect, we applied minimum thickness criteria similar to those of Achtert



et al. (2026): supersaturated layers were required to be at least 100 m thick, increasing to 500 m above 5 km, while subsaturated gaps between them were required to be at least 150 m thick. After initial segmentation, layers that did not meet the criteria were merged iteratively with a neighboring layer, starting with the thinnest, until all remaining layers satisfied the thickness requirements.

However, there are few cases of subsaturated layers that satisfy the thickness condition but have a consistently high value of relative humidity above 95 %. This might artificially increase the count of both multilayer systems and seeding cases. Specifically, a layer may be classified as subsaturated even when it is actually a cloud, meaning the assumed cloud gap between layers does not physically exist. Furthermore, because the seeding calculation accounts for the dryness of intermediate layers, these misclassified, high relative humidity gaps are highly likely to erroneously satisfy the conditions for successful seeding. To ensure that the identified subsaturated layers represent genuinely drier regions, an uncertainty-aware correction was applied. A layer that was initially labeled subsaturated was reclassified as supersaturated when RH^* stayed above $100\% - \delta RH$ throughout the layer, where RH^* denotes the relative humidity values within the subsaturated layer and δRH represents the relative humidity uncertainty of the radiosonde used. Although lowering the initial humidity threshold could also reduce these near-saturated gaps, we retained 100 % as the physical saturation threshold to avoid systematically expanding all detected saturated layers and restricted the correction to gaps whose apparent subsaturation remained within measurement uncertainty.

After the final supersaturated layers were defined, the number of supersaturated layers was used to describe the cloud situation for that launch. If no supersaturated layer was found, the profile is classified as cloud-free. If one supersaturated layer is found, it is classified as a single layer cloud. If more than one supersaturated layer is found, it is classified as a MLC. Cloud cases are further separated into warm and cold conditions using the cloud-top temperature. A case is warm when the cloud-top temperature is above 0 °C and cold when it is below 0 °C. This addition is adapted from Achtert et al. (2026).

To verify the presence of clouds within these supersaturated layers, we used collocated cloud radar reflectivity measurements. For each launch, radar observations were evaluated from 30 min before launch until 30 min after the sounding reached 10 km altitude. A radiosonde-derived supersaturated layer was kept as a cloud layer only when the radar echoes were present through at least 50 % of the minimum layer depth of 100 m (50 % threshold). Here, we retained only a minimum-thickness criterion, as supersaturated layers at higher altitudes frequently show little or no corresponding cloud signal, which can be partly due to signal attenuation. This also may reflect the aerosol-limited conditions of the Arctic atmosphere (Mauritsen et al., 2011), together with the rapid sedimentation of ice particles. An example of this behavior is shown in Fig 1b. Here, the radiosonde identified a thick, vertically extensive supersaturated layer aloft. However, the corresponding radar reflectivity reveals that physical cloud hydrometeors did not uniformly fill this entire supersaturated layer. Instead, active ice cloud formation was localized in the middle part of the layer—around -38 °C , the threshold where homogeneous freezing is expected. At this altitude, despite a distinct decrease in relative humidity with respect to ice, the air remained supersaturated and cold enough to drive ice nucleation. As these crystals sedimented downward through the underlying saturated air, they created a distinct vertical structure where radar-detectable features occupied only a portion of the lower half of the supersaturated layer. A similar vertical structure and radar signal were reported by Ansmann et al. (2025), who found that transported aerosols such as wildfire smoke can initiate ice formation by acting as ice-nucleating particles in supersaturated upper-tropospheric layers.



165 To investigate the potential seeding on identified MLCs, we performed idealized sublimation calculations following Vassel
et al. (2019) and Proske et al. (2021). Cold MLC systems were divided into triplets comprising two saturated layers separated
by a subsaturated gap. An ice particle released from the upper cloud base was classified as seeding if it reached the lower
cloud top before complete sublimation, and as non-seeding otherwise. The calculations considered aggregates, plates, rimed
columns, and sector plates with initial radii of 100 μm , 200 μm and 400 μm . Figure 1c shows no seeding across the lower gap
170 and seeding across the upper gap only for 400 μm particles, excluding sector plates. When radar observations were available,
idealized seeding cases were reclassified as non-seeding if there was no radar signal between the two cloud layers. Profiles
containing both seeding and non-seeding triplets were classified as mixed seeding/non-seeding MLC cases.

Both radiosonde and radar data have limitations in determining cloud layers. Radiosondes can identify potential cloud lay-
ers, but this does not necessarily mean that a cloud has formed, especially in the relatively clean Arctic environment, where
175 cloud formation may be aerosol-limited. Therefore, using radiosonde data alone shows only the potential for MLC formation
and likely leads to an overestimation of cloud occurrence. However, radiosonde profiles provide important thermodynamic in-
formation by identifying layers where clouds can form and persist, as well as layers where falling hydrometeors may begin to
sublimate or evaporate. In this way, the radiosonde observations help distinguish thermodynamic cloud layers from precipita-
tion falling through subsaturated air. Radar reflectivity complements the radiosonde data by constraining whether hydrometeors
180 are actually present within the identified supersaturated layers. However, the radar signal can be attenuated by lower cloud lay-
ers or precipitation, which may reduce its sensitivity at higher altitudes. In addition, small ice crystals may produce weak
reflectivity and can therefore be missed by the radar (Wang et al., 2022). For this reason, we used both the radiosonde-only and
radar-verified statistics to provide a possible range of MLC occurrence frequency. Furthermore, another uncertainty may arise
from the drifting of the radiosonde. The radar observes the vertical column above the site, whereas the radiosonde is transported
185 horizontally by the wind, meaning that the two instruments may not sample the same air column. In this study, this source of
uncertainty is expected to have only a limited influence on the analysis of air mass transformation regimes because WAIs and
CAOs are generally associated with relatively coherent synoptic-scale thermodynamic structures. Kiszler et al. (2023) showed
that radiosonde drift direction can influence the observed humidity profile; however, the differences in their study were mainly
confined to the lowest 500 – 600 m. Part of this variability may also be related to the local topography, which reaches approx-
190 imately 556 m, rather than solely to whether the radiosonde drifted over land or ocean. Radiosonde drift may therefore affect
the exact height and position of individual cloud boundaries, but it is unlikely to change the broader thermodynamic differences
between the air mass regimes.

2.3 Air mass transformation classification

WAMAI or WAI are episodes characterized by the transport of air from lower latitudes into the Arctic, bringing anomalously
195 high temperature and humidity (You et al., 2022). They are often detected from enhanced poleward moisture transport, for
example, by applying thresholds to the vertically integrated northward moisture flux or integrated vapor transport across an
Arctic boundary near 70 °N, together with duration and spatial-extent criteria to avoid false alarms from short, narrow transport
spikes (Woods et al., 2013). Complementary Lagrangian approaches use air mass trajectories to diagnose pathways, the air



mass origin, and moisture uptake along the route (Woods and Caballero, 2016; Ali and Pithan, 2020; Silber and Shupe, 2022; Murto et al., 2022). During poleward transport, diabatic processes, including cloud microphysics, radiation, and turbulence, can substantially modify the air mass, shaping the thermodynamic signature observed over the Arctic (Pithan et al., 2018; Wendisch et al., 2025).

During the Arctic winter, the absence of solar heating allows strong radiative cooling to maintain a cold and stably stratified lower troposphere. Under these conditions, the arrival of lower-latitude air masses produces a pronounced warm and moist anomaly in vertical atmospheric profiles. This vertical signature motivates a profile-based classification approach that relies on the full vertical structure of each sounding. We used the equivalent potential temperature, θ_e , to capture both changes in temperature and moisture, calculated using the approximation of Bolton (1980):

$$\theta_e = \theta \exp \left(\frac{L_v r}{c_{p,dry} T_{LCL}} \right), \quad (1)$$

where θ is potential temperature, L_v is the latent heat of vaporization, r is the water vapor mixing ratio, $c_{p,dry}$ is the specific heat of dry air at constant pressure, and T_{LCL} is the temperature at the lifting condensation level. Because our classification relies on the full vertical structure of θ_e , it is essential that θ_e profiles are comparable and not affected by missing data or measurement errors. Therefore, we first quality-checked the wintertime radiosonde dataset by visually inspecting the profiles and excluding soundings that showed unrealistic or anomalous deviations from the expected vertical structure, such as extended layers of nearly constant temperature. This screening removed approximately 1 % of the total profiles. Then, each sounding was interpolated onto a common vertical grid from 50 m to 10 km with 25 m spacing to ensure consistent dimensions across profiles. Finally, we applied principal component analysis (PCA) (Jolliffe, 2002) to leverage the full vertical profile of θ_e and capture vertically displaced signatures, while reducing dimensionality. Vertical temperature and specific humidity profiles showed little seasonality within DJF, so we did not apply deseasonalization, which could otherwise introduce artificial signals.

Before performing the PCA, the θ_e profiles were standardized level by level by subtracting the wintertime mean and dividing by the wintertime standard deviation at each height. This prevents altitude levels with larger raw variance from dominating the decomposition. To improve interpretability, we applied an orthogonal varimax rotation to the retained PC subspace (Kaiser, 1958; Richman, 1986). This rotation preserves the total explained variance while redistributing it among the modes, effectively concentrating the loadings into more localized vertical layers. Rotated PCA approaches have a long history in atmospheric applications (Richman, 1986; Hannachi et al., 2007), specifically with vertical profiles (Langford et al., 1995; Bukowski et al., 2017). The resulting rotated components (RCs) in Fig. 2 partition the dominant θ_e variability into three vertically distinct structures that can be interpreted as lower-, middle-, and upper-tropospheric contributions. We used the associated rotated-component scores as compact indices of each sounding's thermodynamic state. In particular, the RC2 and RC1 indices describe variability in the lower and mid-troposphere, respectively, where the WAI thermodynamic signature is most pronounced. These levels are consistent with previous studies showing that winter Arctic moist intrusions are associated with strong warming and moistening in the lower to middle troposphere (Woods and Caballero, 2016).

This approach reduces reliance on defining air mass regimes from a single thermodynamic threshold at a prescribed pressure level, which can be sensitive to the chosen variable, altitude, and local thermodynamic structure. Instead, the classification

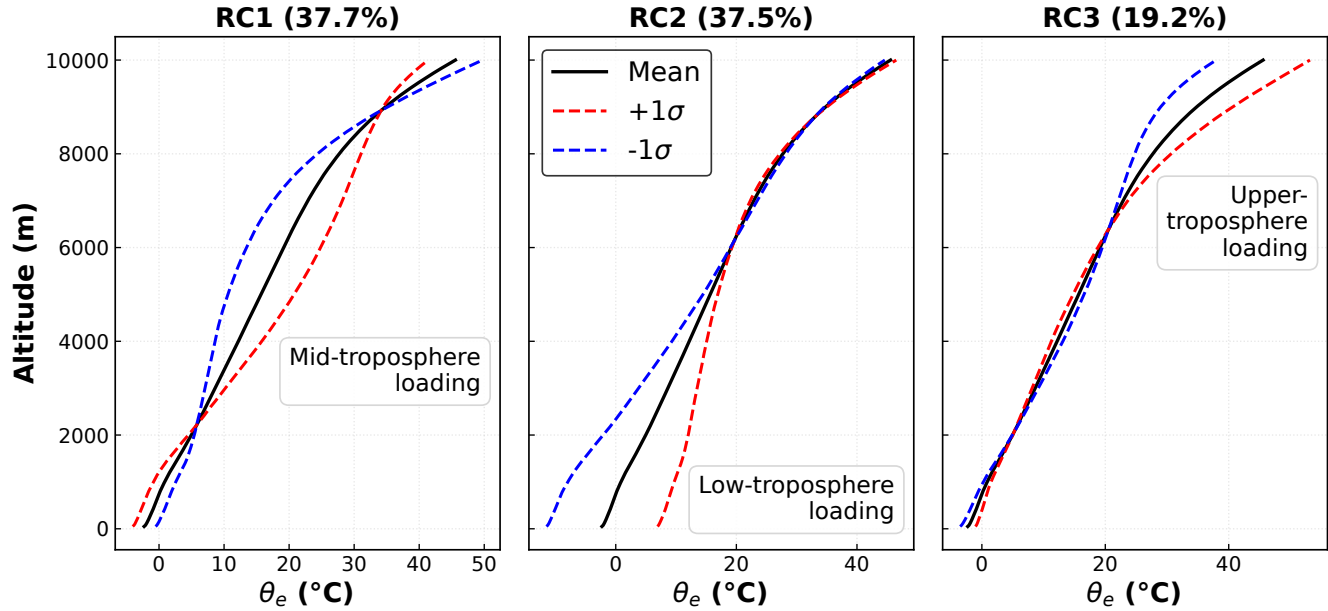


Figure 2. Wintertime θ_e profiles illustrating the thermodynamic structures associated with the first three varimax-rotated components (RCs), derived from standardized θ_e profiles. Black curves show the mean θ_e profile, while red and blue dashed curves show the mean plus and minus the corresponding component loadings, representing positive and negative one-standard-deviation perturbations, respectively. Each panel title gives the percentage of variance explained by that mode. Varimax rotation redistributes variance among modes, producing more vertically localized structures.

is based on the dominant vertically coherent θ_e variability in the observed sounding profiles. Because both principal components and rotated components represent this variability as linear modes about the mean state, positive and negative component scores describe opposite thermodynamic anomalies along the same mode. We then applied symmetric thresholds to the rotated-component scores to identify warm and cold air mass transformation regimes. Positive anomalies in the RC1 and RC2 indices indicate enhanced middle- and lower-tropospheric θ_e , respectively, consistent with the warm and moist structure of WAIs. Conversely, CAOs are associated with the opposite thermodynamic regime, in which cold and relatively dry polar air is transported out of the Arctic and often undergoes rapid transformation over comparatively warmer ocean surfaces (Pithan et al., 2018; Papritz and Grams, 2018). They are commonly identified from the contrast between the lower-tropospheric air temperature and the underlying sea-surface temperature, often using the potential temperature difference between the surface and 850 hPa as a measure of air mass instability (Dahlke et al., 2022). This instability is also reflected in the vertical thermodynamic structure of the lower troposphere, thus making the profile-based classification used here an alternative way to identify CAO-like thermodynamic conditions.

Accordingly, soundings with the RC1 index or RC2 index exceeding one standard deviation above their respective wintertime means were denoted as +RC1 and +RC2, respectively. Soundings for which only one of the two indices exceeded this

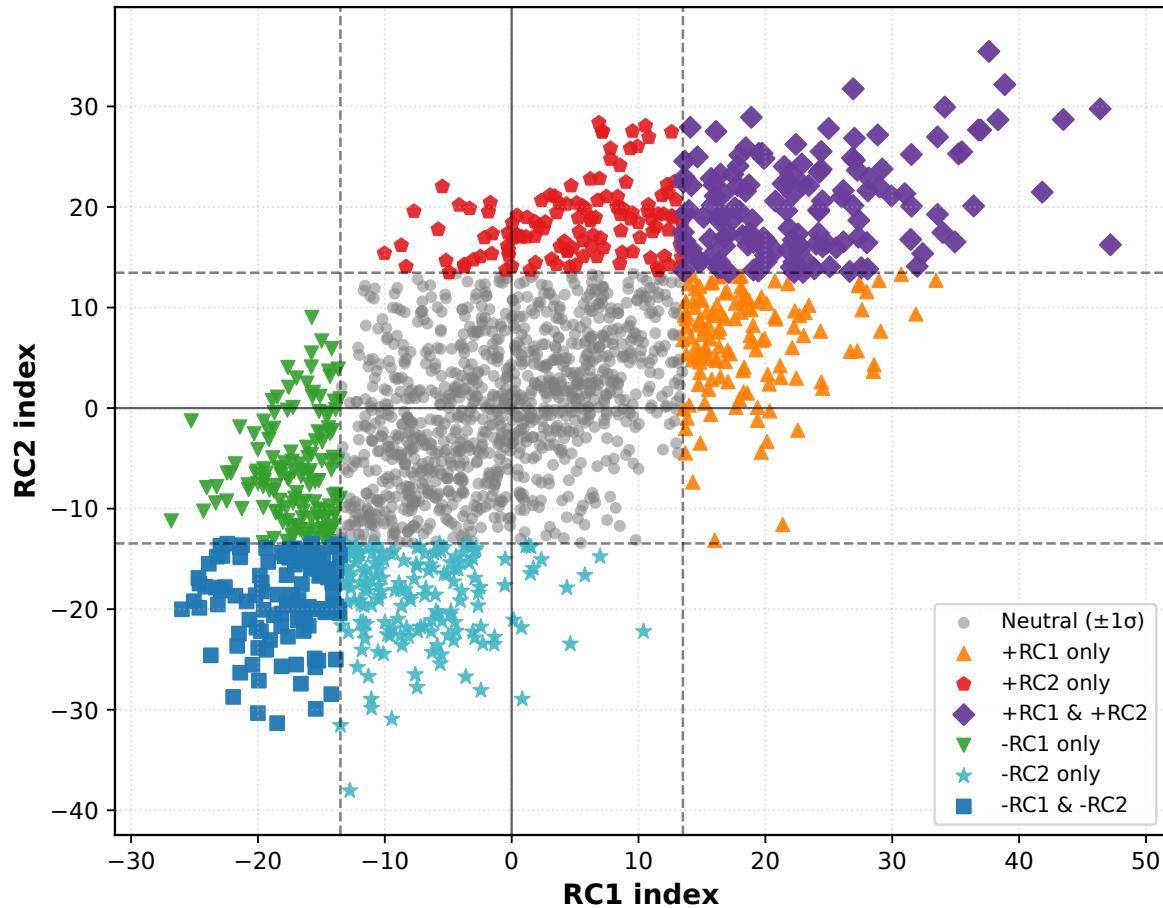


Figure 3. Classification of air mass transformation regimes in RC1 index and RC2 index space. Each point represents one radiosonde profile. The RC1 index describes variability in the mid-tropospheric θ_e structure, while the RC2 index describes variability in the lower-tropospheric θ_e structure. Dashed lines indicate the $\pm 1\sigma$ thresholds used to separate positive and negative regimes, and solid lines mark zero. A profile is labeled +RC1 when the RC1 index exceeds $+1\sigma$ and –RC1 when it falls below -1σ , and similarly for RC2. Grey points denote neutral profiles within $\pm 1\sigma$ for both indices. Colored points show profiles classified into single-index (+RC1, –RC1, +RC2, –RC2) or combined (+RC1 & +RC2, –RC1 & –RC2) regimes.

threshold were further denoted as +RC1-only or +RC2-only, while soundings for which both indices exceeded the threshold simultaneously were denoted as the joint +RC1 & +RC2 regime, see Fig. 3. These positive regimes were used to identify WAIs. The corresponding negative-anomaly cases (–RC1-only, –RC2-only, and the joint –RC1 & –RC2 regime) were defined analogously for soundings falling more than one standard deviation below their respective wintertime means, and used to identify CAOs. We also tested percentile-based thresholds to identify WAIs and CAOs from the RC scores and found no substantive differences; therefore, we retained the standard deviation-based thresholds.



To characterize the atmospheric state of these groups, composite radiosonde profiles of temperature, specific humidity, and meridional wind were constructed.

255 2.4 Reanalysis data for large-scale features

To place the local radiosonde observations in a synoptic-scale context, we used the European Centre for Medium-Range Weather Forecasts reanalysis ERA5 (Hersbach et al., 2020). We retrieved 6-hourly winter fields (DJF) for 2006/07–2024/25 over the Arctic domain, interpolated to a $0.25^\circ \times 0.25^\circ$ grid. For each radiosonde launch, the nearest ERA5 analysis time within the 6-hourly sampling was selected to represent the large-scale environment contemporaneous with the sounding. For our syn-
260 optic diagnostics, we used geopotential height at 500 hPa (Z_{500}) to characterize mid-tropospheric circulation, and temperature at 850 hPa (T_{850}) to capture lower-tropospheric thermal advection, alongside with the vertically integrated northward water vapor flux (f_w ; $\text{kgm}^{-1}\text{s}^{-1}$) for quantifying meridional moisture transport.

These fields were composited for the WAI and CAO groups to characterize the large-scale atmospheric state associated with each class and to evaluate their synoptic consistency. A two-sided Student's t -test ($p < 0.01$) was used to assess grid-point
265 significance similar to You et al. (2022).

3 Results and Discussion

3.1 Vertical profile and synoptic context of the rotated PCA classes

Figure 4 shows composite vertical profiles of temperature, specific humidity, and meridional wind for the +RC1-only, +RC2-only, and joint +RC1 & +RC2 regimes (warm air intrusions), and the corresponding –RC1-only, –RC2-only, and joint
270 –RC1 & –RC2 regimes (cold air outbreaks), relative to the wintertime mean, illustrating the mean conditions and within-regime variability for each classification. Rather than only representing statistical groupings in profile space, these classes reveal vertically coherent structures that resemble the expected signatures of warm, moist intrusion-like conditions and cold, dry outbreak states. The vertical profiles also provide a first indication of how the RC1 and RC2 indices differ in their thermodynamic expression, particularly in the lower troposphere. We then examined whether these local vertical structures are
275 consistent with the associated large-scale circulation patterns using synoptic-scale composites.

3.1.1 Warm air intrusions

For the positive-score regimes, all groups are warmer and moister than the wintertime mean, but their vertical structures differ substantially as shown in Fig. 4a–c. The +RC1-only group shows its strongest temperature anomaly in the middle troposphere, while the lower-tropospheric warming, moistening, and meridional-wind anomaly are relatively weak. This suggests a weak
280 or decaying WAI regime, in which the air masses have already been partly transformed by the sea-ice surface before reaching Ny-Ålesund station, or where the warm anomaly remains elevated and does not fully penetrate the stable lower troposphere.

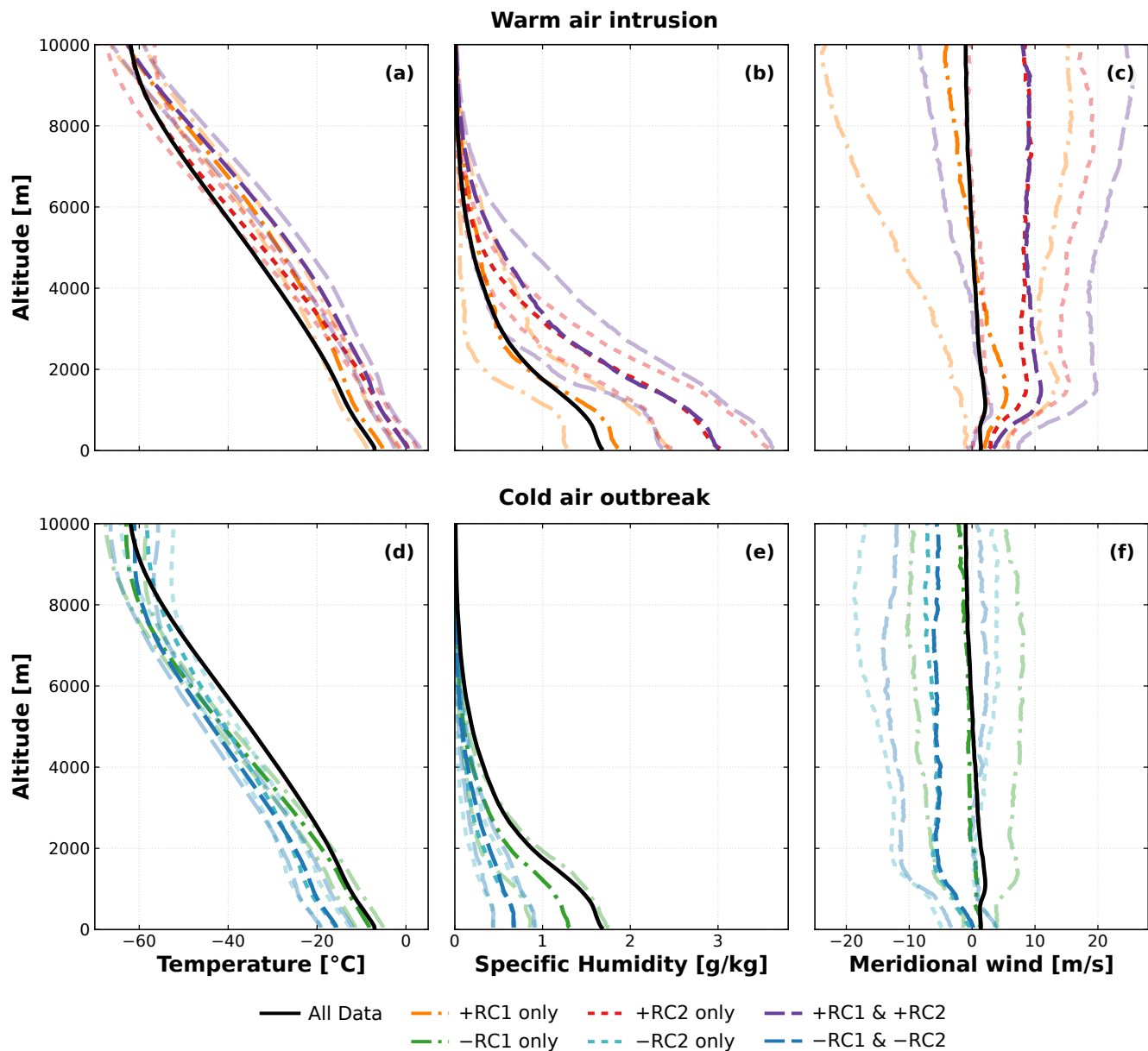


Figure 4. Composite mean vertical profiles of (a,d) temperature, (b,e) specific humidity, and (c,f) meridional wind for each rotated PCA-based subset during warm air intrusions (a–c) and cold air outbreaks (d–f). The black solid line denotes the wintertime mean across all soundings (2006/07–2024/25). Thick colored lines indicate the mean profile of each group, and thinner colored lines indicate ± 1 standard deviation.

This interpretation is also consistent with the idea of air mass modification over cold surfaces (Pithan et al., 2018) and with cases where warm intrusions remain elevated above the boundary layer (You et al., 2022).

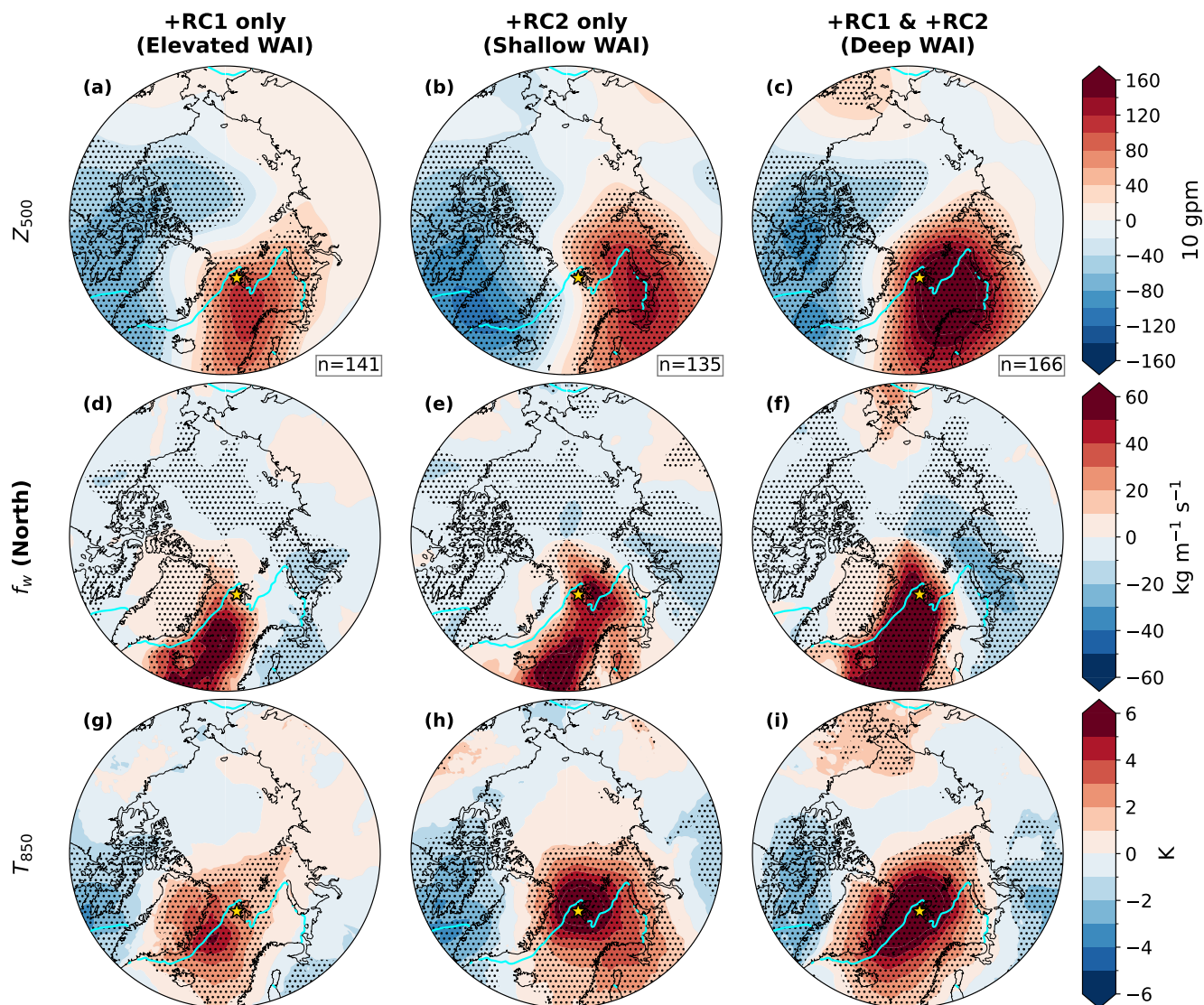


Figure 5. Composite ERA5 anomalies from the climatological mean associated with WAIs during 2006/07–2024/25 winters (DJF): (a–c) geopotential height at 500 hPa, Z_{500} (10 gpm); (d–f) northward water vapor flux, f_w ($\text{kg m}^{-1} \text{s}^{-1}$); and (g–i) temperature at 850 hPa, T_{850} (K). The cyan line indicates the climatological 15 % sea-ice concentration boundary calculated over the same years as the study period and is used to define the sea-ice edge. The star marks the study location, and stippling indicates statistical significance from a two-sided Student's t -test ($p < 0.01$).

In contrast, the +RC2-only group shows strong warming and moistening in the lower troposphere, together with enhanced southerly flow. This indicates an active WAI signal in which warm and moist air is being advected into the station region. However, the anomaly is more confined to the lower troposphere, so this group is interpreted as a shallow active WAI. The joint



+RC1 & +RC2 group combines the lower-tropospheric signal of RC2 with the deeper thermodynamic structure associated with RC1. It is therefore warmer and moister through a deeper part of the column and is interpreted as a deep active WAI.

290 The ERA5 composites in Fig. 5 support this interpretation. All positive-score regimes show a WAI-favorable circulation, with positive geopotential height anomalies over the Eurasian/Barents sector and lower geopotential height toward Greenland and the North Atlantic. This pattern is consistent with previous studies showing that blocking and cyclone activity can steer warm and moist air poleward into the Arctic (Murto et al., 2022; You et al., 2022). However, the pathway and strength of the transport differ among the regimes.

For +RC1-only, the geopotential height and moisture transport anomalies are displaced farther westward, and the warm 295 anomaly at 850 hPa is centered on Greenland's coast. This suggests that the air mass has followed a more sea-ice-influenced pathway before reaching the Ny-Ålesund station, allowing heat and moisture loss near the surface. This explains why the local profile remains warm aloft but shows a weaker lower-tropospheric response.

The joint +RC1 & +RC2 composite shows the clearest and deepest WAI signal. The northward moisture-transport anomaly extends from the North Atlantic toward Svalbard, and the 850 hPa temperature anomaly forms a broad warm tongue along 300 the same pathway. This indicates sustained warm and moist advection over relatively warm ocean surfaces, consistent with the strong lower-tropospheric warming and moistening and the warmer, moister middle troposphere in Fig. 4a,b. This group therefore represents the most vertically developed WAIs. The +RC2-only composite also shows strong lower-tropospheric transport and a positive 850 hPa temperature anomaly near Svalbard. However, compared with the joint group, the anomaly is shallower, and the mid-tropospheric thermodynamic signal is weaker. This suggests an active but more vertically confined 305 intrusion, possibly linked to a different or more regional pathway through the Barents/Scandinavian sector. Thus, RC2 identifies active lower-tropospheric WAI conditions, while the joint +RC1 & +RC2 group captures deeper and more vertically developed WAI events.

Together, the vertical profiles in Fig. 4a–c and ERA5 composites in Fig. 5 show that the RC-based classification separates not only the intensity of WAI conditions but also their vertical structure and likely transport history. The +RC2-only and joint 310 groups are characterized by strong lower-tropospheric warming, moistening, and southerly flow, indicating active warm and moist advection into Ny-Ålesund. The +RC1-only group, by contrast, is characterized by elevated warming and a weaker lower-tropospheric response, suggesting a relatively weak or decaying intrusion, or potentially the early stages of an intrusion event. The distinction between the +RC2-only and joint regimes appears to depend on the intrusion's depth and pathway. Both are active near the surface, but the joint group is associated with a more direct Atlantic pathway and a deeper thermodynamic 315 anomaly, whereas the +RC2-only group is more confined to the lower troposphere and may reflect a different air mass source.

3.1.2 Cold air outbreaks

The negative-score regimes show the cold and dry counterpart of the positive-score cases. All three groups are colder and drier than the wintertime mean as shown in Fig. 4d–f, but the strength and vertical extent of the anomalies again differ between RC1 and RC2. The –RC1-only group has a more elevated cold anomaly and a weaker lower-tropospheric response, while its 320 meridional wind profile remains close to the winter mean. This suggests a weak or modified CAO regime rather than an active

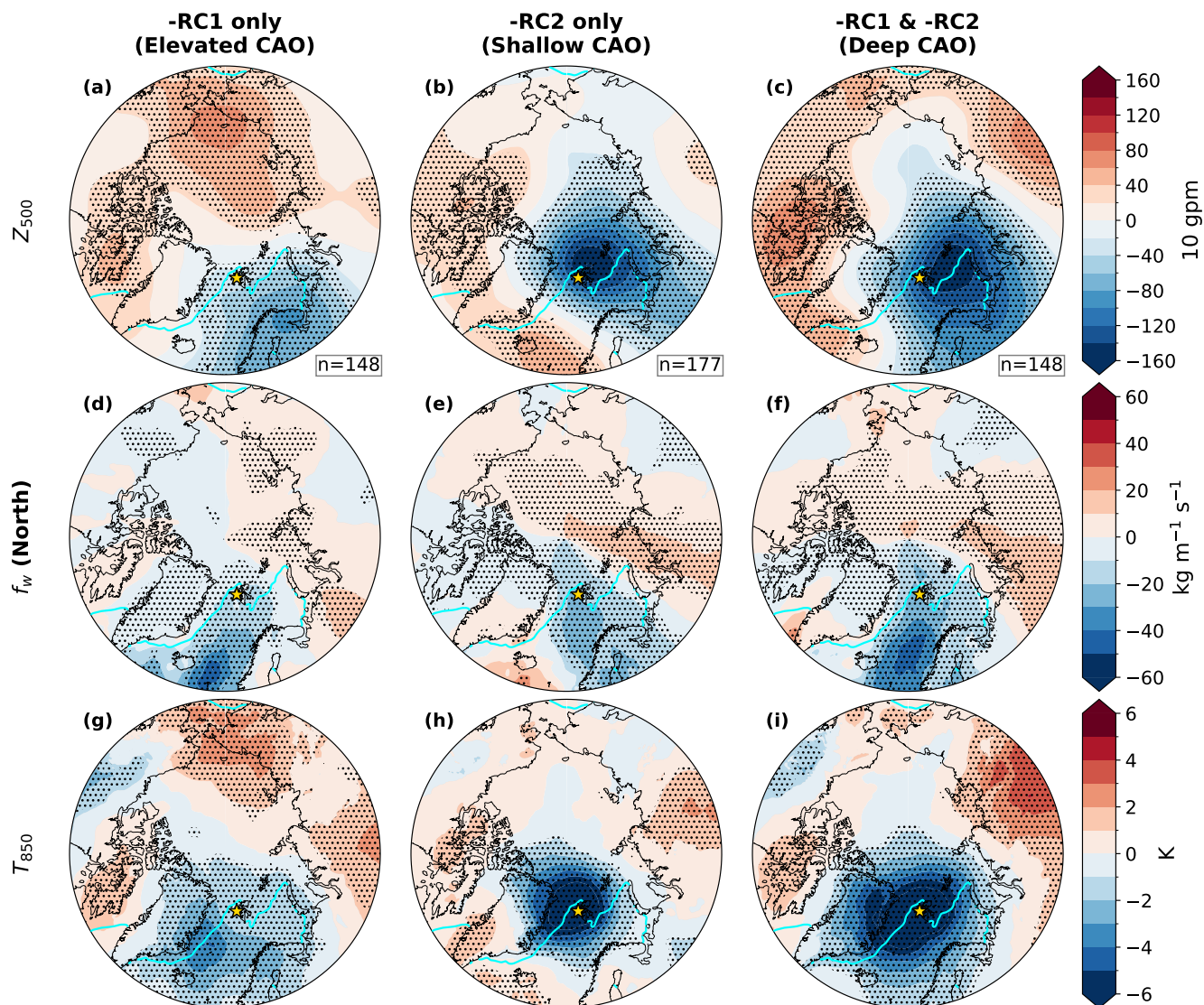


Figure 6. Composite ERA5 anomalies associated with CAOs during 2006/07–2024/25 winters (DJF): (a–c) geopotential height at 500 hPa, Z_{500} (10 gpm); (d–f) northward water vapor flux, f_w ($\text{kg m}^{-1} \text{s}^{-1}$); and (g–i) temperature at 850 hPa, T_{850} (K). The cyan line indicates the climatological 15 % sea-ice concentration boundary calculated over the same years as the study period and is used to define the sea-ice edge. The star marks the study location, and stippling indicates statistical significance from a two-sided Student’s t -test ($p < 0.01$).

outbreak at the station. By comparison, the –RC2-only group shows stronger lower-tropospheric cooling, drying, and northerly flow, indicating a more active cold-air advection signal. This group is therefore interpreted as a shallow CAO. The joint –RC1 & –RC2 group combines strong lower-tropospheric cooling with a deeper cold and dry anomaly, and is interpreted as a deep CAO. These results mirror the positive-score classification but with opposite-signed thermodynamic and dynamical anomalies.



325 The ERA5 composites in Fig. 6 show that the negative-score regimes are associated with a circulation anomaly pattern opposite to the WAI cases. They feature lower geopotential height over the Svalbard–Barents Sea sector and higher geopotential height toward Greenland and the Arctic. This configuration resembles the dipole pattern associated with North Atlantic marine CAOs, where ridging near Greenland and troughing toward Scandinavia promote northerly flow over the Norwegian and Barents seas (Afargan-Gerstman et al., 2020). Together with negative northward moisture-flux anomalies and negative 850 hPa
330 temperature anomalies near the station, this indicates cold and dry advection toward Ny-Ålesund.

Among the three negative-score regimes, the –RC1-only composite shows the weakest and most displaced signal. The 850 hPa cold anomaly is less centered on the station, and the moisture-flux anomaly is weaker than in the –RC2-only and joint groups. In the 500 hPa geopotential-height composite (Fig. 6a), the low-pressure anomaly is also displaced farther south relative to the –RC2-only and joint negative groups. Such a configuration may direct the air mass from the cold and dry
335 continental sector toward the marginal ice zone and the open-water boundary between the Barents and Kara seas before it reaches Svalbard. This would allow some modification through surface heat and moisture uptake, which could explain why the –RC1-only profiles are relatively warmer and moister by the time they arrive at Ny-Ålesund. Consistent with this interpretation, the meridional wind profile in Fig. 4f remains close to the winter mean, suggesting the absence of a pronounced northerly component and indicating a more zonal pathway. By contrast, the –RC2-only and joint –RC1 & –RC2 composites
340 show stronger cold, dry, and northerly signals near Svalbard. These regimes therefore capture the main thermodynamic and dynamical conditions associated with developing CAO events. The –RC2-only group represents a shallower cold outbreak focused in the lower troposphere, while the joint group represents a deeper and more vertically coherent CAO. Similar to the positive-score regimes, the negative-score regimes in Fig. 6d–f and Fig. 6 show that CAO characteristics are mainly tied to RC2, with the –RC2-only and joint –RC1 & –RC2 groups capturing the cold, dry, and northerly conditions required for
345 marine CAO development, while the joint classification identifies the deeper and more developed CAO regime (Pithan et al., 2018).

3.2 Cloud characteristics

Having established the main air-transformation regimes and their associated thermodynamic and dynamical structures, we next examine how these regimes are associated with multilayer cloud occurrence. For simplicity in the following discussion,
350 positive regimes are henceforth referred to as WAIs and negative regimes as CAOs. We assigned the RC groups physical names based on their vertical structure: RC1-only cases are classified as elevated WAI/CAO regimes, with thermodynamic anomalies concentrated aloft and weaker near the surface, consistent with an early or decaying stage; RC2-only cases are termed shallow active WAI/CAO regimes, characterized by the strongest thermodynamic anomalies near the surface; and joint RC1–RC2 cases are classified as deep active WAI/CAO regimes, since their thermodynamic anomalies extend over a greater
355 vertical depth. Because multilayer cloud formation depends closely on the vertical distribution of temperature, moisture, and dynamical forcing, its occurrence is expected to vary systematically across these distinct regimes.

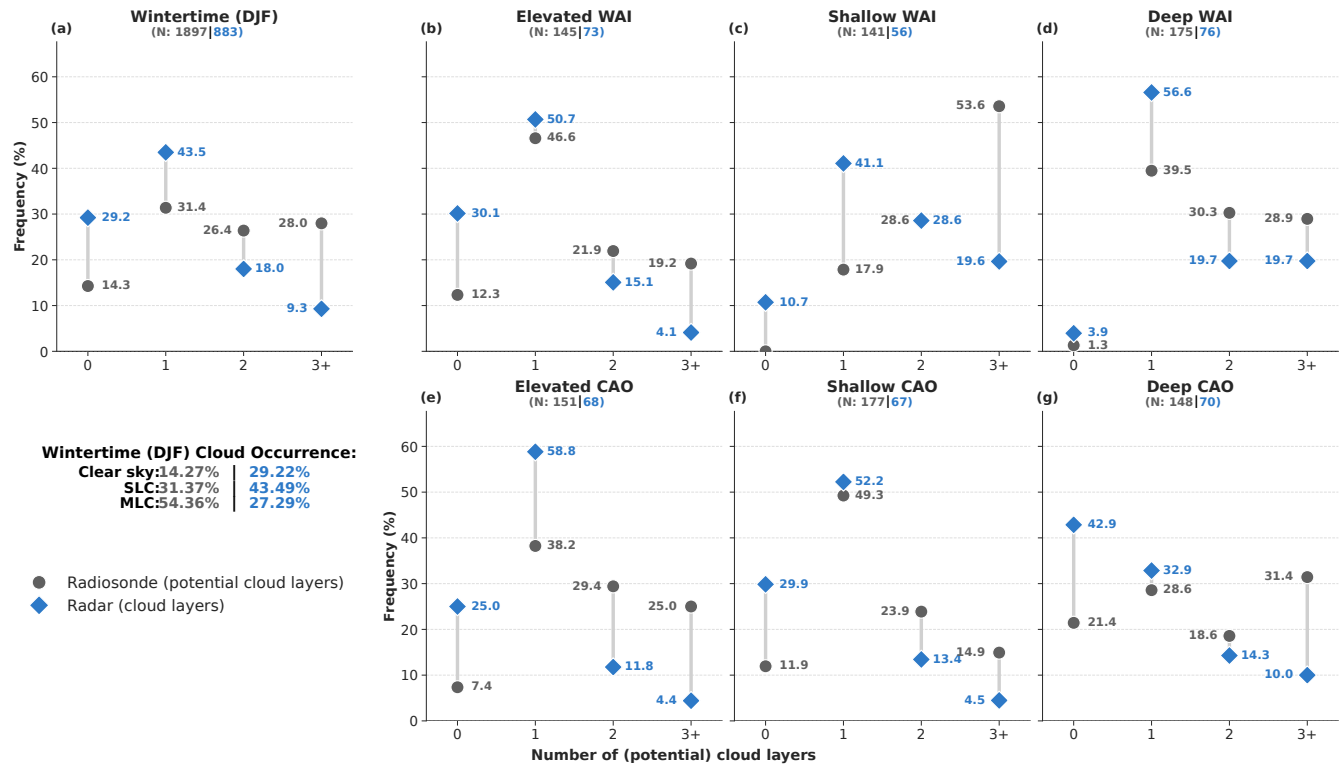


Figure 7. Frequency distribution of cloud layer occurrence during wintertime (DJF) and for each rotated PCA-based classification. The x-axis indicates the number of identified layers, where 0 Gray clear-sky and 3+ denotes three or more layers. Gray markers show potential cloud layers identified from radiosonde profiles (2006/07–2024/25), while blue markers show radar-verified (2017–2024/25) cloud layers. Numbers in the panel titles indicate the sample size for the radiosonde-only and radiosonde–radar classifications, respectively.

3.2.1 MLC occurrence

Previous studies have shown that winter clouds at Ny-Ålesund, Svalbard, are frequently organized in vertically complex structures. Using the Cloudnet target classification product, which is derived from the synergy of several ground-based remote-sensing instruments, Nomokonova et al. (2019) reported a total cloud occurrence of about 76.2 %, with multilayer clouds occurring 44.2 % of the time and single-layer clouds occurring 32.0 % of the time. Among the single-layer clouds, mixed-phase clouds were the most frequent, followed by ice and liquid clouds. Their monthly statistics further showed that single-layer ice clouds become more common during winter and spring, with occurrence values of about 15 –20 %, whereas single-layer liquid clouds are rare outside summer. Almost similar wintertime cloud statistics were obtained by Achtert et al. (2026), who combined radiosonde profiles with cloud radar from the CloudNet product. For Ny-Ålesund, their multi-year winter statistics indicate that cloud-free conditions account for roughly one-fourth of the radiosonde profiles, while single-layer clouds also occur in about one-fourth of the cases and multilayer clouds make up the remaining, slightly larger fraction of about 49 %.



Table 1. Wintertime (DJF) cloud occurrence comparison at Ny-Ålesund, Svalbard, across different studies. Active WAI (CAO) refers to the shallow and deep WAI (CAO) regimes.

Study / Regime	Observation & Period	Cloud Statistics		
		Clear sky	Single layer	Multilayer
Nomokonova et al. (2019)	Ground-based remote-sensing (2017)	~23.8%	~32.0%	~44.2%
Vassel et al. (2019)	Radiosonde and cloud radar (2017)	~37.7%	~28.0%	~34.3%
Achtert et al. (2026)	Radiosonde and cloud radar (2017–2022)	~24%	~27%	~49%
Liu et al. (2012)*	Satellite (2006–2011)	14–17% [†]	57–60% [†]	~26.0%
This study	Radiosonde (2006/07–2024/25)	14.3%	31.4%	54.4%
	Radiosonde and cloud radar (2017–2024/25)	29.2%	43.5%	27.3%
Active WAI [‡]	Radiosonde	0.6%	28.7%	70.7%
	Radiosonde and cloud radar	7.3%	48.9%	43.8%
Active CAO [‡]	Radiosonde	16.7%	39.0%	44.4%
	Radiosonde and cloud radar	36.4%	42.6%	21.1%

*Note: The spatial domain for Liu et al. (2012) is the Atlantic-side Arctic (60–82° N) and the rest is at Ny-Ålesund, Svalbard.

[†]Values calculated based on the study's reported total cloud occurrence (83–86%) and multilayer cloud fraction (26%).

Satellite observations also indicate cloudy winter conditions, but with lower MLC occurrence. Liu et al. (2012) reported winter total cloud occurrence values of about 83–86% in the Atlantic-side Arctic sectors between 60–82° N, whereas winter MLC occurrence was only around 26%. This lower satellite-based MLC fraction likely reflects observational limitations from space, since upper cloud layers can obscure lower layers and shallow or optically thin clouds near the surface may be missed (Schirmacher et al., 2023). The wintertime cloud occurrence statistics from these previous studies and the present work are synthesized for comparison in Table 1.

We retain both the radiosonde-only and radar-verified classifications because they provide complementary information and differ in temporal coverage. The radiosonde-only classification uses approximately two decades of observations starting in 2006, allowing a longer-term assessment of the thermodynamic conditions that can support cloud layer formation. This classification is interpreted as the occurrence of potential cloud layers, because it identifies vertically separated supersaturated layers in the radiosonde profiles—saturated layers do not necessarily mean that there are clouds present in that layer due to the scarcity of aerosol particles as cloud condensation or ice nucleation nuclei in the Arctic. The radar-verified classification is limited to almost one decade of overlapping radiosonde–radar observations, so a shorter period compared to the radiosonde-only classification. It provides a more direct constraint on which of the potential layers are associated with detectable hydrometeors. However, it is also important to note that ground-based radar observations may underestimate cloud occurrence in some situations, particularly for optically thin high-altitude clouds or clouds containing very small ice crystals with weak radar reflectivity that can fall below the instrument detection threshold (Wang et al., 2022). Additionally, thick low-level liquid or



precipitating clouds can attenuate the radar signal and reduce sensitivity to weaker cloud layers above, potentially leading to underdetection of overlying clouds. Together, the two classifications provide a range of wintertime cloud occurrence, where the radiosonde-only result describes the potential for layered cloud formation over a longer time period, whereas the radar-verified result gives a shorter-term but observationally more constrained estimate that is more directly comparable with previous radar-based cloud classifications (Nomokonova et al., 2019). Figure A4 shows how potential cloud layers identified from radiosonde supersaturated layers are redistributed after being verified by radar-detected cloud layers.

For all wintertime cases, the radiosonde-only classification gives 14.3 % clear-sky conditions, 31.4 % single-layer potential cloud cases, and 54.4 % multilayer potential cloud cases (Fig. 7a). Thus, more than half of the winter profiles contain two or more vertically separated supersaturated layers. This does not necessarily imply observed clouds, but rather indicates that the thermodynamic structure of the atmospheric column can support cloud formation if sufficient cloud condensation and ice-nucleating particles are present. In contrast, the radar-verified classification gives a different partitioning, with 29.6 % clear-sky conditions, 44.2 % single-layer cloud cases, and 26.3 % multilayer cloud cases. The substantially lower frequency of radar-verified multilayer clouds indicates that not every saturated layer identified in the radiosonde profile contains radar-detectable particles. This discrepancy is expected, as some saturated layers occur in a pristine Arctic environment where limited aerosol availability can inhibit cloud formation. Furthermore, observational challenges arise when clouds contain minimal condensate or consist of minute ice crystals below the radar's detection threshold (Wang et al., 2022), as well as when thick, precipitating layers attenuate the radar signal and obscure weaker cloud layers above. Therefore, the difference between the radiosonde-only and radar-verified classifications likely reflects both microphysical constraints on cloud development and observational limitations of the radar.

Taken together, the two classifications provide a physically meaningful range for wintertime cloudiness at the Arctic gateway. The radiosonde-only classification suggests that 85.7 % of winter profiles have the potential for cloud formation, whereas the radar-verified classification gives a detectable cloud occurrence of 70.5 %. Previous estimates for wintertime cloud occurrence at Ny-Ålesund fall within this range, as shown in Table 1. This agreement suggests that the combined radiosonde-only and radar-verified classifications bracket the plausible range of wintertime Arctic cloudiness for the period of analysis, while also highlighting the distinction between the potential for cloud formation from the thermodynamic structure and the observed cloud layers from the radar. These overall statistics provide the baseline for the regime-resolved analysis. We next examine how this cloudiness is distributed across the regimes identified in the previous section, and whether multilayer clouds occur preferentially under specific atmospheric conditions.

The regime dependence is clearest for shallow active WAIs. This regime shows the strongest MLC potential, with 82.2 % of the profiles containing two or more saturated layers and 53.6 % containing three or more layers (Fig. 7c). The radar-verified cloud statistics also show a nearly equal partitioning between SLC and MLC occurrence, with 44.6 % and 44.7 %, respectively. This indicates that active WAI conditions strongly favor vertically complex wintertime cloud structures, as hypothesized by Tan et al. (2023).

The active deep WAI case is almost always cloudy. Clear-sky occurrence is limited to only 3.9 %, while SLC and MLC occurrence reach 56.6 % and 39.4 %, respectively (Fig. 7d). Although total cloud occurrence is very high in this regime, the



larger SLC fraction suggests that the vertical cloud structure is less frequently multilayered than during an active shallow WAI. This indicates that the warmest and moistest regime does not necessarily provide the most favorable conditions for multilayer cloud formation. Instead, multilayer development appears to require not only enhanced moisture and warmth, but also a vertical thermodynamic structure that supports the formation of distinct, separated supersaturated layers. By contrast, decaying WAI conditions also show a frequent cloud occurrence but weaker multilayer development. In this regime, SLC occurrence reaches 50.7 %, whereas MLC occurrence is only 19.2 % (Fig. 7b). This suggests that residual WAI-related warmth and moisture can maintain cloudiness, but are less effective at sustaining multiple separated cloud layers.

The CAO regimes show a different response. The shallow active CAO regime is dominated by SLC occurrence of 53.7 %, while MLC occurrence is limited to 16.4 % (Fig. 7f). Although 38.8 % of the profiles still contain two or more saturated layers, the cloud field is mainly expressed as single-layer rather than multilayer. The deep active CAO regime, on the other hand, has a stronger suppression of total cloudiness. It has the highest clear-sky occurrence among all regimes, reaching 42.9 %, while SLC and MLC occurrence are 32.9 % and 24.3 %, respectively (Fig. 7g). The decaying/elevated CAO condition is not necessarily a CAO event, but it has a colder and drier thermodynamic signature compared to the wintertime mean. In this regime, SLC occurrence is high at 58.8 %, whereas MLC occurrence is only 14.7 % (Fig. 7e). This indicates that colder and drier conditions can still support frequent cloud formation, but mainly as SLCs rather than vertically complex multilayer systems.

The regime statistics show that WAI states are generally cloudier, with the strongest multilayer development occurring during active (shallow and deep) WAIs. In elevated WAIs, clouds remain frequent but are mainly single-layer. In contrast, the developing shallow CAO regime generally shows reduced multilayer occurrence, whereas deep CAOs feature enhanced clear-sky conditions, yet retain a higher potential for MLCs relative to other CAOs. Ultimately, the main contrast lies between the vertically complex cloud structures favored by active WAI conditions and the clear-sky or single-layer profiles predominant in colder, drier CAO states.

3.2.2 MLC vertical structure

We next examine the vertical distribution of cloud layers in each regime. Figure 8 shows the bases, tops, and median geometrical thicknesses of the first, second, and third layers, with radiosonde-only saturated layers shown on the left side of each violin and radar-verified cloud layers shown on the right.

For the wintertime mean, the radiosonde-only layers extend over a deeper part of the troposphere than the radar-verified clouds, especially for the second and third layers (Fig. 8a). The difference is largest at higher altitudes, indicating that many saturated layers not confirmed by radar occur above the lower troposphere. These elevated layers may be too high or contain too few ice particles, or particles that are too small, to produce detectable radar reflectivity. Additionally, receiver saturation from intense low-level echoes can also restrict the cloud radar's range.

The active WAI regimes show the clearest vertical extension. In the shallow WAI regime, the second and third layers reach higher altitudes and occupy a broader vertical range than in the wintertime mean (Fig. 8c). The radar-verified layers remain lower than the radiosonde-only layers, but they still show clear separation between lower and elevated cloud layers. This

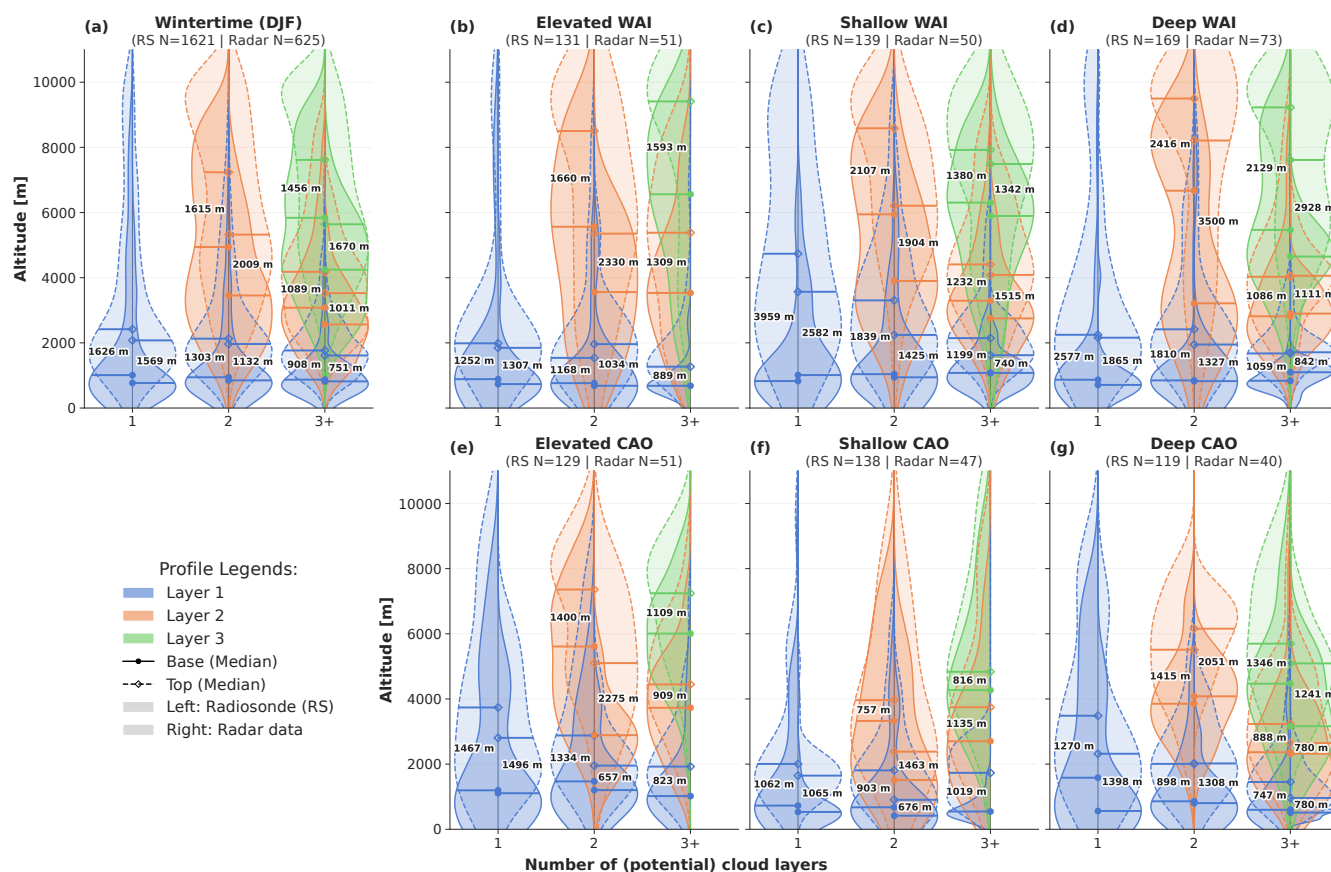


Figure 8. Vertical distribution of cloud layer bases (solid lines) and tops (dashed lines) for the wintertime mean and for each rotated PCA-based classification. The violin plots show the first, second, and third layers as a function of the total number of (potential) cloud layers in the profile. The left side of each violin shows radiosonde-only saturated layers, while the right side shows radar-verified cloud layers, with right-hand distributions omitted where sample sizes are insufficient. Solid and open symbols denote the median layer base and top, respectively. Numbers indicate the geometric mean layer thickness.

agrees with the high multilayer cloud occurrence in this regime. The single-layer cases are relatively deep as well, indicating that active warm and moist air intrusions can produce a vertically extensive cloud layer.

455 The deep WAI regime contains elevated and relatively thick second and third layers (Fig. 8d). The most evident difference is found in the radar-verified elevated layers, which are among the thickest in the regime composites and can exceed 2 km in depth. However, the occurrence statistics show that this regime is still more often classified as SLC than MLC. Thus, the very warm and moist state produces widespread and sometimes vertically deep cloudiness, but the saturated structure is not always separated into distinct radar-detectable layers. When multilayer clouds are detected, they tend to include a deep elevated cloud

460 layer.



The elevated WAI regime shows weaker vertical cloud development (Fig. 8b). Radar-verified clouds are mainly concentrated in the lower troposphere, and the second and third layers are less vertically extended than in the active shallow WAI regime. Layer thicknesses do not show the same strong contrast as in the deep WAI regime. This is consistent with a more decaying WAI state, where residual warmth and moisture maintain cloudiness but produce fewer elevated layers.

465 The CAO regimes exhibit a more vertically confined cloud structure. In the shallow CAO regime, radar-verified clouds are predominantly low-level and relatively shallow (Fig. 8f). While some elevated saturated layers remain in the radiosonde-only classification, they are rarely confirmed by the radar. The deep CAO regime similarly features mainly low-level radar-verified layers and fewer elevated structures (Fig. 8g), consistent with its larger overall clear-sky fraction. Where elevated radar-verified layers do occur, they tend to be relatively thick, possibly reflecting aerosol-limited cloud formation or weak radar signals from
470 thinner, sparsely populated cloud layers. The weak/decaying CAO regime shows a comparable pattern of frequent low-level clouds but limited radar-verified secondary or tertiary layers compared to the active WAI regimes (Fig. 8e).

The vertical distributions and layer thicknesses place the occurrence statistics in the context of the differing cloud structures across the regimes. Active WAI regimes, particularly the shallow WAI, favor deeper and more vertically separated cloud layers, consistent with their enhanced multilayer cloud occurrence. Deep WAI conditions are also highly cloudy, but when elevated
475 radar-verified layers occur they are often relatively thick, suggesting that only the more developed elevated cloud layers are consistently detected. In contrast, decaying WAI and CAO regimes are dominated by lower-level clouds, with fewer radar-verified secondary and tertiary layers and, particularly for deep CAO, a larger clear-sky fraction. The frequent occurrence of elevated radiosonde-only saturated layers indicates that saturation alone does not necessarily correspond to a radar-detectable cloud. This may reflect both radar detection limitations and differences in cloud microphysics. In particular, CAO air masses
480 may provide less favorable conditions for cloud particle formation because of their relatively pristine aerosol environment, whereas WAIs can transport moisture and aerosol-rich air into the Arctic (Dada et al., 2022).

3.2.3 Seeding potential

After examining where cloud layers occur, we next focus on potential seeding of lower cloud layers from upper cloud layers. The subsaturated gaps between cloud layers control whether ice particles falling from an upper cloud layer will reach and seed
485 the lower cloud layer. We evaluated the seeding potential for different initial aggregate ice-particle radii (see Fig. A5 for other ice-crystal habits) under the wintertime mean, WAI, and CAO regimes. For this analysis, we only consider the active (shallow and deep) WAI regimes and the developing (shallow and deep) CAO regimes. Only radar-verified cases are used, so that the analysis is restricted to profiles where cloud particles were actually observed. For completeness, the corresponding seeding potential plot for radiosonde-only cases is presented in Fig. A6.

490 As the seeding potential is known to depend strongly on ice-particle size (e.g., Vassel et al., 2019; Proske et al., 2021; Achtert et al., 2026), we evaluated the seeding classification for three initial particle radii: 100 μm , 200 μm and 400 μm (Fig. 9a). For an initial radius of 100 μm , non-seeding cases dominate across all regimes, with seeding occurring in only 24 % of cases, as the smaller particles generally sublime completely while descending through the intervening subsaturated layer. With increasing initial radius, seeding becomes more frequent, reaching 81 % of cases for an initial radius of 400 μm . The larger particles have

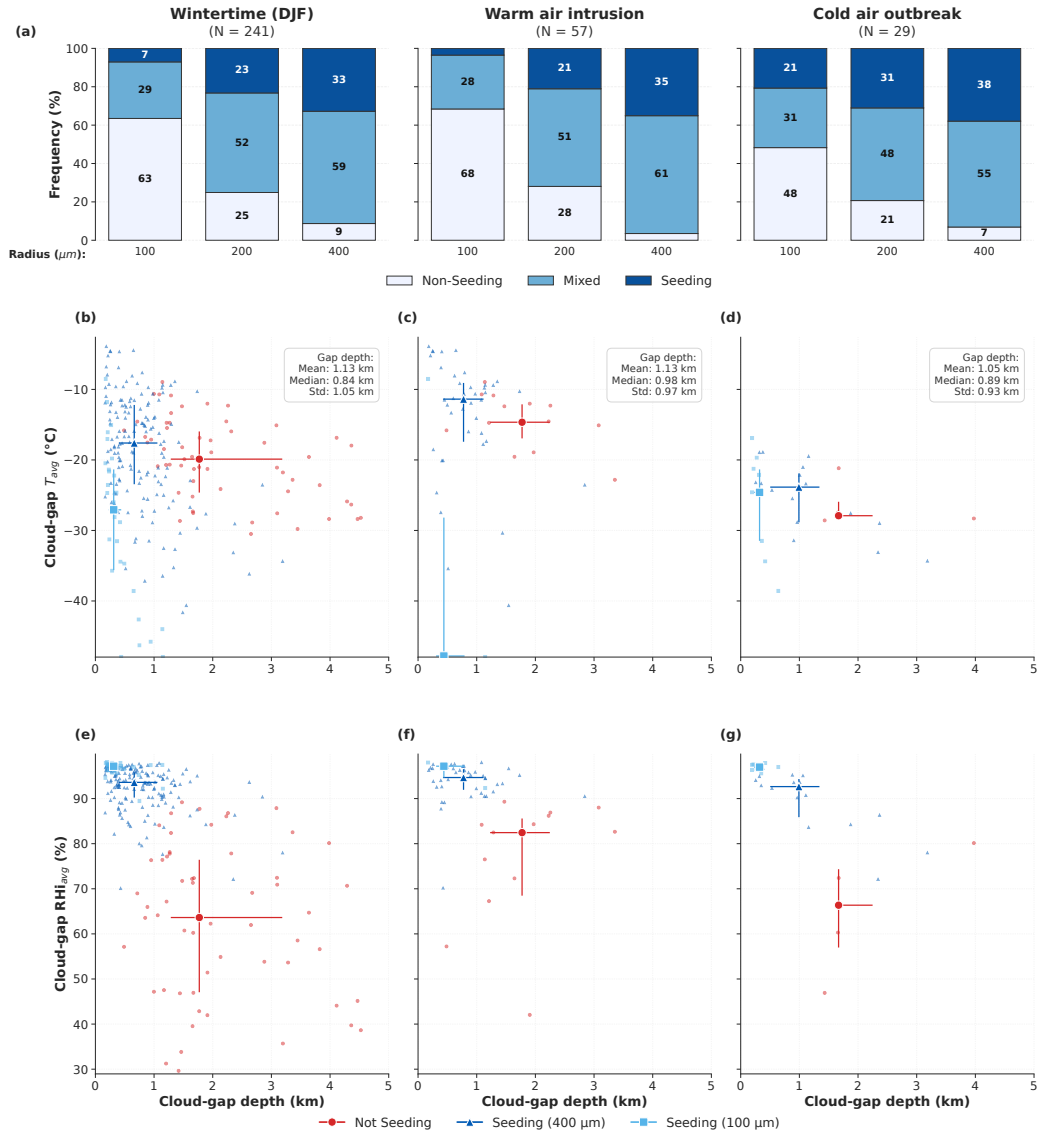


Figure 9. Seeding potential for radar-verified multilayer clouds during wintertime (DJF 2017-2024/25), warm air intrusion (WAI), and cold air outbreak (CAO) conditions. (a) Percent frequency of the seeding classification for three assumed initial aggregate ice-crystal radii, $r_0 = 100\mu\text{m}$, $200\mu\text{m}$ and $400\mu\text{m}$. Stacked bars show non-seeding, mixed, and seeding cases; the number of multilayer-cloud profiles (N) is given above each regime. (b–d) Cloud-gap depth versus mean cloud-gap temperature, T_{avg} , for (b) wintertime, (c) WAI, and (d) CAO conditions. (e–g) As in (b–d), but for mean cloud-gap relative humidity with respect to ice, $\text{RH}_{i,\text{avg}}$. Individual points represent cloud layer triplets consisting of an upper saturated layer, a subsaturated gap, and a lower saturated layer. Red circles indicate non-seeding cases, dark-blue triangles indicate seeding for $r_0 = 400\mu\text{m}$, and light-blue squares indicate seeding for $r_0 = 100\mu\text{m}$. Large symbols and whiskers show the median and interquartile range for each class. Insets in (b–d) give the mean, median, and standard deviation of the cloud-gap depth.



greater fall velocities and lose a smaller fraction of their mass during descent, increasing their likelihood of traversing the subsaturated layer and reaching the lower cloud layer, where they can initiate glaciation.

The regime dependence is most pronounced for the smallest particles due to the difference in cloud-gap temperature. Under WAI conditions, where the cloud-gap is relatively warmer, only a small fraction ($\sim 18\%$) of the $100\text{ }\mu\text{m}$ cases are classified as seeding, whereas the corresponding fraction is substantially larger ($\sim 41\%$) under CAO conditions (Fig. 9a). This divergence is consistent with the warmer subsaturated layers characteristic of the WAI regime (Fig. 9c, d), which promote more rapid sublimation of descending small ice crystals. Conversely, the colder CAO environment reduces sublimation and allows small particles to survive over a wider range of conditions. Although CAOs are generally characterized as cold, dry air because of their low specific humidity, this does not necessarily imply a greater departure from ice saturation, because the saturation vapor pressure over ice decreases strongly with temperature (Murphy and Koop, 2005). As shown in Fig. 9g and Fig. A6g, subsaturated layers during CAOs are also frequently close to ice saturation.

Accordingly, when the assumed initial radius was increased to $400\text{ }\mu\text{m}$, the seeding fractions converge and become nearly identical across the WAI, CAO, and wintertime mean statistics. This shift indicates that once particles are sufficiently large, variations in cloud-gap temperature become secondary for the overall seeding classification. Instead, Fig. 9b–g demonstrate that seeding is then primarily governed by the depth of the subsaturated gap and by how close that gap remains to ice saturation, which both act as the dominant controls for these larger particles. Typically, the dryness of the gap is directly proportional to its depth. Hence, seeding triplets are concentrated at short fall distances, with a median cloud-gap depth of 0.7 km to 1.0 km , compared with about 1.7 km for the non-seeding triplets, and most of them occur for gaps shallower than about 1.5 km . In addition, in both air mass regimes, gaps that remain close to ice saturation ($\text{RH}_{i,\text{avg}} \gtrsim 90\%$) are almost exclusively classified as seeding, largely independent of the gap depth over the observed range (Fig. 9e–g). In the colder CAO regime, this limit is relaxed, and seeding still occurs for gaps with $\text{RH}_{i,\text{avg}}$ down to about 80% , whereas under WAI conditions the non-seeding cases extend to nearly 90% and their median gap humidity ($\sim 82\%$) is markedly higher than during the wintertime mean and CAO conditions ($\sim 64\%$ to 66%). Across deeper gaps, non-seeding cases predominate even for $400\text{ }\mu\text{m}$ particles. Consequently, microphysical size alone is insufficient to guarantee cloud seeding, as large ice particles will still fully sublimate if the intervening subsaturated layer is dry ($\text{RH}_{i,\text{avg}}$ below about 90%) and deep (exceeding about 1 km).

Ultimately, this analysis reveals that multilayer cloud interactions through the seeder–feeder mechanism are dictated by a combination of microphysical and geometrical constraints. Small ice particles are strongly sensitive to the thermodynamic structure of the subsaturated layer, with the colder conditions associated with CAOs favoring their survival. In contrast, larger particles are more resistant to sublimation and are therefore more likely to reach and seed the lower cloud layer under both WAI and CAO conditions. This size dependence can be explained by their greater ice mass and lower surface-area-to-mass ratio, which reduce the fractional mass lost through sublimation. Larger particles also generally have higher fall speeds, shortening their residence time within the subsaturated layer and further increasing their likelihood of reaching the lower cloud layer. Importantly, this does not imply that the seeder–feeder process is inherently more efficient during CAOs. When placed in a broader physical context, a developing CAO typically produces few and relatively small ice crystals because of the clean air mass and the limited moisture available. In contrast, the enhanced aerosols and moisture availability during an active WAI

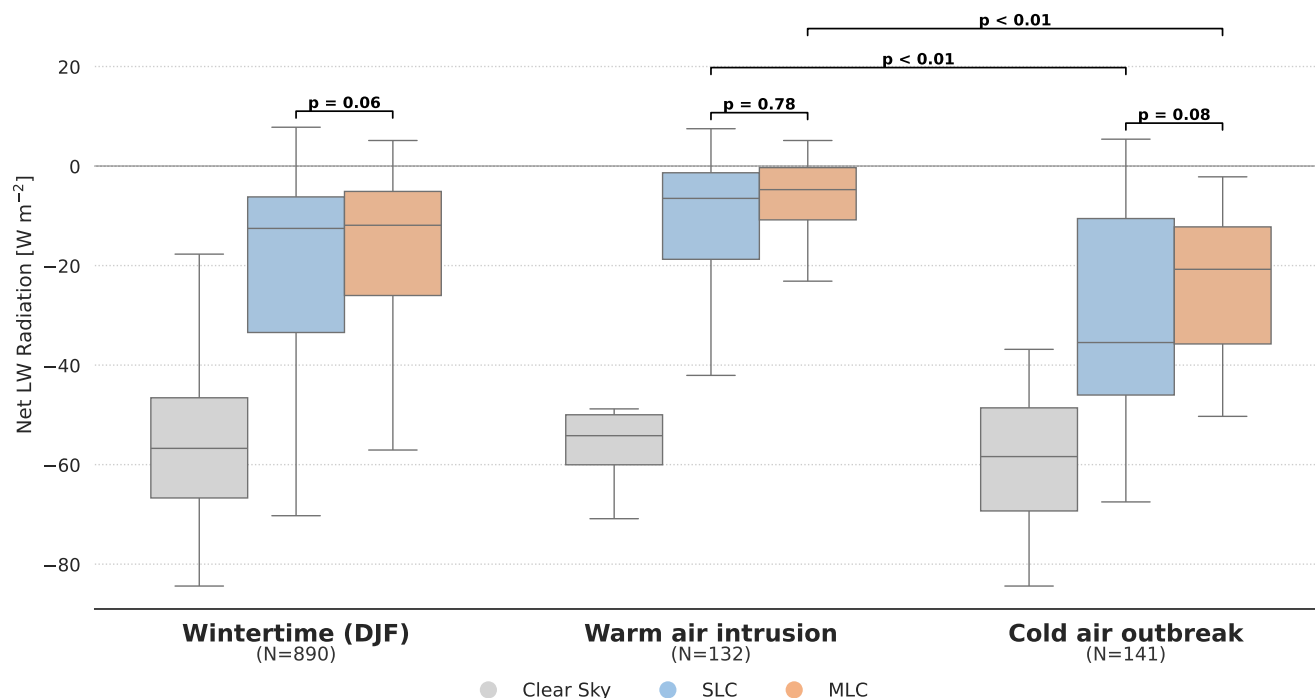


Figure 10. Distribution of surface net longwave (LW) radiation during wintertime (DJF 2017-2024/25), the WAI regime, and the CAO regime, separated by sky condition: clear-sky, single-layer cloud (SLC), and multilayer cloud (MLC). Negative values indicate net LW cooling at the surface. Boxes show the interquartile range, horizontal lines mark the median, whiskers extend to the most extreme data points that lie within 1.5 times the interquartile range of the quartiles, and data points outside the whiskers are omitted. Brackets indicate pairwise comparisons using Welch's two-sample *t*-tests, performed separately for SLC and MLC conditions within each regime and between the WAI and CAO regimes. Comparisons are labeled with their corresponding *p*-values.

530 foster the growth of more numerous and larger aggregate ice crystals (Dada et al., 2022). Because these larger particles are more resilient to sublimation and more likely to survive the fall, seeding events may therefore be more prevalent under active WAIs. Although the number of radar-verified cases is substantially smaller, in particular for the CAO regime ($N = 29$), a consistent behavior is also found for the potential seeding in the radiosonde-only dataset (Fig. A6).

3.2.4 Radiative impact

535 The distinct cloud vertical structures associated with different air mass transformation regimes influence the wintertime surface longwave radiative balance. Here, negative net longwave radiation represents an energy loss from the surface and therefore contributes to surface cooling. Clouds reduce this loss by absorbing surface-emitted radiation and emitting longwave radiation back toward the surface. This effect is particularly important in the Arctic, where the cold and dry clear sky atmosphere produces relatively weak downward longwave radiation. The magnitude of the cloud longwave effect depends on cloud tem-



perature, height, and emissivity, which are in turn influenced by cloud microphysical properties (Shupe and Intrieri, 2004). To examine how this response varies with cloud structure and air mass conditions, we compared clear-sky, SLC, and MLC conditions across all DJF observations and within the WAI and CAO subsets.

Figure 10 shows that clouds substantially reduce surface longwave cooling by almost 40 W m^{-2} . Across all DJF observations, clear-sky conditions exhibit the most-negative net longwave fluxes around -56.8 W m^{-2} , whereas both SLCs and MLCs shift the distributions toward zero. This clear-sky–cloudy contrast is consistent with observations at Ny-Ålesund showing that wintertime longwave fluxes increase with increasing cloud fraction and decreasing lowest cloud-base height (Yeo et al., 2018). MLCs generally show less-negative net longwave values than SLCs, with mean fluxes around -17.65 W m^{-2} for MLC and -20.46 W m^{-2} for SLC, although their distributions overlap considerably, with only weak statistical support for a difference in means ($p = 0.06$). Thus, cloud occurrence produces the dominant radiative contrast, whereas the additional difference associated with multilayer structure is more modest.

The vertical configurations shown in Fig. 8 may partly explain this tendency. MLC systems frequently contain a lowermost cloud layer close to the surface together with one or more elevated layers. The lowering of the lowest cloud base in MLC systems might result in slightly warmer cloud-base temperatures: averaged across all DJF observations, the lowest radar-detected cloud base sits at 1105 m with a mean temperature of -12.2°C in MLC systems, compared with 1293 m and -14.1°C in SLC systems, roughly 190 m lower yet nearly 2 K warmer. This could reduce the temperature difference between the cloud base and the surface and, consequently, weaken the net longwave cooling, as the enhanced downwelling longwave flux from the warmer cloud base may more closely compensate the surface upward longwave flux. Thus, the weaker cooling during MLC conditions may be related partly to the altitude and temperature of the lowest cloud layer, rather than to the number of cloud layers alone. However, this relationship likely also depends on the surface temperature and the degree of coupling between the cloud layer and the surface.

The weakest surface cooling occurs during WAIs. Even under clear-sky conditions, net longwave radiation tends to be less negative (on average -52.12 W m^{-2}) compared with all DJF observations, consistent with the warmer and moister atmospheric column associated with these events. Moist anomalies at Ny-Ålesund have previously been linked to atmospheric moisture transport from the North Atlantic, increased cloud occurrence and cloud water content, and an enhanced longwave cloud radiative effect outside summer (Nomokonova et al., 2020). Under cloudy WAI conditions, SLCs and MLCs both shift toward near-zero values and show substantial overlap, with no significant difference in mean net longwave radiation ($p = 0.78$). Downward longwave radiation during winter moisture intrusions is strongly controlled by cloud temperature and optical depth (Dimitrelos et al., 2023), while the longwave effect of liquid-containing clouds becomes less sensitive to additional liquid water as liquid water path (LWP) increases and the cloud becomes sufficiently emissive (Shupe and Intrieri, 2004; Chen et al., 2006; Ebell et al., 2020). Thus, one warm and sufficiently emissive cloud layer may already provide much of the reduction in surface longwave cooling during WAIs, leaving only a small additional effect associated with multilayer structure, which can be seen in Fig. A7a. Because LWP and optical depth are not explicitly examined here, this interpretation remains tentative.

In contrast, CAOs are characterized by stronger surface longwave cooling. Clear-sky values are strongly negative, consistent with weaker downward longwave emission from the colder and drier atmospheric column. Clouds still substantially reduce



575 this heat loss, but cloudy CAO cases remain more negative than their WAI counterparts. Similar contrasts between warm- and cold-air advection have been observed at Ny-Ålesund, where warm-air periods were associated with increased cloudiness and downward longwave radiation, whereas cold-air periods were characterized by reduced cloudiness and lower optical radiance (Yeo et al., 2018; Nomokonova et al., 2020). The separation between the SLC and MLC distributions appears larger during CAOs, with MLCs tending to produce weaker cooling. However, the distributions remain broad and overlapping, and the within-regime difference is not statistically significant due to small sample size. The apparent CAO contrast should therefore be interpreted as a tendency rather than a robust separation between cloud categories. Nevertheless, this tendency makes the specific radiative behaviors of single- versus multi-layer clouds during CAO a compelling avenue for future investigation.

Overall, the surface longwave response reflects the combined influence of the atmospheric thermodynamic state and the macrophysical and microphysical properties of the cloud system. WAIs shift the radiative balance toward weaker cooling, whereas CAOs favor stronger and more variable surface heat loss. The additional relationships presented in Appendix Fig. A7 show that less-negative net longwave fluxes are generally associated with warmer and lower cloud bases, particularly during WAIs. During CAOs, however, clouds with similar bottom temperatures and base altitudes produce a broader range of radiative responses. This variability may partly reflect the occurrence of optically thin, low-liquid-water clouds, for which the longwave effect is highly sensitive to small changes in LWP below approximately 30 g m^{-2} (Shupe and Intrieri, 2004). Observations at Ny-Ålesund similarly identify LWP as a primary control on the longwave radiative effect of liquid-containing clouds, with decreasing sensitivity as clouds become sufficiently opaque (Ebell et al., 2020). In MLC systems, the presence of additional cloud layers may increase the total column LWP, thereby enhancing the downward longwave radiation reaching the surface. At the same time, elevated layers can influence the surface response indirectly by increasing the longwave radiation received at the top of the lowest cloud layer. This additional radiative input reduces cloud-top radiative cooling and may modify the turbulence, microphysical structure, and evolution of the lower cloud (Chen et al., 2020). Thus, the greater radiative variability of CAO MLCs may reflect both the increased condensate associated with multiple layers and the interaction between the lowest cloud and the overlying layers, rather than the number of detected cloud layers alone.

4 Conclusions

This study used almost two decades of wintertime (2006/07–2024/2025 DJF) radiosonde observations from Ny-Ålesund, Svalbard, to examine how multilayer cloud occurrence and structure vary during air mass transformations at an Atlantic-sector Arctic gateway. The thermodynamic state of the atmosphere was classified using principal component analysis followed by varimax rotation of equivalent potential temperature profiles from GRUAN-processed radiosondes. The first two rotated components primarily reflect middle- and lower-tropospheric temperature–moisture variability, respectively, with positive (negative) index values indicating warming and moistening (cooling and drying). Positive RC2 regimes are associated with lower-tropospheric warming, moistening, and southerly flow, consistent with active (shallow and deep) WAI events, whereas negative RC2 regimes exhibit colder, drier conditions and northerly flow, consistent with developing (shallow and deep) CAOs. These interpretations are supported by ERA5 synoptic-scale composites showing poleward moisture transport during positive regimes and cold,



dry northerly advection during negative regimes. To systematically evaluate MLC characteristics, we use two cloud classification frameworks following Vassel et al. (2019). The radiosonde-only approach identifies saturated environments where clouds
610 can potentially form, while the radar-verified approach uses cloud radar reflectivity to confirm radar-detectable hydrometeors within the radiosonde-identified potential cloud layers.

Using the radiosonde-only classification, we find that wintertime multilayer cloud potential is common at Ny-Ålesund, similar to results from previous studies (e.g., Nomokonova et al., 2019; Achtert et al., 2026; Vassel et al., 2019). For all winter profiles, 14.3 % are classified as clear-sky, 31.4 % as single-layer potential cloud cases, and 54.4 % as multilayer potential
615 cloud cases. When the radiosonde-derived layers are constrained with cloud radar reflectivity, the cloud occurrence changes substantially: clear-sky conditions increase to 29.2 %, single-layer clouds account for 43.5 %, and radar-verified multilayer clouds occur in 27.3 % of cases. This prominent difference highlights that humidity-derived layers often capture thermodynamic cloud potential rather than physical, radar-detectable hydrometeors. This gap may be attributed by both microphysical and instrument limitations, where clean, aerosol-limited conditions of the Arctic restrict cloud droplet activation in supersat-
620 urated air, while the cloud radar itself is limited by signal attenuation and the presence of optically small hydrometeors. This interpretation is strongly supported by the vertical distribution of these discrepancies, which shows that undetected cloud layers are most prevalent at higher altitudes. The discrepancy between the two methodologies is largest in the upper troposphere, where low aerosol concentrations may restrict hydrometeor formation, while radar sensitivity decreases with range and may also be affected by attenuation from lower-level clouds. An overestimation of cloud layer occurrence at high altitudes is also
625 seen in the model studies by Wallentin et al. (2026a).

The vertical structure of multilayer clouds is found to depend heavily on the prevailing air mass regime. The shallow WAI regime shows the most pronounced multilayer development, with more than 80 % of the profiles containing two or more saturated layers. Radar-verified clouds within this regime are also frequently multilayered, indicating that active warm and moist advection is associated with vertically complex cloud structures. Meanwhile, the deep WAI regime is almost always
630 cloudy and frequently features deep, elevated cloud layers, yet it is more often classified as single-layer than multilayer. This indicates that the warmest and most humid atmospheric conditions do not necessarily maximize multilayer occurrence. In contrast, CAO events predominantly favor clear-sky conditions or single-layer, low-level clouds.

However, while multilayer cloud formation is less frequent during CAOs, when these events do occur, their environment is highly favorable for the seeder-feeder mechanism. To evaluate these regime-dependent interactions, we investigated this
635 seeder-feeder process by analyzing the subsaturated gaps within multilayer clouds and simulating the survival of descending ice crystals with varying initial radii and crystal habits. This seeding analysis shows that the microphysical interaction between cloud layers is closely controlled by both particle size and the vertical extent of the subsaturated gap. Small ice particles with a radius of 100 μm are more likely to survive within CAO conditions but typically sublime during WAIs due to the different thermodynamic conditions of the subsaturated layer. For larger particles of 400 μm , this regime-dependent difference weakens,
640 and the physical fall distance between cloud layers becomes the dominant control. When the gap between cloud layers exceeds around a kilometer, seeding becomes unlikely even for large particles, demonstrating that the vertical separation of cloud layers is a primary constraint on seeder-feeder efficiency regardless of prevailing air mass regime.



Finally, we quantified the surface radiative impact of these different cloud systems. We found that clouds substantially reduce wintertime surface longwave cooling, but the impact of multiple layers depends heavily on the prevailing thermodynamic regime. During WAI events, the strong suppression of surface heat loss is driven primarily by cloud occurrence. Because the atmospheric column is relatively warm and moist, a single emissive cloud layer is typically sufficient to bring the net longwave fluxes close to zero, rendering the additional warming from elevated layers minimal. Conversely, the colder and drier atmosphere during CAOs leads to much stronger clear-sky surface cooling. In this regime, the difference between single-layer and multilayer clouds becomes more noticeable, with multilayer systems tending to have stronger surface insulation. However, the high variability observed during CAO events indicates that the surface radiative response relies on more than just the number of cloud layers. Instead, it likely reflects the specific microphysical properties of the lowest cloud layer and its complex radiative interactions with the overlying clouds. Surface flux measurements alone cannot distinguish whether an elevated layer is radiatively decoupled from the low-level cloud or actively modifying its evolution and lifetime, an ambiguity the comparable radiative effect of single- and multilayer clouds during WAI illustrates directly. Closing this gap will require synergistic ground-based remote sensing to retrieve cloud microphysical properties such as phase partitioning, particle size, and liquid and ice water content, together with numerical simulations of MLCs to resolve inter-layer radiative interactions directly.

In conclusion, our results show that air mass transformation controls not only whether clouds form at Ny-Ålesund, but also whether they organize into vertically separated multilayer systems, whether seeding between layers is possible, and how strongly the cloud modifies wintertime surface longwave cooling. Radiosondes were used to reveal the thermodynamic structures that can support cloud formation, while radar observations identify which of those potential layers contain detectable hydrometeors. Radar-verified statistics show roughly twice the occurrence of clear-sky conditions as the radiosonde-only analysis. Narrowing this gap with additional observations would be a valuable direction for future work but lies beyond the scope of this study. Nonetheless, together the two records show that active WAI events favor vertically complex cloud structures. In contrast, CAOs more often produce low-level single-layer clouds or clear-sky conditions. While both single-layer and multilayer clouds reduce surface cooling across all conditions, the difference between these systems becomes more pronounced during CAOs, where multilayer clouds exert a stronger radiative influence, about 7 W m^{-2} more than single-layer clouds. In WAIs, by contrast, cloudy conditions showed no significant difference ($p = 0.78$). Our results suggest that this radiative signature is therefore not simply a function of how many layers a cloud system contains, but instead reflects the specific microphysical state of the lowest cloud layer and its coupling to the layers above. This dependence appears strongest under CAOs, where clouds are present in the regime where longwave sensitivity to liquid water path is highest, unlike the radiatively saturated clouds typical of WAIs.

Code and data availability. The multilayer cloud classification algorithm is available at https://github.com/maikenv/Classification_algorithm_of_multilayer_clouds.git (Vassel, 2019), and the modified version used in this study is available at https://github.com/llvillanueva/MLC_classification_algorithm/tree/main. Radiosonde data from Ny-Ålesund, including GRUAN RS92 and RS41 products, were obtained from



Sommer et al. (2012), Maturilli (2017), and GRUAN Ny-Ålesund site (<https://www.gruan.org/network/sites/ny-aalesund>). Cloudnet-processed remote-sensing products for Ny-Ålesund used in this study are publicly available from the ACTRIS Cloudnet data portal (<https://cloudnet.fmi.fi/site/ny-alesund>). Surface radiation measurements were obtained from PANGAEA. ERA5 reanalysis data were obtained from the Copernicus Climate Data Store (<https://cds.climate.copernicus.eu/>).

680 **Appendix A: Discontinuity in cloud occurrence statistics from the homogenized radiosonde record around 2006**

The multilayer cloud detection algorithm adopted from Vassel et al. (2019) was applied to the homogenized Ny-Ålesund radiosonde record by Maturilli and Kayser (2016) to examine the long-term behavior of clear-sky, single-layer cloud (SLC), and multilayer cloud (MLC) occurrence. The resulting annual frequencies in Fig. A1 show a marked discontinuity around 2006. Before 2006, clear-sky conditions occur much more frequently, while MLC occurrence is substantially lower. After 2006, the
685 record becomes more stable and is characterized by lower clear-sky occurrence and higher MLC occurrence. Since this change is very abrupt, we investigated a correlation with the radiosonde type used.

The timing of the discontinuity closely matches the radiosonde instrumentation history at Ny-Ålesund. The early part of the record is based mainly on Vaisala RS80-A and RS90 radiosondes, whereas from May 2006 onward the record uses RS92 and, subsequently, RS41 radiosondes. The transition from RS90 to RS92 in 2006 coincides with the sharp change in the derived
690 cloud occurrence statistics. This is particularly relevant because the more recent RS92 and RS41 observations have been reprocessed within the GRUAN framework, which applies advanced corrections and quality-control procedures to improve the accuracy and long-term consistency of humidity measurements.

The comparison of relative-humidity profiles in Fig. A2 further supports this interpretation. In the RS92/RS41 period, 4696 out of 6969 profiles, or about 67%, contain at least one level reaching the 95 % relative-humidity threshold. In contrast, during
695 the earlier RS80-A/RS90 period, only 1162 out of 5049 profiles, or about 23 %, reach this threshold. This much lower fraction of near-saturated profiles in the non-GRUAN period suggests a systematic dry bias in the earlier radiosonde record. Since the MLC algorithm depends directly on whether vertically resolved humidity layers exceed the saturation threshold, this bias would reduce the number of detected cloudy layers and could artificially increase clear-sky occurrence. It may also convert profiles that would otherwise be classified as MLC into SLC, or SLC into clear-sky conditions.

700 Previous studies have documented several humidity-related limitations of older radiosonde measurements, including reduced reliability at low temperatures, sensor time-lag effects, icing, and radiation-related dry bias (Treffeisen et al., 2007; Wang et al., 2002; Miloshevich et al., 2004). These effects are especially problematic for Arctic cloud detection because many cloud layers occur close to saturation or ice saturation. The GRUAN-processed RS92 and RS41 products are specifically designed to correct known systematic errors and provide more traceable humidity profiles with uncertainty estimates (Dirksen et al., 2014; Sommer
705 et al., 2023). In contrast, the earlier homogenized RS80-A/RS90 period does not provide the same level of vertically resolved, GRUAN-based correction.

For this reason, the pre-2006 part of the homogenized radiosonde record is not used in the main analysis. Although the early record is valuable for temperature and humidity climatology, the impact of the remaining humidity uncertainty on cloud

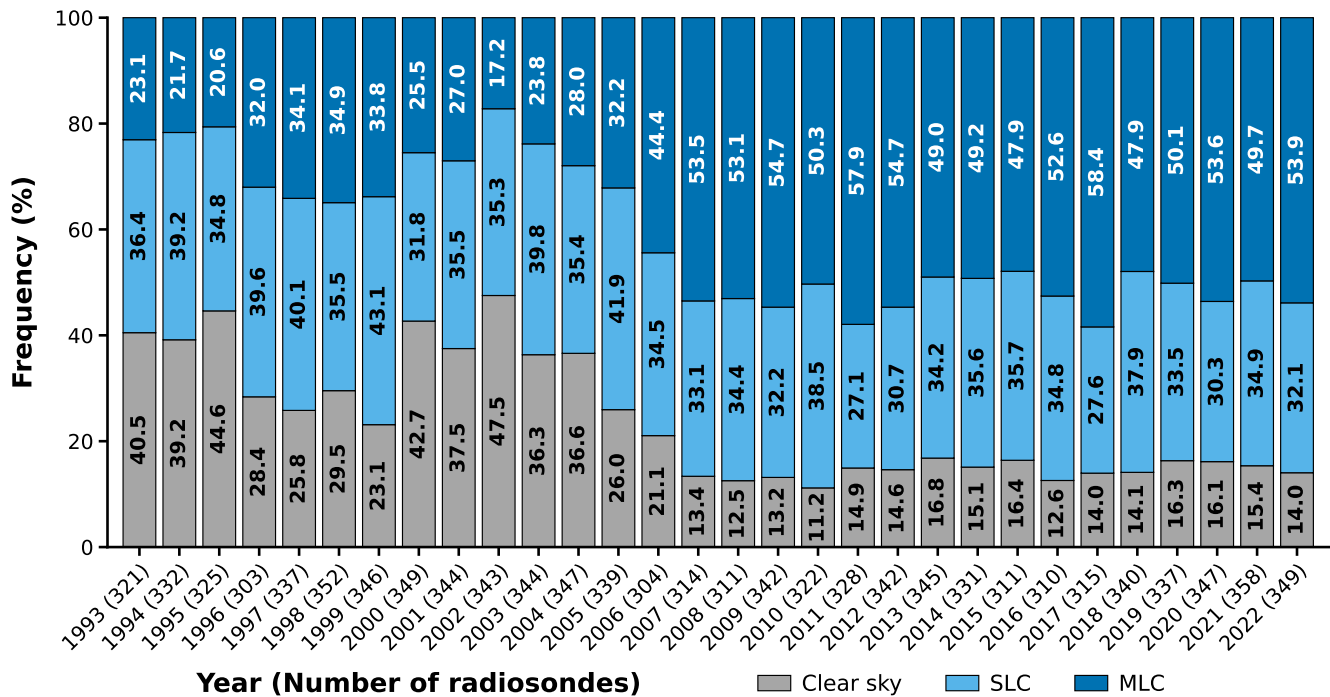


Figure A1. Interannual variability in cloud layer frequencies identified using the MLC algorithm. Stacked bars represent the annual relative frequency (%) of clear sky (grey), single layer cloud (SLC, light blue), and multilayer cloud (MLC, ≥ 2 layers, dark blue) from 1993 to 2022. Values overlaid on the bars denote the specific percentage for that category in a given year.

layer detection is difficult to quantify because the vertical distribution of the residual dry bias is unknown. Including this period would therefore risk introducing an artificial long-term change in cloud layer occurrence. We therefore restrict the main analysis to the GRUAN post-processed RS92/RS41 period, during which no significant long-term changes in potential cloud layering or overall potential cloudiness are detected.

Author contributions. LV and LI developed the idea of this study. LV, LI and HC developed the methodology. LV conducted the analysis and wrote the manuscript draft, under the supervision of LI and HC. All authors contributed to the discussion and interpretation of the results and commented on the manuscript.

Competing interests. The authors declare no competing interests. LI is a member of the Editorial Board of Atmospheric Chemistry and Physics.

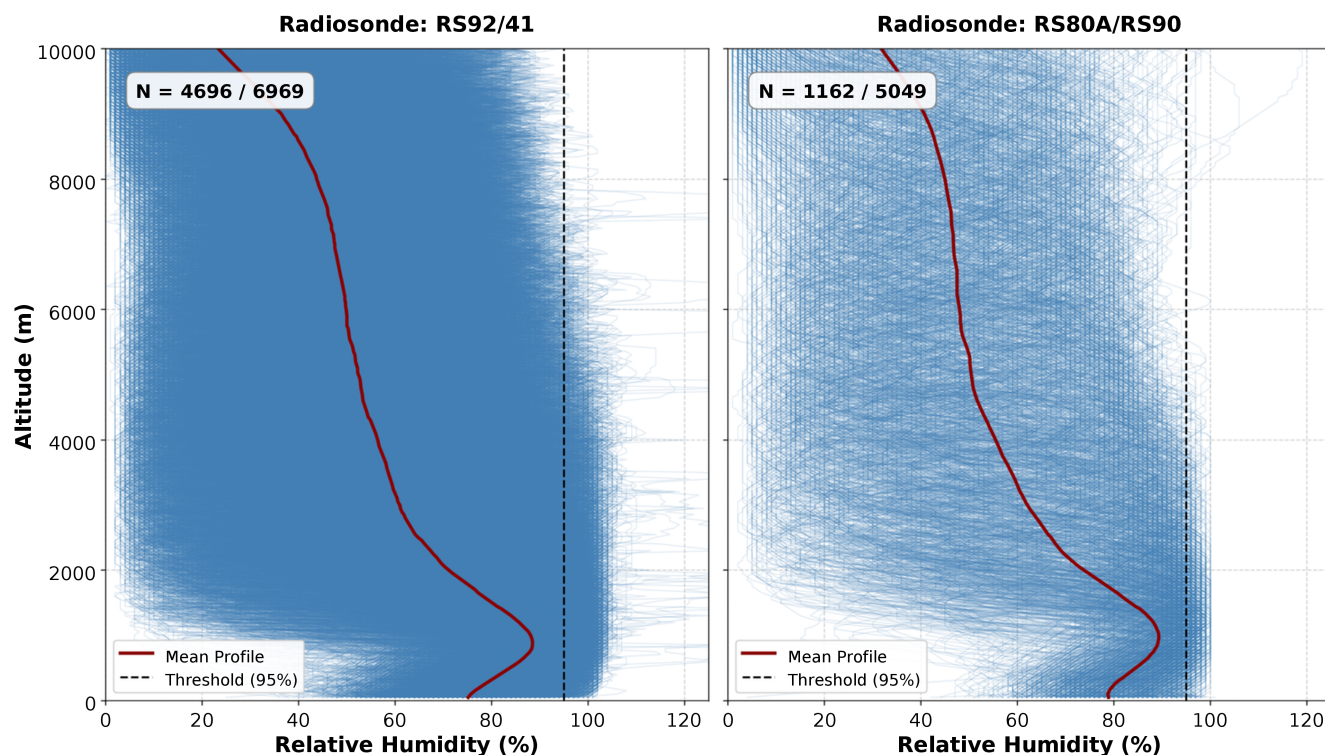


Figure A2. Vertical profiles of relative humidity for RS92/41 (left) and RS80A/RS90 (right) radiosonde. The data are filtered to include only profiles containing at least one measurement $\geq 95\%$ relative humidity, indicated by the vertical dashed black line. Individual sounding profiles are represented by faint blue lines, while the composite mean profile for each subset is denoted by the thick dark red line. The text boxes in the upper left of each panel indicate the number of profiles meeting this condition out of the total number of available profiles (N) for that respective sensors category.

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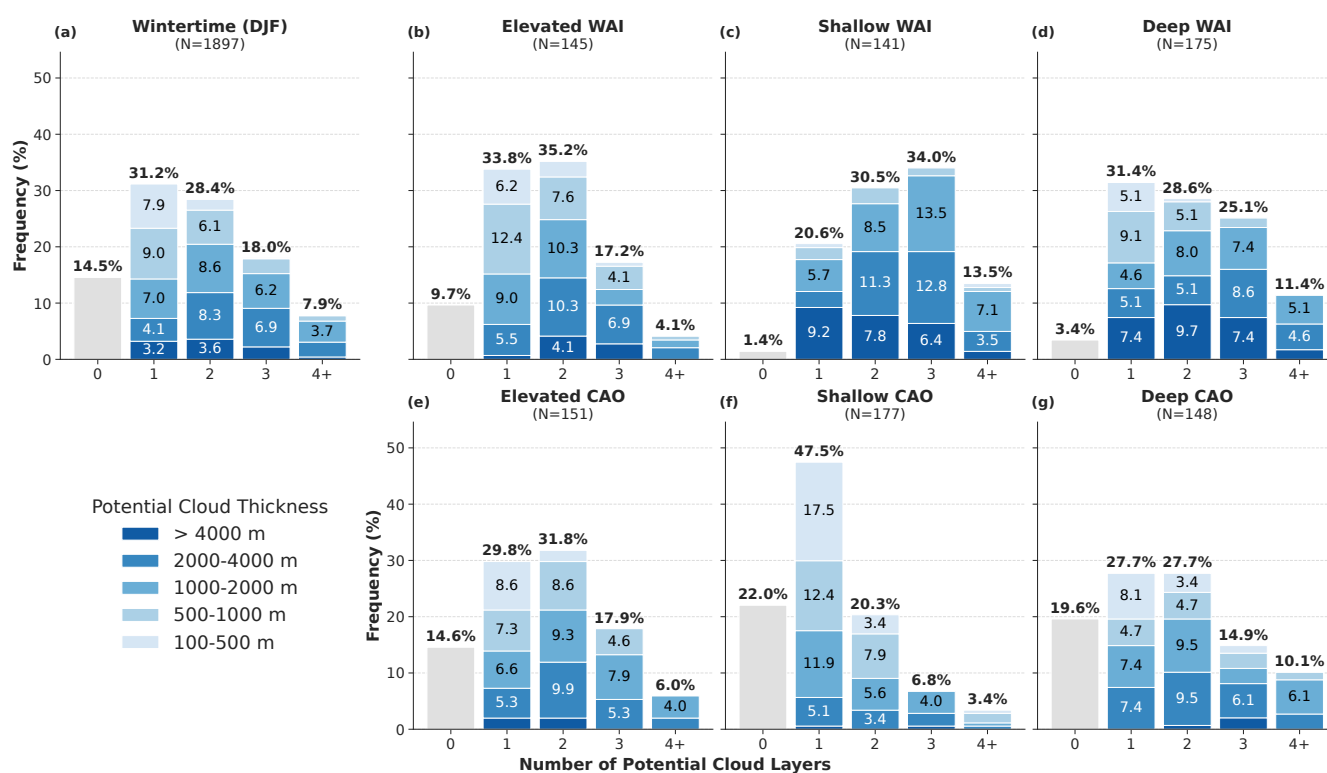


Figure A3. Frequency distribution of the number of potential cloud layers during winter and within each air mass transformation regime. Bars show the relative frequency of profiles with 0, 1, 2, 3, or 4+ potential cloud layers, and colors indicate the thickness category of the detected layers. Bold values above each bar indicate the total frequency for each cloud layer category, while numbers within the bars show the contribution from each layer-thickness category.

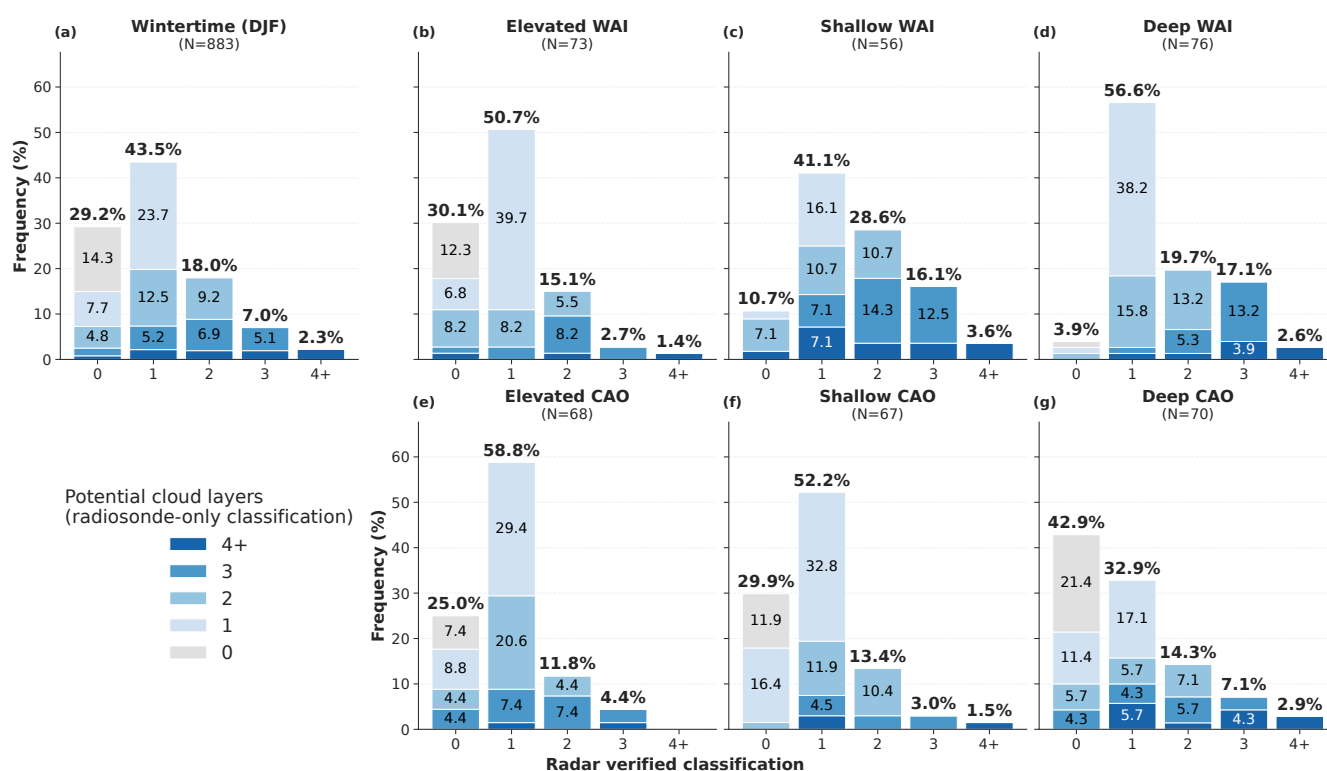


Figure A4. Frequency distribution of radar-verified cloud layers during winter and within each air mass transformation regime. Bars show the relative frequency of profiles with 0, 1, 2, 3, or 4+ cloud layers, while colors indicate the corresponding number of potential cloud layers identified from radiosonde supersaturated layers before radar constraint. Bold values above each bar indicate the total frequency for each cloud layer category, and numbers within the bars show the contribution from each potential-layer category.

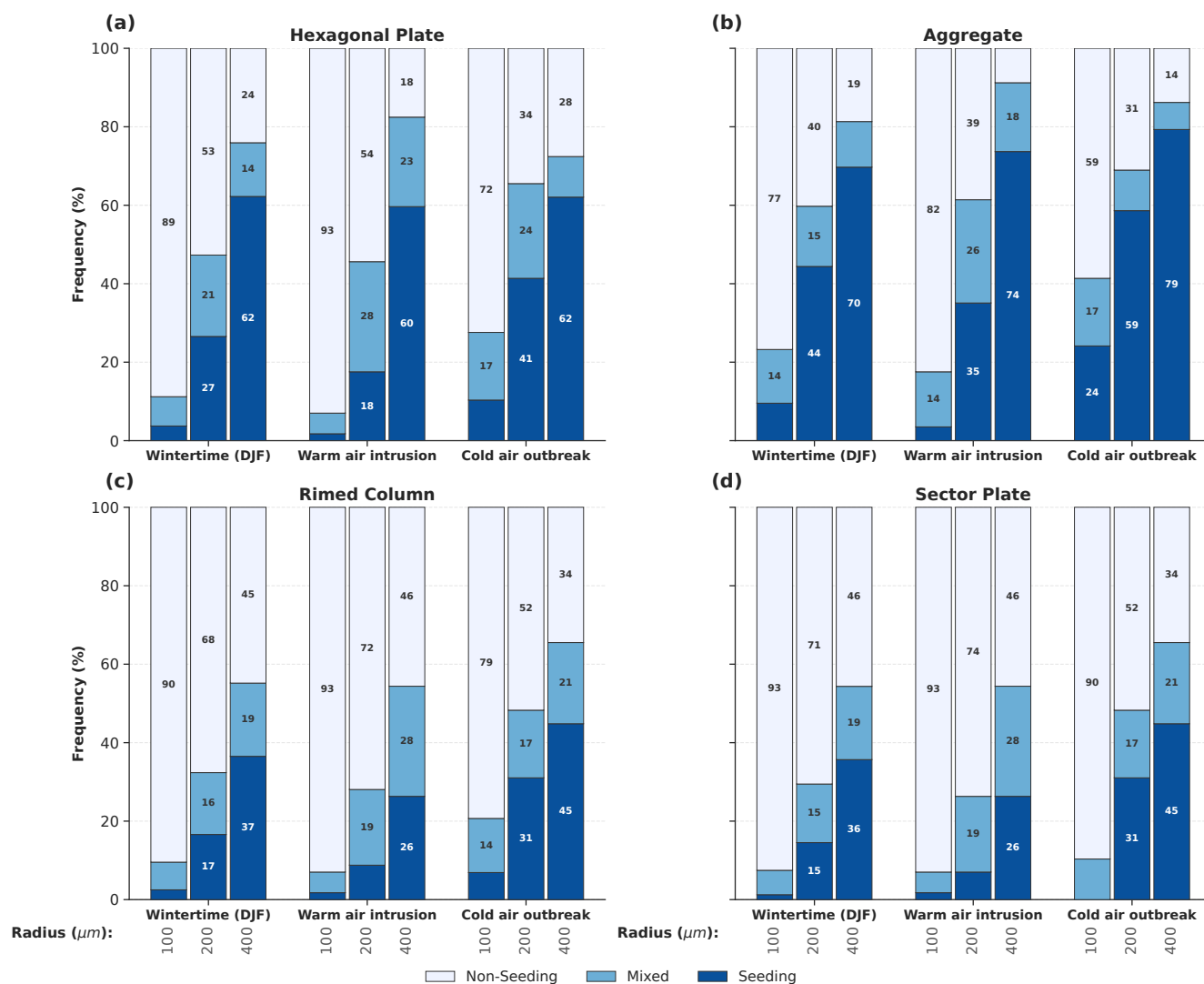


Figure A5. Seeding potential of the multilayer clouds for the wintertime mean, active (shallow and deep) WAI regime, and developing (shallow and deep) CAO regime. Percent frequency of the seeding classification is shown for three assumed initial ice-crystal radii, $r_0 = 100, 200, 400 \mu\text{m}$, across four distinct crystal habits: (a) Hexagonal Plate, (b) Aggregate, (c) Rimed Column, and (d) Sector Plate. Stacked bars show non-seeding, mixed (seeding and non-seeding), and seeding cases.

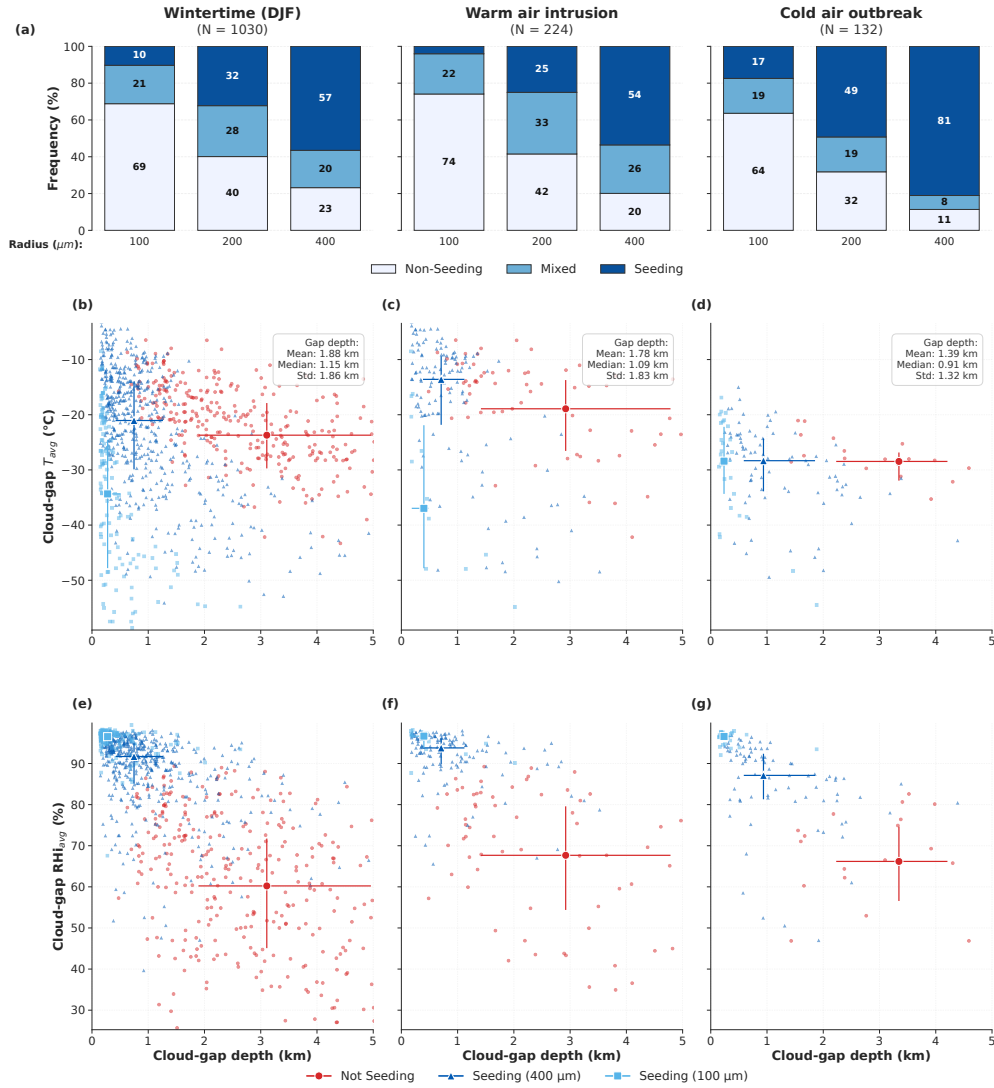


Figure A6. Seeding potential for radiosonde-only multilayer clouds during wintertime (DJF 2004/05–2024/25), warm air intrusion (WAI), and cold air outbreak (CAO) conditions. (a) Percent frequency of the seeding classification for three assumed initial aggregate ice-crystal radii, $r_0 = 100\mu\text{m}$, $200\mu\text{m}$ and $400\mu\text{m}$. Stacked bars show non-seeding, mixed, and seeding cases; the number of multilayer-cloud profiles (N) is given above each regime. (b–d) Cloud-gap depth versus mean cloud-gap temperature, T_{avg} , for (b) wintertime, (c) WAI, and (d) CAO conditions. (e–g) As in (b–d), but for mean cloud-gap relative humidity with respect to ice, $RH_{i,avg}$. Individual points represent cloud layer triplets consisting of an upper saturated layer, a subsaturated gap, and a lower saturated layer. Red circles indicate non-seeding cases, dark-blue triangles indicate seeding for $r_0 = 400\mu\text{m}$, and light-blue squares indicate seeding for $r_0 = 100\mu\text{m}$. Large symbols and whiskers show the median and interquartile range for each class. Insets in (b–d) give the mean, median, and standard deviation of the cloud-gap depth.

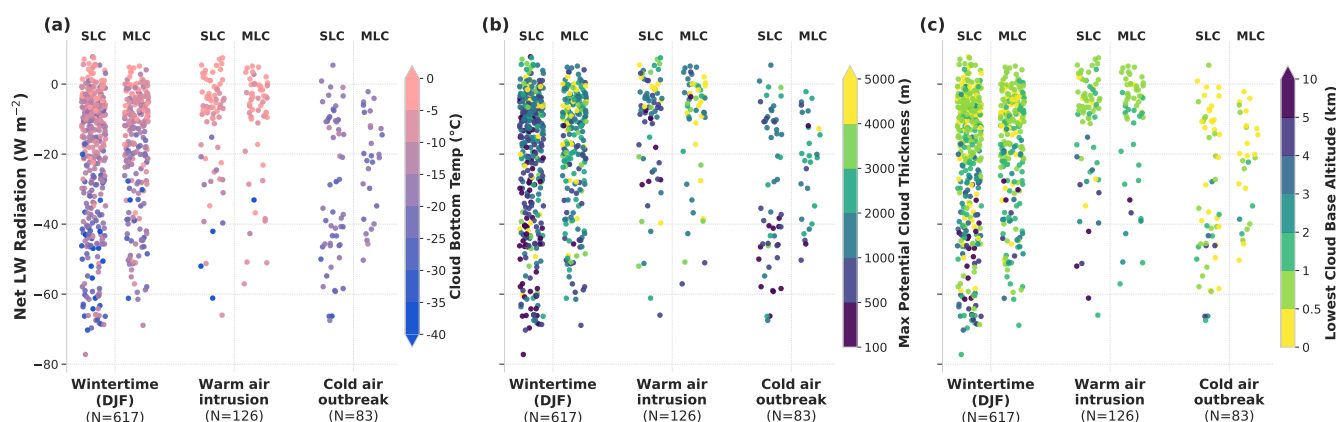


Figure A7. Net longwave radiation for single-layer clouds (SLC) and multilayer clouds (MLC) across the wintertime mean (DJF 2017-2024/25), warm air intrusion, and cold air outbreak regimes. Points represent individual observations and are colored according to (a) cloud base temperature ($^{\circ}\text{C}$), (b) maximum potential cloud thickness (m), and (c) lowest cloud base altitude (km). The total number of profiles (N) for each regime is indicated on the horizontal axis.



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