

Impacts of Mesoscale Subsidence on the Dynamics and Cloud Physics of Marine Cold Air Outbreaks



INAUGURAL-DISSERTATION

ZUR

ERLANGUNG DES DOKTORGRADES

DER MATHEMATISCH-NATURWISSENSCHAFTLICHEN FAKULTÄT

DER UNIVERSITÄT ZU KÖLN

VORGELEGT VON

FIONA MARIA PAULUS

AUS DUISBURG

KÖLN, APRIL 2026

BERICHTERSTATTENDE:

Prof. Dr. Roel Neggers

Prof. Dr. Felix Pithan

DATUM DER DISPUTATION: 16th June 2026

Impacts of Mesoscale Subsidence on the Dynamics and Cloud Physics of Marine Cold Air Outbreaks

ABSTRACT

Marine cold air outbreaks (MCAOs) are a fundamental pathway of meridional energy and moisture exchange between the Arctic and the mid-latitudes. They occur when cold, dry Arctic air masses are advected over comparatively warm open ocean, leading to strong sensible and latent heat fluxes. The resulting rapid destabilisation of the lower troposphere promotes the development of deep, cloud-topped boundary layers. During this transformation, initially stable air masses rapidly transition into a convective regime characterised by organised mixed-phase cloud fields and precipitation. Despite their importance for the Arctic and mid-latitude weather, MCAOs remain challenging to represent accurately in atmospheric models. Their multiscale characteristics require consistent treatment of processes including turbulent mixing, cloud microphysics, and large-scale forcing. In particular, large-scale subsidence has been identified as an important, yet poorly constrained, control on boundary layer development, entrainment, and cloud evolution. Quantifying subsidence is especially difficult in high-latitude regions due to limited observations, and its role in MCAO evolution remains insufficiently understood.

This thesis addresses these limitations by combining airborne observations with Lagrangian large eddy simulations (LES) to investigate the impacts of mesoscale subsidence on Arctic MCAOs. It utilises comprehensive observations during the HALO-($\mathcal{AC}$)³ campaign, which provide detailed information on atmospheric structure, cloud properties, and air-sea interactions. Dropsonde observations collected along mesoscale flight patterns enable the derivation of vertical profiles of horizontal divergence on spatial scales of about 100 km. Through mass continuity, these divergence estimates yield profiles of atmospheric subsidence. In this way, subsidence is quantified under Arctic conditions, and its magnitude, vertical structure, and uncertainty are constrained through independent observations from a different aircraft.

These observational constraints are subsequently used to initialise and force a Lagrangian LES. The evolution of an air mass is followed along its trajectory from sea-ice to open ocean, allowing a continuous description of the boundary layer transformation under observation-based forcing and boundary conditions from the extensive HALO-($\mathcal{AC}$)³ dataset. This simulation is evaluated against independent airborne remote-sensing observations and is shown to reproduce key observed features, including boundary layer depth, thermodynamic structure, integrated water vapour, and liquid and ice water path. This provides confidence in the model's ability to capture the essential processes governing MCAO evolution. Based on this simulation framework, the sensitivity of boundary layer and cloud processes to

mesoscale subsidence is systematically examined by varying its magnitude while maintaining its vertical and temporal structure consistent with the control case. In addition, the sensitivity to aerosol–cloud interactions is examined through experiments with varying concentrations of cloud condensation nuclei.

The results show that subsidence substantially influences the boundary layer depth and cloud phase. Changes in cloud condensation nuclei affect the droplet activation, precipitation formation, and phase partitioning, and these microphysical processes interact with the dynamical constraints imposed by subsidence. The system response is further shaped by boundary layer dynamics, particularly atmospheric boundary layer internal decoupling, which occurs earlier under weaker mesoscale subsidence and initiates convective graupel formation, thereby enhancing the conversion of liquid water droplets and altering cloud-phase evolution. In contrast, stronger subsidence suppresses boundary layer growth, delays decoupling, and limits cloud development, while weaker subsidence permits a deeper atmospheric boundary layer and supports more vigorous cloud evolution.

Einfluss mesoskaliger Absinkbewegungen auf die Dynamik und Wolkenphysik von marinen Kaltluftausbrüchen

ZUSAMMENFASSUNG

Marine Kaltluftausbrüche sind ein zentraler Mechanismus für den meridionalen Energie- und Feuchtigkeitsaustausch zwischen der Arktis und den mittleren Breiten. Sie entstehen, wenn kalte, trockene arktische Luftmassen über den vergleichsweise warmen offenen Ozean transportiert werden, was zu starken sensiblen und latenten Wärmeflüssen führt. Die daraus resultierende schnelle Destabilisierung der unteren Troposphäre begünstigt die Entstehung tiefer, wolkenbedeckter Grenzschichten. Im Verlauf dieser Transformation verwandeln sich die ursprünglich stabilen Luftmassen rasch in ein konvektives Regime, das durch organisierte Mischphasenwolken und Niederschlag gekennzeichnet ist. Trotz ihrer großen Bedeutung für das Wetter in der Arktis und den mittleren Breiten bleibt die realistische Darstellung von marinen Kaltluftausbrüchen in Atmosphärenmodellen eine Herausforderung. Ihre multiskaligen Eigenschaften erfordern eine konsistente Berücksichtigung unterschiedlicher Prozesse wie Turbulenz, Wolkenmikrophysik und großskaliger Bewegungen. Insbesondere sind großskalige Absinkbewegungen ein wichtiger, bislang jedoch nur unzureichend erfasster Einflussfaktor auf die Grenzschichtentwicklung und die Wolkenentwicklung. Die Quantifizierung der Absinkbewegungen ist in hohen Breitengraden aufgrund begrenzter Beobachtungsmöglichkeiten besonders schwierig, und ihre Rolle bei der Entwicklung von marinen Kaltluftausbrüchen ist nach wie vor nicht vollständig erfasst.

In dieser Dissertation werden diese Fragen adressiert, indem flugzeuggestützte Beobachtungen mit Lagrangschen Large-Eddy-Simulationen (LES) kombiniert werden, um den Einfluss mesoskaliger Absinkbewegungen auf arktische marine Kaltluftausbrüche zu analysieren. Dabei werden umfassende Beobachtungen während der HALO-($\mathcal{AC}$)³-Kampagne genutzt, die detaillierte Informationen zur atmosphärischen Struktur, zu Wolkeneigenschaften und zu Luft-Meer-Wechselwirkungen liefern. Entlang mesoskaliger Flugrouten gewonnene Dropsondendaten ermöglichen die Bestimmung vertikaler Profile der horizontalen Divergenz auf räumlichen Skalen von etwa 100 km. Über die Massenerhaltung lassen sich daraus Profile der atmosphärischen Absinkbewegung ableiten, die auf diese Weise unter arktischen Bedingungen quantifiziert werden können. Unsicherheiten werden durch unabhängige Messungen verschiedener Flugzeuge eingeschränkt.

Diese Beobachtungsdaten dienen anschließend zur Initialisierung und zum Antrieb eines Lagrangschen LES-Modells. Die Entwicklung der Luftmasse wird entlang ihrer Trajektorie vom Meereis bis zum offenen Ozean verfolgt, wodurch eine kontinuierliche Beschreibung der Grenzschichttransformation unter beobachtungsbasierten Randbedingungen sowie der großskaligen Zirkulation aus dem umfassenden HALO-($\mathcal{AC}$)³-Datensatz ermöglicht

wird. Die Simulation wird anhand unabhängiger flugzeuggestützter Fernerkundungsbeobachtungen validiert und reproduziert zentrale beobachtete Eigenschaften, darunter die Grenzschichthöhe, thermodynamische Struktur, integrierter Wasserdampf sowie Flüssig- und Eiswasserpfade. Dies stellt sicher, dass das Modell die wesentlichen Prozesse der Entwicklung von Kaltluftausbrüchen abbildet. Aufbauend auf diesem Simulationsansatz wird eine systematische Sensitivitätsanalyse durchgeführt, bei der die Reaktion von Grenzschicht- und Wolkenprozessen auf mesoskalige Absinkbewegungen untersucht wird. Hierzu wird die Stärke der beobachteten Vertikalbewegung variiert, während die vertikalen und zeitlichen Strukturen konsistent mit der Kontrollsimulation beibehalten werden. Darüber hinaus wird die Sensitivität der Aerosol-Wolken-Wechselwirkung durch Experimente mit variierenden Konzentrationen von Wolkenkondensationskernen untersucht.

Die Ergebnisse zeigen, dass großskaliges Absinken die Tiefe der Grenzschicht und die Wolkenphase maßgeblich beeinflusst. Variationen in der Konzentration von Wolkenkondensationskernen beeinflussen die Tröpfchenaktivierung, die Niederschlagsbildung sowie die Phasenaufteilung, wobei diese mikrophysikalischen Prozesse mit den durch das Absinken vorgegebenen dynamischen Randbedingungen wechselwirken. Die Systemreaktion wird zudem durch die Dynamik der Grenzschicht bestimmt, insbesondere durch ihre interne Entkopplung, die bei schwächerem mesoskaligem Absinken früher eintritt und die Bildung von konvektivem Graupel auslöst. Dies fördert die effiziente Umwandlung von flüssigen Wassertropfen zu Eiskristallen und beeinflusst die Entwicklung der Wolkenphase. Im Gegensatz dazu hemmt stärkeres Absinken das Wachstum der Grenzschicht, verzögert die Entkopplung und begrenzt die Wolkenentwicklung, während schwächeres Absinken eine tiefere atmosphärische Grenzschicht und eine intensivere Wolkenentwicklung ermöglicht.

Contents

I	SCIENTIFIC MOTIVATION	I
1.1	Arctic Amplification	4
1.1.1	Surface Albedo Feedback	4
1.1.2	Lapse-Rate Feedback	5
1.1.3	Planck Feedback	6
1.1.4	Water Vapour and Cloud Feedback	6
1.1.5	Meridional Energy Transport	7
1.2	Marine Cold Air Outbreaks	9
1.2.1	Large-Scale Circulation	10
1.2.2	Atmospheric Boundary Layer Deepening	11
1.3	Clouds in Marine Cold Air Outbreaks	12
1.3.1	Roll Cloud Formation	12
1.3.2	Roll-to-Cell Transition	13
1.3.3	Microphysical Processes	13
	Warm Cloud Microphysics	14
	Ice and Mixed-Phase Cloud Microphysics	14
1.3.4	Aerosol-Cloud Interactions	16
1.3.5	Radiative Impacts	17
1.3.6	Challenges for Numerical Representation	18
1.4	Scope of this Thesis	19
1.4.1	Research Objectives	19
2	MODELLING MARINE COLD AIR OUTBREAKS	22
2.1	Governing Equations	23
2.1.1	Ideal Gas Law	24
2.1.2	Continuity Equation	25
2.1.3	Navier-Stokes Equations	25
2.1.4	First Law of Thermodynamics	26
2.1.5	Conservation Equation for Scalars	27
2.2	Atmospheric Turbulence	27
2.2.1	Kolmogorov Hypothesis	27

2.3	Approaches to Atmospheric Modelling	29
2.4	Large Eddy Equations	30
2.4.1	Filtering	31
2.4.2	Filtered Governing Equations	32
2.4.3	Subfilter-Scale Closure	33
2.4.4	Surface Model	34
2.4.5	Parametrisation of Microphysics	36
2.5	Lagrangian Large Eddy Simulation	37
2.6	The Dutch Atmospheric Large Eddy Simulation Model	38
3	OBSERVING MARINE COLD AIR OUTBREAKS	40
3.1	HALO-(AC) ³ Campaign	41
3.1.1	Quasi-Lagrangian Observations	43
3.1.2	Additional Observations	44
3.2	Case Description	45
	MCAO Index	46
	Stability Parameter	46
3.3	Setup of Lagrangian Large Eddy Simulation	47
3.3.1	Trajectory Calculation	48
3.3.2	Initialisation	48
3.3.3	Boundary Conditions	49
3.3.4	Large-Scale Forcings from Observations	49
4	AIRBORNE MEASUREMENTS OF MESOSCALE DIVERGENCE AT HIGH LATITUDES DURING HALO-(AC) ³	55
4.1	Introduction	58
4.2	Data and Methods	60
4.2.1	HALO-(AC) ³ Campaign	60
4.2.2	Line-Integral Method	62
4.2.3	Regression Method	63
4.3	Results	65
4.4	Conclusions	71
4.5	Appendices	73
4.5.1	Appendix A - Horizontal Displacement of Dropsondes	73
4.5.2	Appendix B - Variability in the Line-Integral Method	74
4.5.3	Appendix C - Error Estimation	74
4.5.4	Appendix D - Profiles From Regression Method and Line-Integral Method with ERA5 For All Grid Boxes	76
4.5.5	Appendix E - Boundary Layer Height Calculation	79
4.5.6	Appendix F - Mean Profiles of Horizontal Wind	80
4.5.7	Appendix G - Fit Parameters of Error Statistics	80

5	IMPACTS OF MESOSCALE ATMOSPHERIC SUBSIDENCE ON CLOUD GLACIATION AND DECOUPLING IN ARCTIC MARINE COLD AIR OUTBREAKS	82
5.1	Introduction	86
5.2	Observations	89
5.2.1	The HALO-($\mathcal{AC}$) ³ Campaign	89
5.2.2	Observational Datasets	90
5.3	Model Configuration	93
5.3.1	The Dutch Atmospheric Large Eddy Simulation	93
5.3.2	Trajectory Calculation	95
5.3.3	Observational Forcing and Boundary Conditions	96
5.3.4	Experimental Setup	100
5.4	Results I: Evaluation of the Control Simulation	100
5.4.1	Atmospheric Boundary Layer and Cloud Evolution	101
5.4.2	Temperature and Humidity Profiles	102
5.4.3	Integrated Water Paths	106
5.4.4	Surface Heat Fluxes	107
5.5	Results II: Subsidence Impacts	110
5.5.1	Sensitivity Analysis	110
5.5.2	Glaciation and Decoupling	114
5.6	Conclusions	119
5.7	Appendices	120
5.7.1	Appendix A - Cold Air Outbreak and Stability Indices	120
5.7.2	Appendix B - Trajectory Calculation	122
5.7.3	Appendix C - Statistics of Observed Mesoscale Subsidence	122
5.7.4	Appendix D - Mean Absolute Percentage Error and Agreement	123
6	IMPACTS OF CLOUD CONDENSATION NUCLEI ON THE EVOLUTION OF CLOUDS IN MARINE COLD AIR OUTBREAKS	125
6.1	Introduction	126
6.2	Model Configuration	127
6.3	Results	128
6.3.1	CCN Effects across Subsidence Regimes	128
6.3.2	Ice-Phase Partitioning and Hydrometeor Response	132
6.3.3	Boundary Layer Decoupling	135
6.4	Summary and Outlook	136
7	CONCLUSIONS AND OUTLOOK	138
7.1	Summary and Conclusions	138
7.2	Outlook	144
	ACKNOWLEDGMENTS	147

LIST OF MATHEMATICAL SYMBOLS	150
LIST OF ABBREVIATIONS	153
REFERENCES	156

1

Scientific Motivation

While global mean surface temperatures have increased by 1.1 K over the period from 2011 to 2020 relative to the pre-industrial baseline (1850 to 1900; [IPCC, 2023](#)), the Arctic has warmed at a rate three to four times faster in terms of near-surface air temperature ([Rantanen et al., 2022](#); [Zhou et al., 2024](#)). This disproportionate warming is referred to as Arctic Amplification (AA; [Serreze and Francis, 2006](#); further introduced in Section 1.1) and represents a robust response to CO₂ forcing across future climate projections ([Sweeney et al., 2023](#)). AA is caused by multiple interacting feedback mechanisms involving the atmosphere, ocean, sea-ice, land surface, and cryosphere. Although many of these processes are individually well understood, their interactions and relative contributions remain partly uncertain ([Screen and Simmonds, 2010a](#); [Wendisch et al., 2023](#); [Previdi et al., 2020](#); [Taylor et al., 2022](#)).

The observed Arctic warming exhibits a pronounced spatial and seasonal structure, with the strongest temperature increases occurring in autumn and winter. These patterns are closely linked to sea-ice loss and enhanced ocean-atmosphere heat exchange in regions experiencing the most rapid warming ([Screen and Simmonds, 2010a](#); [Serreze et al., 2009](#)). The rate of

accelerated warming and corresponding loss of the cryosphere has far-reaching implications. Arctic communities face reduced access to hunting grounds and food sources as animal migration, reproduction, and behaviour shift (Novy, 2025; Hovelsrud et al., 2011; Ford et al., 2019). Drinking water and food security are threatened by contamination and environmental change (AMAP, 2017), while thawing permafrost compromises infrastructure stability (Schneider von Deimling et al., 2021). Declining sea-ice and frozen ground reduce the safety of traditional travel routes and ice roads, limiting local mobility (Ford et al., 2019; Olsen et al., 2021).

At the same time, a seasonally ice-free Arctic may enable expanded trans-Arctic shipping, shorter trade routes, and shifts in fisheries and tourism, yet these opportunities are accompanied by environmental risks, infrastructure demands, and governance challenges (Melia et al., 2016; Hovelsrud et al., 2011). Ecosystem responses are equally complex: Arctic greening modifies the surface energy balance and ecosystem functioning (Frost et al., 2020; Hartfield et al., 2018), while extreme events such as wildfires and insect outbreaks can induce Arctic browning (Phoenix et al., 2025; Hartfield et al., 2018). Ocean warming and sea-ice loss alter the light availability for marine life and air-sea exchange processes, with implications for marine photosynthesis and CO₂ uptake (McGuire et al., 2009). Competing mechanisms such as enhanced light-driven productivity, increased mixing, and stronger stratification from freshwater input affect Arctic Ocean carbon cycling and greenhouse gas concentrations (McGuire et al., 2009; Bruhwiler et al., 2021). Concurrently, large wintertime CO₂ losses from permafrost regions have been observed and are expected to increase with further warming (Natali et al., 2019).

Although direct climate forcings such as rising greenhouse gas concentrations, changes in aerosols and ozone-depleting substances, and land-use changes, contribute to Arctic warming, they alone cannot explain the observed magnitude of AA (Previdi et al., 2021). The amplification emerges from the interplay between poleward energy transport and local feedback mechanisms that either enhance or dampen the imposed forcing (Goosse et al., 2018). Understanding how these processes combine to shape Arctic temperature change is therefore central not only to regional climate projections but also to the broader climate system response.

A key pathway through which AA may exert global influence is via its impact on the large-scale atmospheric circulation. The Northern Hemisphere polar jet stream is primarily governed by the meridional temperature gradient between the equator and the Arctic through a thermal-wind balance. As AA weakens the lower-tropospheric meridional temperature gradient, an influence on the jet is expected, although the sign and magnitude of the response depend on the vertical structure of warming (Lee et al., 2019; Robinson et al., 2023). AA is characterised by maximum warming near the surface extending into the mid-troposphere (Wu and Smith, 2016). In contrast, tropical upper tropospheric warming and polar lower-stratospheric cooling strengthen the upper-level meridional temperature gradient, exerting a competing influence on the upper-level jet (Lee et al., 2019).

Idealised experiments demonstrate that AA can induce an equatorward shift of the tropospheric jet and weaken the stratospheric polar vortex, with stratosphere-troposphere coupling contributing to the tropospheric response (Wu and Smith, 2016). Recent studies emphasise the role of free-tropospheric temperature gradients, linking observed jet slowdown to enhanced Arctic free-tropospheric warming and gradient weakening (Robinson et al., 2023), while stronger eddy feedbacks are associated with larger equatorward jet shifts (Screen et al., 2022). Regions of weakened poleward gradients are associated with reduced zonal winds and more meridional, wavier flow (Francis and Vavrus, 2015). Recent reanalysis indicates increased wintertime directional variability, together with jet weakening and more frequent easterly flow near the Arctic boundary, whereas over the North Atlantic, no significant trend in annual-mean jet speed is detected, despite increasing vertical shear (Degirmendžić and Marosz, 2025; Lee et al., 2019). Many climate simulations indicate jet slowdown associated with Arctic free-tropospheric warming, yet a majority project future jet strengthening under strong CO₂ forcing due to pronounced subtropical upper-level warming (Robinson et al., 2023). At the same time, difficulties in representing observed polar jet strengthening over the North Atlantic persist (Blackport and Fyfe, 2022), and future jet shifts reflect competing influences of sea-ice loss and ocean warming with substantial model spread (Screen et al., 2022).

Changes in jet position and structure are dynamically relevant because they modulate the pathways through which polar air masses can intrude into the mid-latitudes. A southward jet extension can facilitate equatorward advection of cold Arctic air (Smith and Sheridan, 2018). When such outbreaks occur over open ocean, they are referred to as marine cold air

outbreaks (MCAOs). During these events, cold, dry Arctic air is advected over comparatively warm water, producing strong low-level instability and intense turbulent air-sea heat fluxes (Fletcher et al., 2016a; Kolstad et al., 2009; Brümmer, 1996). MCAOs frequently occur in the cold-air sector of midlatitude cyclones embedded within the storm tracks linked to the jet stream (Fletcher et al., 2016a). The resulting surface fluxes rapidly deepen and moisten the marine boundary layer, reshape cloud structures, and modify downstream weather (Fletcher et al., 2016b; Brümmer, 1996). In extreme cases, the enhanced boundary layer instability favours the development of small, intense cyclones, called polar lows (Papritz and Grams, 2018; Rasmussen, 2003).

Marine cold air outbreaks thus represent a dynamically and thermodynamically sensitive interface between Arctic change and midlatitude weather. Their intensity, regional occurrence, and structure depend on the evolving temperature gradients, sea-ice distribution, and large-scale circulation patterns potentially shaped by AA (Kolstad and Bracegirdle, 2008; Ayarzagüena and Screen, 2016; Overland et al., 2021).

The following section provides a brief introduction to the key feedback mechanisms underlying AA (Section 1.1). It then introduces MCAOs (Section 1.2), discussing their characteristics and large-scale circulation. Finally, clouds in MCAOs are examined in greater detail, including their microphysical processes and interactions with aerosols, radiative impacts, and the challenges associated with their numerical representation (Section 1.3).

1.1 ARCTIC AMPLIFICATION

The magnitude and structure of AA arise from a combination of external forcings and internal feedback processes. While poleward energy transport sets the large-scale context, local feedbacks involving sea-ice, clouds, water vapour, lapse-rate changes, and surface albedo strongly modulate the Arctic temperature response (Goosse et al., 2018). A schematic overview of the principal processes is provided in Figure 1.1. In the following section, these feedback mechanisms are examined in more detail.

1.1.1 SURFACE ALBEDO FEEDBACK

The increasing global temperatures accelerate the melting of sea-ice and snow in the Arctic and expose the open ocean, decreasing the surface reflectivity and enhancing solar absorption

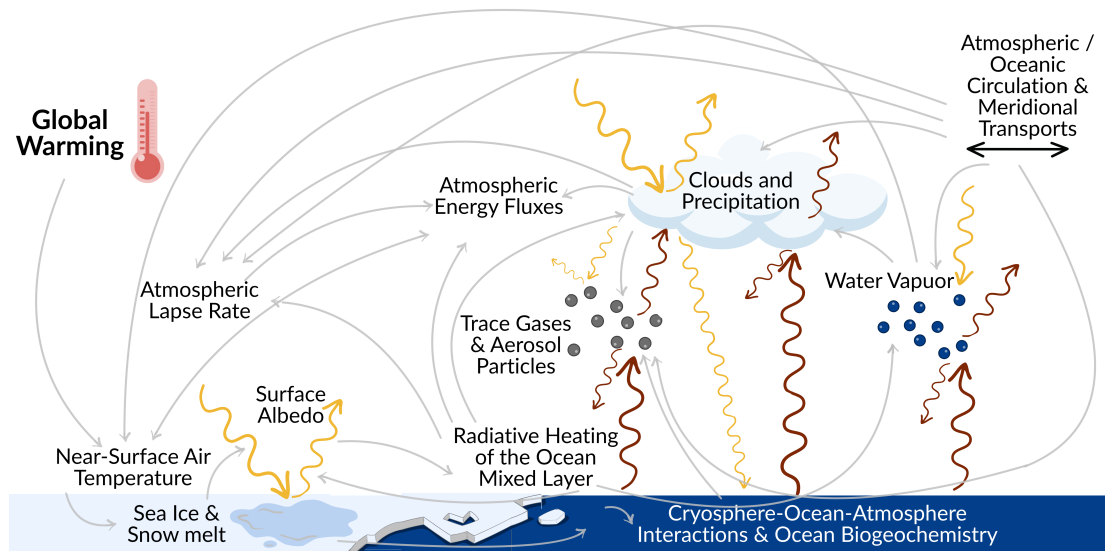


Figure 1.1: Schematic of important feedback mechanisms impacting Arctic Amplification and their interactions. Yellow arrows depict the solar (shortwave) radiation, and red arrows depict the terrestrial (longwave) radiation. Adapted from [Goosse et al. \(2018\)](#) and [Wendisch et al. \(2023\)](#).

([Dickinson et al., 1987](#); [Perovich et al., 2007](#)). The ocean warms by the absorbed radiation and emits more longwave radiation upward, warming the near-surface air and reinforcing the sea-ice melt ([Manabe and Wetherald, 1975](#); [Sellers, 1969](#); [Budyko, 1969](#)). Apart from this direct feedback loop during polar day, the surface albedo feedback has an additional delayed effect in autumn. The solar energy warms the upper ocean layers (ocean heat uptake), continuously heating the near-surface air and delaying refreezing in autumn. Once the sea-ice cover reforms, it acts as an insulating lid and limits further ocean heat loss to the atmosphere (e.g., [Screen and Simmonds, 2010a,b](#); [Boeke and Taylor, 2018](#); [Dai and Deng, 2021](#)).

1.1.2 LAPSE-RATE FEEDBACK

The lapse-rate feedback arises from the vertical structure of atmospheric warming and its radiative effects (e.g., [Manabe and Wetherald, 1975](#); [Hansen et al., 1997](#); [Colman, 2003](#)). While tropical moist convection enhances upper-tropospheric warming and produces a negative lapse-rate feedback that increases outgoing longwave radiation (e.g., [Bony et al., 2006](#); [Lambert and Taylor, 2014](#)), the Arctic is characterised by stable stratification and frequent near-surface temperature inversions. These lead to bottom-heavy warming, which weakens the increase in outgoing longwave radiation and generates a positive lapse-rate feedback ([Bin-](#)

tanja et al., 2011; Pithan and Mauritsen, 2014; Lauer et al., 2020). The Arctic lapse-rate feedback is shaped by both local processes, such as sea-ice loss and boundary layer warming, and remote influences from poleward atmospheric energy transport (Feldl et al., 2020). Across the CMIP6 models, differences in the strength of the Arctic lapse-rate feedback are strongly correlated with the magnitude of AA, with models exhibiting stronger near-surface inversions and weaker poleward atmospheric energy transport tending to produce stronger lapse-rate feedback and larger high-latitude warming (Linke et al., 2023).

1.1.3 PLANCK FEEDBACK

The Earth can be considered as a black body radiating thermal radiation based on its temperature (Planck, 1914). With rising temperatures, the thermal radiative flux also increases, radiating away more energy and therefore stabilising the Earth's temperature. This is called Planck feedback, and is the strongest baseline constraint limiting climate change (Cronin and Dutta, 2023; Feng et al., 2023). The radiative response to temperature increases is non-linear, with more heat being radiated away for larger temperatures; therefore, the Planck feedback is not as effective in limiting the effects of climate forcings in the colder polar regions than in the warmer tropics. This leads to an increase in the difference between warming in the Tropics and the Arctic and thus contributes to AA (Henry and Merlis, 2019; Previdi et al., 2020).

1.1.4 WATER VAPOUR AND CLOUD FEEDBACK

The rising sea-surface temperatures due to radiative heating of the ocean surface and mixed layer enhance evaporation and cause larger water vapour concentrations in the atmosphere. In addition, according to the Clausius-Clapeyron relationship (Clapeyron, 1834; Clausius, 1856), a warmer atmosphere can hold substantially more water vapour (roughly 6 % to 7 % per 1 K warming), thereby preventing rapid saturation and allowing atmospheric moisture concentrations to rise (O'Gorman and Muller, 2010). Without this thermodynamic constraint, increased evaporation would quickly lead to saturation and precipitation rather than sustained increases in atmospheric humidity. The additional water vapour acts as a greenhouse gas, intensifying the downward longwave radiative fluxes and leading to a positive feedback that enhances the near-surface air temperatures (Francis and Hunter, 2007; Ghatak and Miller, 2013; Patel and Kuttippurath, 2023). In addition, the reduced sea-ice cover desta-

bilises the boundary layer and enhances the moisture availability, promoting the formation of clouds, which are often mixed-phase in the Arctic (Jenkins and Dai, 2021; Gierens et al., 2020). Observations suggest that the Arctic boundary layer often alternates between two preferred regimes: a radiatively clear state and an opaque cloudy state associated with persistent mixed-phase clouds (Stramler et al., 2011). The cloud cover in the warming Arctic has increased, particularly during fall and winter (e.g., Vavrus, 2004; Kay and Gettelman, 2009; Vavrus et al., 2009; Wu and Lee, 2012; Philipp et al., 2020). These low-level clouds emit substantial downwelling longwave radiation, thereby enhancing surface warming. This effect is especially relevant during the polar night, when no incoming solar radiation is reflected, and associated cooling effects are absent (Taylor et al., 2013; Cao et al., 2017; Arouf et al., 2024; Tao et al., 2025). During polar day, the clouds' high albedo can reduce incoming solar radiation and cool the surface. However, this cooling effect is comparatively weak over sea-ice-covered regions because snow and sea-ice already exhibit a high albedo (Intrieri et al., 2002; Shupe and Intrieri, 2004; Wendisch et al., 2023). The net radiative impact of these clouds strongly depends on their phase composition, as liquid and ice have distinct radiative properties (Tan and Storelvmo, 2019) (see Section 1.3). Although global climate models generally simulate a net positive cloud feedback, uncertainties remain large (Bony et al., 2006; Zelinka et al., 2020; Sherwood et al., 2020; Lutsko et al., 2022). Especially in the Arctic, the sign and magnitude of the cloud feedback remain insufficiently constrained due to uncertainties in cloud-radiation interactions and the representation of mixed-phase cloud processes in models (Nakanishi and Michibata, 2025). Therefore, the interactions between the cloud feedback and other AA mechanisms, and in contrast to the tropical cloud feedback, remain incompletely understood. Consequently, the overall contribution of cloud feedback to AA is still uncertain (Previdi et al., 2020). Nevertheless, evidence consistently highlights the central role of low-level mixed-phase clouds in shaping the Arctic climate system (Zhang et al., 2018; Tan and Storelvmo, 2019; Huang et al., 2021).

1.1.5 MERIDIONAL ENERGY TRANSPORT

Transport processes between the Arctic and the lower latitudes respond to external radiative forcings and are coupled to the local-feedbacks governing AA (e.g., Previdi et al., 2021; Pithan et al., 2018; Wendisch et al., 2023). In a warming climate, atmospheric poleward energy transport changes in both magnitude and partitioning. While AA tends to weaken the dry-

static transport by reducing the tropospheric meridional temperature gradient, latent energy transport often increases as the atmosphere moistens (see Section 1.1.4; [Hwang et al., 2011](#); [Graversen and Langen, 2019](#); [Feldl et al., 2020](#); [Russotto and Biasutti, 2020](#)). However, poleward transport of heat and moisture is not a steady flux. It occurs mainly through episodic air mass exchanges between the mid-latitudes and the Arctic region ([Pithan et al., 2018](#); [Papritz, 2020](#); [Murto et al., 2022](#)).

The main pathways for these exchanges involve warm and moist air intrusions (WAIs) entering the Arctic and cold air outbreaks moving from the Arctic toward lower latitudes ([Pithan et al., 2018](#); [Wendisch et al., 2023, 2025](#)). These processes facilitate the redistribution of energy and moisture between the two regions.

During WAIs, mid-latitude air masses bring large amounts of heat and moisture into the Arctic (e.g., [Woods and Caballero, 2016](#); [Binder et al., 2017](#); [Lauer et al., 2023](#); [Walbröl et al., 2024](#)). As these air masses cool and interact with clouds and the surface, rapid radiative and microphysical changes occur. Such events strongly modify the Arctic surface energy balance ([Persson et al., 2017](#); [Murto et al., 2023](#); [Papritz et al., 2023](#)) through increased water vapour and cloud cover, enhancing the downward longwave radiation and promoting surface warming and sea-ice melt (see Section 1.1.4 [Tjernström et al., 2015](#); [Yang and Magnusdottir, 2017](#); [Johansson et al., 2017](#)). In these events, blocking patterns and cyclones channel warm, humid air into the Arctic basin through preferred pathways ([Fearon et al., 2021](#); [Murto et al., 2022](#); [Lauer et al., 2023](#)).

The complementary branch of this exchange occurs during cold air outbreaks. When cold and dry Arctic air flows over comparatively warm ocean and land surfaces, intense sensible and latent heat fluxes rapidly warm and moisten the lower atmosphere ([Fletcher et al., 2016a](#); [Papritz et al., 2019](#); [Papritz, 2020](#)). In particular, Marine cold air outbreaks over the open ocean lead to deepening boundary layers and the development of organised convection, such as cloud streets (e.g., [Müller et al., 1999](#); [Renfrew and Moore, 1999](#); [Fletcher et al., 2016b](#)). Through these rapid air mass transformations, MCAOs transfer heat and moisture from the ocean to the atmosphere, contributing to cloud formation and precipitation and are found to be responsible for a substantial fraction of wintertime ocean heat loss in the Nordic Seas ([Papritz and Spengler, 2017](#); [Stoll et al., 2018](#); [Terpstra et al., 2021](#); [Svingen et al., 2023](#)). A more in-depth discussion of the processes in MCAOs will be given in Section 1.2.

Taken together, these episodic exchanges illustrate that meridional energy transport cannot be interpreted solely as a large-scale diffusive transport. Instead, it must be understood in terms of dynamically driven air mass transformations that interact with regional feedback processes (Papritz, 2020; Pithan et al., 2018; Wendisch et al., 2025). Enhanced poleward moisture transport was found to increase Arctic cloudiness and greenhouse forcing, modify lapse-rate feedbacks, and accelerate sea-ice decline, thereby reinforcing AA (Woods and Caballero, 2016; Mortin et al., 2016; Yang and Magnúsdóttir, 2017; Johansson et al., 2017; Feldl et al., 2020).

1.2 MARINE COLD AIR OUTBREAKS

MCAOs, as introduced in Section 1.1.5, can quickly shift the atmospheric boundary layer (ABL) from stable to convective within a short downstream distance from the sea-ice edge (Brümmer, 1996; Michaelis et al., 2022; Kirbus et al., 2024). With increasing distance to the sea-ice edge, the ABL deepens, warms, and moistens; its depth is controlled primarily by the air-sea temperature contrast, inversion strength and large-scale subsidence, with wind shear and synoptic flow influencing the convective organisation and evolution (Fletcher et al., 2016b; Murray-Watson et al., 2023; Schirmacher et al., 2024).

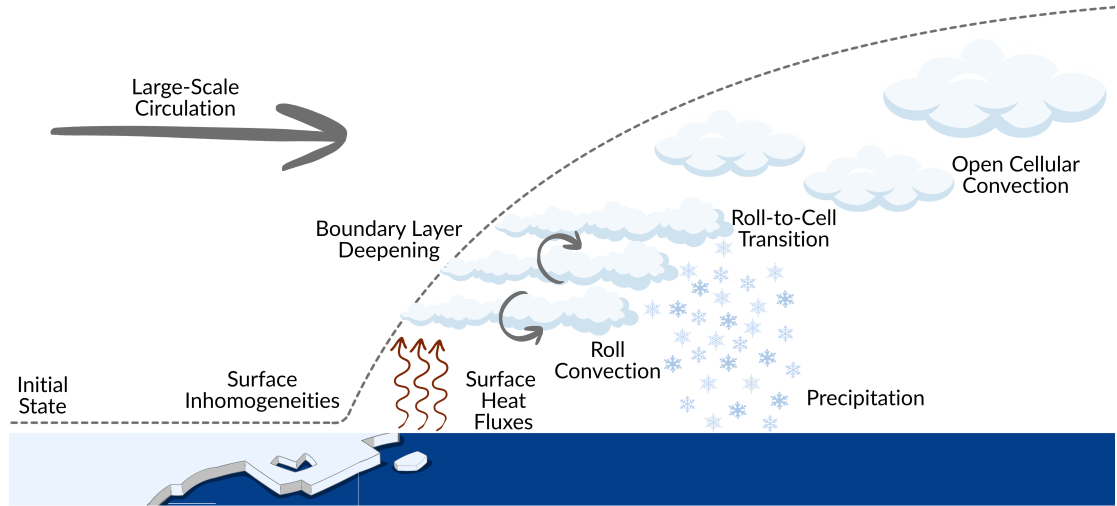


Figure 1.2: Schematic of the ABLs development and organised convection during MCAOs. The boundary layer top is indicated as a dashed line.

A defining feature of MCAOs is the development of organised turbulent structures in the marine boundary layer. Convection develops in organised mesoscale patterns, initially as roll convection aligned with the mean wind (e.g., Brümmer and Pohlmann, 2000; Gryschka and Raasch, 2005; Gryschka et al., 2008; Wu and Ovchinnikov, 2022; Murray-Watson and Gryspeerdt, 2024; Klingebiel et al., 2025). These rolls contribute to the vertical turbulent transport, thereby modulating the boundary layer growth (Gryschka et al., 2014). As the distance to the sea-ice increases and the ABL stability decreases, these organised patterns can shift to cellular convection, reflecting a changing balance between surface fluxes and entrainment at the boundary layer top, impacting the static stability (Brümmer, 1999; Brümmer and Pohlmann, 2000; McCoy et al., 2017; Tornow et al., 2021). A schematic overview of the relevant processes in MCAO development is given in Figure 1.2.

MCAOs are multiscale phenomena and result from the coupling of multiple processes, like surface fluxes, cloud microphysics and boundary layer turbulence (Wu and Ovchinnikov, 2022; Murray-Watson and Gryspeerdt, 2024; Juliano et al., 2024; Wu et al., 2025). The vertical distribution of heat and moisture is largely governed by small-scale turbulent mixing, while the cloud fields are organised by mesoscale activity (Gryschka et al., 2014; de Roode et al., 2019; Juliano et al., 2024). Larger-scale circulations largely impact the outbreak’s structure and duration (Kolstad et al., 2009; Dahlke et al., 2022). The interactions of these scales present a significant challenge for atmospheric models, which must accurately represent key processes such as entrainment, cloud formation, and organised convection (Klein et al., 2009; Tomassini et al., 2017). Therefore, the combination of comprehensive observations and process-resolving modelling approaches, including large eddy simulations (LES), is essential for advancing the understanding and representation of MCAO dynamics (Gryschka and Raasch, 2005; de Roode et al., 2019; Michaelis et al., 2022; Juliano et al., 2024, 2026).

1.2.1 LARGE-SCALE CIRCULATION

Synoptic-scale circulation patterns govern the development of MCAOs by facilitating the southward transport of Arctic air masses. These events frequently coincide with extratropical cyclones and often develop in the wake of associated cold fronts (Fletcher et al., 2016a; Terpstra et al., 2021). In the Fram Strait between Svalbard and Greenland, intense MCAOs have been linked to dipole-like sea-level pressure patterns, with low pressure over the Barents Sea or eastern Arctic and high pressure near Greenland (Dahlke et al., 2022; Walbröl et al., 2024).

After their initiation, the evolution of MCAOs has been found to be further modulated by large-scale dynamical processes (Brümmer, 1996; Wendisch et al., 2025). The main driver of their evolution is the air-sea temperature contrast, as discussed in Section 1.2, but factors such as mesoscale subsidence and the stability aloft may influence the development of the boundary layer and the longevity of outbreaks. Several studies indicate that variations in subsidence can suppress cloud growth and alter the early evolution of outbreaks, under similar thermodynamic forcing (Brümmer, 1997; McCoy et al., 2017; Young et al., 2018; Schirmacher et al., 2024).

In the presence of global warming and AA, changes in sea-ice distribution and associated shifts in meridional temperature gradients may modify the large-scale circulation, including the position and strength of storm tracks, frontal zones, and the jet stream, which in turn might alter the frequency, pathways, and intensity of MCAOs; however, these responses remain uncertain and regionally variable (Kolstad and Bracegirdle, 2008; Pedersen et al., 2016; Screen, 2017; Dahlke et al., 2022).

1.2.2 ATMOSPHERIC BOUNDARY LAYER DEEPENING

During the initial stage of an MCAO over sea-ice, air masses are typically very cold, dry, and stably stratified, and are characterised by shallow boundary layers and often strong temperature inversions (Wu and Ovchinnikov, 2022; Kirbus et al., 2024). Upon reaching the marginal sea-ice zone (MIZ), extremely strong sensible and latent heat fluxes destabilise the boundary layer, causing rapid near-surface diabatic heating and moistening, at rates exceeding 6 K h^{-1} and $0.3 \text{ g kg}^{-1} \text{ h}^{-1}$, respectively, shortly after crossing the MIZ (Fletcher et al., 2016a; Kirbus et al., 2024). Consequently, static stability is quickly reduced, initiating shallow convection and subsequent cloud formation (Brümmer, 1996).

The boundary layer subsequently deepens rapidly with increasing downstream distance from the sea-ice edge. Brümmer (1996) reported an ABL growth from initially about 100 m to 300 m to 900 m to 2200 m after approximately 300 km off the sea-ice edge. Similarly, Schirmacher et al. (2024) observed cloud formation and deepening after about 15 km over the open ocean, transitioning to a near-continuous cloud cover after about 20 km.

This evolution is broadly consistent with classical convective boundary layer theory, in which the surface buoyancy flux is expected to drive the turbulent growth of the mixed layer (Mahrt

and Lenschow, 1976; Stevens, 2007). In an idealised framework, the boundary layer depth then scales as $z_i \propto t^{\frac{1}{2}}$, reflecting a balance between surface-driven turbulence and cloud-top entrainment across the inversion (Mahrt and Lenschow, 1976). Observational studies of MCAOs typically record this rapid initial deepening, but deviations from the idealised scaling occur due to spatial variations in surface fluxes, changing inversion strength, and broader dynamical factors such as subsidence and mesoscale divergence (Brümmer, 1996; Wu and Ovchinnikov, 2022; Schirmacher et al., 2024).

1.3 CLOUDS IN MARINE COLD AIR OUTBREAKS

During MCAOs extensive fields of boundary layer stratocumulus clouds are generated. These clouds are typically not randomly distributed but exhibit a high degree of spatial organisation, meaning that convection arranges into coherent mesoscale patterns with characteristic structures, scales, and orientations (Brümmer and Pohlmann, 2000). As the MCAO propagates downstream, a distinct transformation between different organised structures emerges, which will be discussed below.

1.3.1 ROLL CLOUD FORMATION

The initial convection of an MCAO often forms roll vortices aligned with the mean boundary layer wind, creating cloud streets (Renfrew and Moore, 1999; Gryschka et al., 2008). This represents the direct convective response of the initially stable air mass after crossing the MIZ, where it is rapidly destabilised from below (Murray-Watson et al., 2023). Roll clouds emerge from the interaction of buoyancy-driven turbulence and low-level wind shear (Atkinson and Wu Zhang, 1996; Weckwerth et al., 1997; Gryschka and Raasch, 2005). Their intensity and coherence depend on the outbreak strength and local conditions (Wu and Ovchinnikov, 2022). The MIZ is highly variable; local differences in ice concentration and edge geometry can strongly influence where rolls form and how organised they become (Liu et al., 2006; Gryschka et al., 2008). Roll circulations have been found to contribute significantly to the vertical transport of heat and moisture and can thereby modulate the boundary layer depth and air mass transformation (Gryschka et al., 2014). In this near-ice-edge regime, cloud streets have been observed to extend more than a hundred kilometres downwind (Wu et al., 2025).

1.3.2 ROLL-TO-CELL TRANSITION

Organised rolls often evolve into cellular convection after a few hundred kilometres downstream of the ice edge, although the transition distance varies substantially by case (Raasch, 1990; Müller and Chlond, 1996; Gryschka and Raasch, 2005; Wu and Ovchinnikov, 2022). As the boundary layer deepens with increasing distance from the sea-ice edge, buoyancy-driven turbulence often becomes increasingly important relative to low-level shear, which weakens roll organisation and favours broader cellular patterns with cloud top heights typically between 1 km and 3 km (Gryschka et al., 2008; McCoy et al., 2017; Kirbus et al., 2023a; Mages et al., 2023).

Recent studies have linked the roll-to-cell transition in MCAOs to a weakening of the coupling between the surface mixed layer and the cloud layer, often associated with increasingly layered thermodynamic profiles of temperature and humidity. These studies suggest that mixed-phase microphysical processes, especially riming and precipitation, promote evaporation or sublimation below clouds, which cools and stabilises the subcloud layer, enhances stratification, and reduces vertical mixing (Abel et al., 2017; Murray-Watson et al., 2023; Murray-Watson and Gryspeerdt, 2024; Wu et al., 2025). As a result, roll circulations are less efficiently maintained and the cloud field transitions toward cellular convection (Juliano et al., 2026). Similar processes have been identified in subtropical stratocumulus transitions, where decoupling is associated with a layered boundary layer structure, diagnosed by the development of negative buoyancy flux $\overline{w'\theta_v'}$ below cloud base (Bretherton and Wyant, 1997; Sandu and Stevens, 2011; van der Dussen et al., 2013).

In MCAOs, cellular convection often occurs in an open-cellular form, while closed cells tend to arise under more stable boundary layer conditions and weaker air-sea temperature contrasts (McCoy et al., 2017). Closed cells are characterised by a relatively continuous cloud cover within the cellular pattern, whereas open cells exhibit relatively clear centres surrounded by cloudy cell walls and have been associated with stronger precipitation (Agee, 1987; Atkinson and Wu Zhang, 1996; Abel et al., 2017; McCoy et al., 2017).

1.3.3 MICROPHYSICAL PROCESSES

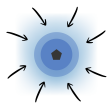
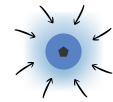
The evolution of these clouds is largely influenced by microphysical processes between cloud hydrometeors (Morrison et al., 2012, 2020). Compared with warm liquid clouds, aerosol-

cloud interactions in mixed-phase clouds involve both droplet activation and ice formation, leading to complex interactions among liquid droplets, ice crystals, and precipitation particles. The following section is largely based on [Pruppacher and Klett \(2012\)](#) and [Morrison et al. \(2020\)](#) and introduces warm-cloud, ice, and mixed-phase microphysical processes, which are specifically relevant for MPCs, in more detail.

WARM CLOUD MICROPHYSICS

Warm clouds consist of hydrometeors in the liquid phase, i.e., cloud droplets and potentially rain. Similar processes also occur in mixed-phase clouds among supercooled liquid droplets.

Cloud droplets form in air masses that are supersaturated with respect to liquid water, when aerosol particles take up water vapour and, upon reaching a critical supersaturation, become activated and grow freely into macroscopic droplets, thereby acting as cloud condensation nuclei (CCN). Their activation depends on the particle size and composition, with typical activated concentrations ranging from about 50 cm^{-3} to 100 cm^{-3} in clean maritime or pristine air to about 1000 cm^{-3} in continental air masses ([Bigg et al., 1996](#); [Stohl et al., 2006](#); [Garrett et al., 2010](#)).



Once activated, droplets grow by condensation, depleting water vapour and limiting supersaturation. This process is most efficient for small droplets and slows with increasing droplet radius.

ICE AND MIXED-PHASE CLOUD MICROPHYSICS

In MPCs, additional ice processes occur, as well as interactions between liquid and frozen hydrometeors.

At warmer temperatures (between 0°C and about -38°C), ice formation occurs primarily via heterogeneous nucleation on ice nucleating particles (INPs). INP activation strongly depends on temperature and aerosol properties, and their concentrations are typically orders of magnitude lower than those of CCN, ranging from about 10^{-5} cm^{-3} near -10°C to at about 10^{-2} cm^{-3} at about -25°C ([DeMott et al., 2025](#)).





Ice formation can occur under supersaturation with respect to ice or through heterogeneous freezing pathways such as immersion freezing, in which an INP within a droplet triggers freezing as temperature decreases.

Additionally, supercooled droplets can collide with INPs, triggering contact freezing. Due to the scarcity of INPs, supercooled liquid droplets can persist in MPCs.



Similar to the condensational growth of liquid droplets, ice particles can grow by vapour deposition when the air is supersaturated with respect to ice.

A key process in MPCs is the Wegener-Bergeron-Findeisen process (WBF): because the saturation vapour pressure over liquid water is higher than over ice, a cloud that is near liquid saturation is supersaturated with respect to ice. Vapour therefore diffuses from evaporating droplets toward growing ice crystals. The efficiency of this process depends on ice number concentration, liquid water content, and spatial inhomogeneities.



Ice particles formed through these processes may initially be pristine crystals and can grow into snowflakes if they become large enough to precipitate. When falling through the cloud, they may collide with supercooled droplets, which freeze upon contact in a process known as riming. This can produce rimed ice particles or snowflakes, and under strong riming conditions, dense graupel particles.

Collisions between ice crystals can also lead to aggregation, where particles stick together to form larger snowflakes. This process is most efficient for large, branched crystals and typically produces low-density aggregates.



Secondary ice production can increase the ice number concentration without additional INP activation. Proposed mechanisms include fragmentation during droplet freezing and splintering during riming, although their relative importance remains subject to ongoing research (e.g., [Zhao et al., 2021](#); [Zhao and Liu, 2022](#); [Seidel et al., 2024](#); [Wallentin et al., 2025](#)).

1.3.4 AEROSOL-CLOUD INTERACTIONS

Acting as CCN and INPs, aerosol populations play an important role in shaping the microphysical structure and evolution of Arctic MPCs, by affecting the phase partitioning, precipitation pathways, and cloud longevity (Lance et al., 2011; Eirund et al., 2019; Wex et al., 2019; Wu et al., 2025).

The Twomey effect describes how increased CCN lead to more numerous but smaller droplets, whereas lower CCN concentrations favour fewer and larger droplets (Twomey, 1974). These differences in droplet size distributions directly influence collision-coalescence processes and warm-rain formation, with higher CCN concentrations generally suppressing precipitation and promoting liquid water retention (Stevens and Feingold, 2009; Chen et al., 2021; Barrett and Hoose, 2023). In MPCs, however, these liquid-phase responses are tightly coupled to ice processes initiated by INPs, such as particle growth via the WBF process (Young et al., 2018; Eirund et al., 2019; Wex et al., 2019).

In Arctic MPCs, aerosol effects therefore emerge from the coupled influences of CCN and INPs on both liquid and ice microphysics. LES studies demonstrate that CCN perturbations can modify droplet size and liquid water content, which in turn can influence riming efficiency, while INP concentrations directly regulate ice nucleation rates (Eirund et al., 2019; Barrett and Hoose, 2023; Chylik et al., 2023). These interactions are strongly interdependent: smaller droplets associated with high CCN can suppress riming, whereas increased INPs can accelerate glaciation, deplete liquid water, and enhance precipitation (Lance et al., 2011; Eirund et al., 2019; Barrett and Hoose, 2023). As a result, aerosol perturbations can shift the balance between liquid retention and ice formation, with important consequences for cloud persistence, radiative properties, and precipitation (Wex et al., 2019; Tornow et al., 2021; Wu et al., 2025).

However, aerosol-cloud-precipitation interactions in Arctic MPCs are often buffered by compensating feedbacks and regime-dependent responses (Eirund et al., 2019; Tornow et al., 2021; Chylik et al., 2023). For example, enhanced ice formation associated with increased INPs can rapidly reduce liquid water through the WBF process, while CCN-induced increases in liquid water may be offset by enhanced precipitation or ice production (Eirund et al., 2019). The magnitude and persistence of aerosol-cloud effects depend strongly on environmental conditions such as surface type, thermodynamic structure, and aerosol avail-

ability (Chylik et al., 2023; Wu et al., 2025). For instance, Eirund et al. (2019) showed that CCN-induced increases in liquid water path may be transient over the open ocean due to enhanced precipitation, but more persistent over sea-ice. Observational and modelling studies further suggest that aerosol perturbations, through both CCN and INP pathways, can influence boundary layer structure via radiative and dynamical feedbacks, including entrainment, inversion strength, and turbulence (Chylik et al., 2023; Tornow et al., 2021; Sterzinger and Igel, 2024).

1.3.5 RADIATIVE IMPACTS

The radiative effect of low-level MPCs in MCAOs depends strongly on cloud phase and liquid water content (Shupe and Intrieri, 2004). Liquid-containing low clouds are highly emissive in the longwave and can enhance downwelling infrared radiation enough to warm the surface (Intrieri et al., 2002; Morrison et al., 2011, 2012). Meanwhile, they cool the cloud top through strong radiative emission. This cloud-top cooling reduces the static stability of the cloud layer and encourages buoyancy-driven turbulence, which helps maintain near-cloud-top saturation, supports ongoing condensational growth (see Section 1.3.3) and can offset, though not eliminate, losses from ice growth and precipitation (Morrison et al., 2012; Shupe et al., 2013; Rauterkus and Ansorge, 2020; Neggers et al., 2025). As a result, these clouds can persist as resilient boundary layer features, sustained by coupled interactions among radiation, turbulence, entrainment, microphysics, and often surface fluxes (Morrison et al., 2012; Neggers et al., 2019).

MCAOs in the Arctic are strongly seasonal and occur mainly during the cold season (autumn to spring), with the highest frequency and intensity found in winter (Fletcher et al., 2016a; Dahlke et al., 2022). During these cold-season months, MPCs play an important role in the surface energy budget. In the absence of solar radiation, their radiative impact is dominated by longwave processes, with low-level MPCs enhancing downwelling longwave radiation and thereby warming the surface (Shupe and Intrieri, 2004; Bennartz et al., 2013).

Because Arctic cloud radiative forcing depends strongly on phase partitioning, mixed-phase cloud regimes that often develop during MCAOs can be effective at promoting surface warming under low-sun or polar night conditions and may contribute to a positive cloud feedback (Section 1.1.4; Shupe and Intrieri, 2004; Kapsch et al., 2016).

1.3.6 CHALLENGES FOR NUMERICAL REPRESENTATION

The representation of MCAOs in atmospheric models poses challenges due to the multitude of interacting processes at different scales. Global and regional models often underresolve mesoscale organisation and turbulent transport, while high-resolution process models typically rely on prescribed or idealised large-scale forcing, making MCAO simulations sensitive to how key processes are represented (Field et al., 2014; Tomassini et al., 2017).

An important modelling difficulty for MCAOs is that clouds and precipitation not only respond to surface forcings but also feed back to the evolving air mass. Precipitation development can alter the cloud structure by removing condensate, modifying sub-cloud thermodynamics and vertical mixing, and helping to promote transitions between overcast, broken, and cellular cloud regimes (Lebo et al., 2008; Tornow et al., 2021; Juliano et al., 2024; Wu et al., 2025). Thus, small errors in representing precipitation, cloud-top cooling, or cloud fraction can result in large deviations in simulated MCAO evolution (Tomassini et al., 2017; de Roode et al., 2019).

The presence of Arctic mixed-phase stratocumulus clouds in MCAOs complicates the modelling further. Accurately modelling their persistence requires careful representation of microphysical processes, radiative cooling, and mixing (Morrison et al., 2005; Lebo et al., 2008; Solomon et al., 2011; de Boer et al., 2011; Morrison et al., 2011). Bulk microphysical schemes are widely used, but simulated cloud phase and lifetime remain highly sensitive to uncertainties in ice nucleation, depositional growth, sedimentation, secondary ice production, and ice habit; small errors can therefore lead to unrealistic glaciation and radiative biases (Fridlind et al., 2007; Klein et al., 2009; Morrison et al., 2011, 2020; Das et al., 2025).

As discussed in Section 1.2.1, differences in large-scale subsidence have been shown to influence the initial development of MCAOs (Brümmer, 1997; McCoy et al., 2017; Young et al., 2018; Schirmacher et al., 2024). During such air mass transformations, subsidence acts as a large-scale dynamical forcing that can modulate the evolving boundary layer, but has been found to produce contrasting effects on cloud evolution depending on the cloud and boundary layer state. In some surface-coupled mixed-phase layers, it can strengthen the capping inversion and enhance cloud-top radiative cooling and overturning, supporting cloud persistence and precipitation formation, whereas in stronger cases, sudden subsidence can force the inversion downward faster than the cloud layer can adjust, leading to rapid cloud dissipa-

tion and reduced surface downwelling longwave radiation (Young et al., 2018; Neggers et al., 2019).

Idealised and high-resolution simulations indicate that subsidence can control mixed-phase cloud lifetime and glaciation susceptibility by modifying boundary layer depth and thermodynamic conditions for ice formation, with stronger subsidence reducing the ice-nuclei concentrations required for cloud dissipation (Das et al., 2025). Because these interactions span scales from mesoscale forcings to turbulence and microphysics, and because subsidence remains difficult to constrain observationally, it represents an important source of uncertainty in numerical simulations (Wendisch et al., 2024).

1.4 SCOPE OF THIS THESIS

The challenges outlined in section 1.3.6 highlight the importance of mesoscale subsidence in representing MCAOs in LES. While subsidence is known to influence Arctic MPCs, its role in MCAO development remains insufficiently understood and poorly constrained due to limited observations, particularly in high-latitude regions. This thesis aims to address these gaps by combining observational estimates of subsidence from dropsonde measurements with LES, using observation-based subsidence profiles as large-scale forcing and systematically varying the subsidence forcing to isolate its effects.

1.4.1 RESEARCH OBJECTIVES

As highlighted in Section 1.3.6, subsidence plays an important role in shaping the thermodynamic structure of the Arctic atmosphere, influencing boundary layer stability, humidity, and cloud evolution. Despite its importance, subsidence remains difficult to observe directly due to its small magnitude and strong variability. This difficulty has led to the development of indirect estimation methods, established in subtropical studies, that rely on divergence measurements.

Therefore, the first research question of this study is:

HOW CAN ATMOSPHERIC SUBSIDENCE BE QUANTIFIED IN THE
ARCTIC USING OBSERVATIONS?

This question is addressed in Chapter 4, through analysing airborne measurements obtained during the HALO- $(\mathcal{AC})^3$ campaign (High Altitude and Long Range Aircraft (HALO); Arctic Amplification: Climate Relevant Atmospheric and Surface Processes and Feedback Mechanisms $(\mathcal{AC})^3$; further introduced in Section 3.1). Using dropsonde observations along mesoscale flight patterns, vertical profiles of horizontal divergence and, consequently, subsidence are derived from mass continuity. The analysis further evaluates the applicability and robustness of these methods under Arctic conditions. By assessing methodological uncertainties and sensitivities, this study offers a framework for constraining subsidence in the Arctic by observations and establishes a basis for its use in LES applications.

Building on these observations, the second research question explores its dynamical implications for LES modelling:

**CAN A LAGRANGIAN LARGE EDDY SIMULATION, CONSTRAINED
WITH OBSERVATION-DERIVED LARGE-SCALE FORCINGS SUCH AS
SUBSIDENCE, REALISTICALLY REPRODUCE AN OBSERVED
MARINE COLD AIR OUTBREAK?**

To address this, in Chapter 5, a Lagrangian LES is conducted along the observed air mass trajectory, initialised and constrained using the extensive dataset obtained during HALO- $(\mathcal{AC})^3$. The simulations incorporate observation-derived large-scale forcings, including subsidence established from the first research question, enabling a consistent representation of the evolving large-scale environment. The model performance is evaluated against independent airborne and remote-sensing observations, focusing on key characteristics of the MCAO, including the thermodynamic boundary layer structure, cloud evolution, and integrated water paths.

Following the successful evaluation of the Lagrangian LES, the third research question focuses on the process-level impacts of subsidence on cloud and boundary layer dynamics:

**WHAT IS THE IMPACT OF ATMOSPHERIC SUBSIDENCE ON CLOUD
GLACIATION AND BOUNDARY LAYER DECOUPLING DURING
ARCTIC MARINE COLD AIR OUTBREAKS?**

A series of sensitivity experiments is conducted, systematically varying the magnitude of observation-derived subsidence while retaining its temporal and vertical structure. This isolates the influence of mesoscale vertical motion on the coupled evolution of the atmospheric boundary layer and mixed-phase cloud properties along the air mass trajectory. The analysis focuses on how variations in subsidence affect the development of the atmospheric boundary layer, the onset and progression of decoupling, and the associated changes in cloud structure and phase composition along the air mass trajectory.

After assessing subsidence effects on the boundary layer evolution, the final research question extends the analysis to include aerosol-cloud interactions and their relative importance:

HOW DOES THE INFLUENCE OF SUBSIDENCE ON CLOUD
EVOLUTION COMPARE TO THE EFFECTS OF THE
CONCENTRATION OF CLOUD CONDENSATION NUCLEI?

The Lagrangian LES are extended in Chapter 6, by incorporating sensitivity experiments using different CCN concentrations, representing conditions from pristine to continental air masses. These simulations are combined with the previously analysed subsidence configurations, enabling a systematic comparison between dynamical forcing and aerosol-driven microphysical effects. This setup examines how large-scale subsidence interacts with CCN-driven microphysical processes along the air mass trajectory and focuses on changes in droplet numbers and sizes, effects on precipitation formation, and phase partitioning due to aerosol concentrations, and how these interact with the dynamical constraints imposed by subsidence.

Before addressing these research questions, the methods used in this study are introduced in more detail. Chapter 2 outlines the conceptual and technical foundations of modelling MCAOs, and Section 3.2 introduces the research campaign, describes the MCAO case studied in this thesis, and presents the observational method to observe subsidence used throughout this work. Subsequently, the research questions are addressed in dedicated chapters: question one in Chapter 4, questions two and three in Chapter 5, and question four in Chapter 6. The concluding chapter Chapter 7 synthesises the main findings and provides an outlook on the implications of this thesis for future research.

2

Modelling Marine Cold Air Outbreaks

As described in Section 1.2, MCAO are characterised by strong instability in the ABL due to the pronounced temperature contrast between relatively warm open ocean and near-surface Arctic air. This contrast drives intense sensible and latent heat fluxes, strong turbulent mixing, and rapid boundary layer growth (Renfrew and Moore, 1999; Li et al., 2023; Dahlke et al., 2022). The resulting motions span scales from small-scale turbulence to mesoscale structures such as rolls and cellular convection (Etling and Brown, 1993; Gryschka and Raasch, 2005). Owing to this multiscale and transient, Lagrangian nature, disentangling the interactions between dynamical, thermodynamical, and microphysical processes from observations alone remains challenging (Papritz and Grams, 2018; Dahlke et al., 2022; Wendisch et al., 2024).

Atmospheric modelling has therefore been central in advancing the understanding of MCAOs by enabling controlled investigations of coupled processes. Early studies using idealised, mesoscale, and large eddy models established the foundations of boundary layer dynamics and organised convection, including roll convection and cloud streets (e.g., Mason,

1985; Shapiro et al., 1987b; Huang and Raman, 1991; Raasch, 1990). Later work clarified roll dynamics and their evolution into cellular convection (e.g., Etling and Brown, 1993; Müller and Chlond, 1996; Müller et al., 1999), highlighting the importance of turbulent transport and surface heterogeneity such as sea-ice (e.g., Liu et al., 2006). With advances in LES, studies increasingly resolved turbulence and mesoscale organisation, showing that roll convection naturally emerges in strongly forced marine boundary layers and plays a key role in vertical transport (e.g., Gryschka and Raasch, 2005; Gryschka et al., 2008; Skillingstad and Edson, 2009; Gryschka et al., 2014).

More recent work has expanded simulation setups to include MPCs and precipitation feedbacks (e.g., Abel et al., 2017; Tornow et al., 2021; Wu et al., 2025). Studies further emphasise the roles of model resolution, process coupling, and microphysical complexity, as coherent turbulent structures and mesoscale organisation strongly affect vertical exchange processes (Brilouet et al., 2020) and grey-zone modelling highlights representation challenges (Tomassini et al., 2017; de Roode et al., 2019; Honnert et al., 2020). Model comparisons with observations and intercomparison studies reveal strong sensitivities to large-scale forcing and uncertainties in turbulence closure, cloud structure, and the representation of microphysics (Li et al., 2021, 2023; Juliano et al., 2024; Wendisch et al., 2025; Juliano et al., 2026).

The following chapter introduces the governing equations on which atmospheric models are based (Section 2.1). It then provides a brief overview of atmospheric turbulence (Section 2.2) and outlines different approaches to atmospheric modelling (Section 2.3). Building on this foundation, the techniques used in LES are discussed in greater detail (Section 2.4). Subsequently, Lagrangian LES are introduced (Section 2.5), followed by a description of the specific atmospheric model employed in this thesis (Section 2.6).

2.1 GOVERNING EQUATIONS

To describe fluid motion, atmospheric models solve a set of fundamental conservation equations of dynamics and thermodynamics. These equations express the conservation of mass, momentum, and energy and provide the physical basis for representing atmospheric flows across scales. The following section summarises the governing equations relevant for this study and is largely based on Stull (1988) and van Heerwaarden et al. (2017).

Since atmospheric variability is generally stronger in the vertical than in the horizontal, vertical motion, which is largely driven by buoyancy, plays a central role in atmospheric dynamics and modelling. To isolate the dynamically relevant, buoyancy-driven motions from fast acoustic disturbances, the governing equations can be formulated under the anelastic approximation (Ogura and Phillips, 1962). Therefore, density ϱ , pressure p , and thermodynamic variables, such as the virtual potential temperature θ_v , are represented as small perturbations of a hydrostatic reference state that depends only on height z , such as

$$\varrho = \varrho_0(z) + \varrho', \quad p = p_0(z) + p', \quad \theta_v = \theta_{v0}(z) + \theta_v'. \quad (2.1)$$

The anelastic formulation retains the vertical variation of the reference density, while excluding sound-wave modes. A frequently used special case is the Boussinesq approximation, which is recovered when the reference density is assumed constant with height (Boussinesq, 1897). In this limit, the anelastic equations simplify to the incompressible form while buoyancy remains represented through density or virtual potential temperature perturbations (Randall, 2013).

2.1.1 IDEAL GAS LAW

The atmosphere can be approximated as an ideal gas composed of particles that move randomly and interact primarily through elastic collisions. Under this assumption, pressure, density, and temperature are related by the ideal gas law

$$p = \varrho \mathcal{R} T_v. \quad (2.2)$$

Here, $\mathcal{R} = 287 \text{ J K}^{-1} \text{ kg}^{-1}$ is the gas constant for dry air, and T_v is the virtual temperature accounting for the lower density of moist air compared to dry air.

Using the anelastic approximation, with a scaling height for the density $\mathcal{H}_\varrho$, based on a reference density profile ϱ_0

$$\mathcal{H}_\varrho \equiv \left(\frac{1}{\varrho_0} \frac{d\varrho_0}{dz} \right)^{-1}, \quad (2.3)$$

and the definition of the virtual potential temperature

$$\theta_v = T_v \left(\frac{p_0}{p} \right)^{\frac{R}{c_p}}, \quad (2.4)$$

the ideal gas law can be expressed as

$$\frac{\theta_v'}{\theta_{v0}} = \frac{p'}{\varrho_0 g \mathcal{H}_\varrho} - \frac{\varrho'}{\varrho_0}. \quad (2.5)$$

2.1.2 CONTINUITY EQUATION

In classical physics, mass cannot be created or destroyed; therefore, it must be conserved. In fluid dynamics, this principle is expressed by the continuity equation. In the following, the Einstein summation convention is used, where indices $i, j, k \in \{1, 2, 3\}$ are summed over whenever they appear twice. The continuity equation then reads

$$\frac{d\varrho}{dt} + \varrho \frac{\partial u_i}{\partial x_i} = 0, \quad (2.6)$$

with the three dimensional wind components u_i . The continuity equation thus states that all mass needs to be either transported in and out of a parcel of air or alter its density.

Under the anelastic approximation, density perturbations are neglected in the continuity equation, but the vertical variation of the reference density is retained. Mass conservation, therefore, becomes

$$\frac{\partial(\varrho_0 u_i)}{\partial x_i} = 0. \quad (2.7)$$

2.1.3 NAVIER-STOKES EQUATIONS

The motion of an air parcel is governed by the Navier-Stokes equations, originally formulated by [Navier \(1821, 1827\)](#) and [Stokes \(1845\)](#) by applying Newton's second law to fluid motion. The resulting momentum equation is given by

$$\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} = -\delta_{i3}g - 2\epsilon_{ijk}\Omega_j u_k - \frac{1}{\varrho} \frac{\partial p}{\partial x_i} + \frac{1}{\varrho} \frac{\partial \tau_{ij}}{\partial x_j}. \quad (2.8)$$

The first term $\frac{\partial u_i}{\partial t}$ describes the inertia of the air parcel, while the second term $u_j \frac{\partial u_i}{\partial x_j}$ represents nonlinear advection. The term $-\delta_{i3}g$ accounts for the gravitational acceleration, where δ_{ij} denotes the Kronecker-Delta (for the definition see page 152). The Coriolis acceleration arising from Earth's rotation is given by $-2\epsilon_{ijk}\Omega_j u_k$, with ϵ_{ijk} , the Levi-Civita Symbol (for the definition see page 152). The pressure-gradient force is expressed by $-\frac{1}{\rho} \frac{\partial p}{\partial x_i}$ and drives the flow toward dynamical equilibrium. The final term $\frac{1}{\rho} \frac{\partial \tau_{ij}}{\partial x_j}$ represents viscous stresses caused by molecular momentum transfer. Assuming an incompressible Newtonian fluid, the viscous stress tensor reduces to $\nu \frac{\partial^2 u_i}{\partial x_j^2}$, where ν denotes the kinematic viscosity. Under the anelastic approximation, and assuming that the reference state satisfies the hydrostatic balance,

$$\frac{dp_0}{dz} = -\rho_0 g , \quad (2.9)$$

together with the continuity equation Equation (2.7), the momentum equation becomes

$$\frac{\partial u_i}{\partial t} = -\frac{1}{\rho_0} \frac{\partial(\rho_0 u_i u_j)}{\partial x_j} - \frac{\partial}{\partial x_i} \left(\frac{p'}{\rho_0} \right) + \delta_{i3} g \frac{\theta_v'}{\theta_{v0}} - 2\epsilon_{ijk} \Omega_j u_k + \nu \frac{\partial^2 u_i}{\partial x_j^2} . \quad (2.10)$$

2.1.4 FIRST LAW OF THERMODYNAMICS

The first law of thermodynamics states that energy is conserved during thermodynamic processes. In atmospheric flows, it governs the evolution of sensible and latent heat through advection, diffusion, radiation, and phase changes. Using potential temperature θ as a prognostic thermodynamic variable, the conservation equation can be written as

$$\frac{\partial \theta}{\partial t} + u_i \frac{\partial \theta}{\partial x_i} = \nu_\theta \frac{\partial^2 \theta}{\partial x_i^2} + S_\theta , \quad (2.11)$$

where S_θ represents diabatic sources and sinks of heat, including radiative heating and latent heat release. The term on the left-hand side describes the local tendency and advection of potential temperature, while the diffusion term on the right-hand side accounts for molecular heat transport.

Under the anelastic approximation, the thermodynamic equation becomes

$$\frac{\partial \theta}{\partial t} = -\frac{1}{\varrho_0} \frac{\partial (\varrho_0 u_i \theta)}{\partial x_i} + \kappa_\theta \frac{\partial^2 \theta}{\partial x_i^2} + \frac{\theta_0}{\varrho_0 c_p T_0} Q . \quad (2.12)$$

Here, κ_θ denotes the thermal diffusivity, and Q represents external sources and sinks of heat.

2.1.5 CONSERVATION EQUATION FOR SCALARS

Analogous to the conservation of potential temperature expressed by the first law of thermodynamics, other atmospheric quantities, such as humidity and trace gas concentrations, obey scalar conservation equations. An arbitrary atmospheric scalar with a concentration of ξ evolves according to

$$\frac{\partial \xi}{\partial t} + u_i \frac{\partial \xi}{\partial x_i} = \nu_\xi \frac{\partial^2 \xi}{\partial x_i^2} + S_\xi . \quad (2.13)$$

The first term represents temporal storage, while $u_i \frac{\partial \xi}{\partial x_i}$ describes advection by the flow. Molecular diffusion is represented by $\nu_\xi \frac{\partial^2 \xi}{\partial x_i^2}$, where ν_ξ denotes the scalar diffusivity. The source-sink term S_ξ accounts for all remaining processes affecting the scalar, such as chemical reactions, emissions, deposition or phase transitions.

2.2 ATMOSPHERIC TURBULENCE

Turbulent flows are characterised by irregular, random fluctuations in velocity and other flow properties over a wide range of spatial and temporal scales. Despite this apparent disorder, turbulent flows exhibit reproducible statistical behaviour, allowing them to be described in a probabilistic sense. A key feature of turbulence is the presence of motions of multiple scales, from large energy-containing structures influenced by geometry and boundary conditions to small dissipative motions governed primarily by viscosity. These multiscale interactions are responsible for the enhanced transport and mixing of momentum, heat, and scalar quantities observed in turbulent flows (Pope, 2001).

2.2.1 KOLMOGOROV HYPOTHESIS

Turbulence consists of eddies of different sizes, with energy entering at the largest scales and cascading to progressively smaller eddies until it is dissipated by viscous effects at the small-

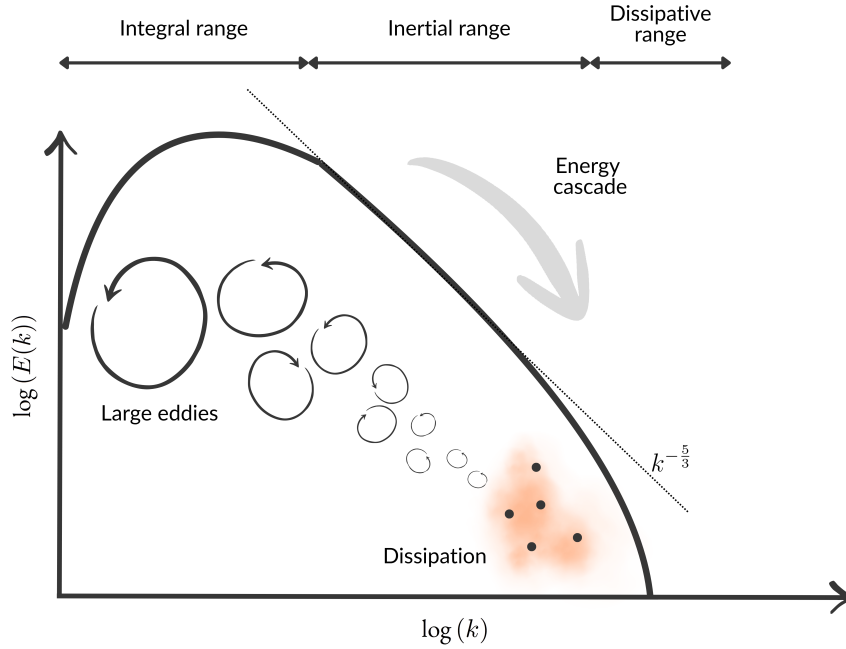


Figure 2.1: Schematic of the kinetic energy spectrum $E(k)$ as a function of the wave number k with the integral range describing large eddies, the inertial range with the characteristic scaling indicated as a dashed line and the dissipative range where the energy is transferred to heat (indicated in red).

est scales and converts into heat. These processes of turbulent kinetic energy transfer across different scales of turbulent motion are described by the Kolmogorov theory. A schematic of the corresponding energy spectrum $E(k)$ is shown as a function of the wave number k in Figure 2.1.

The largest eddies in the integral range at length scales comparable to the characteristic flow dimension contain most of the turbulent kinetic energy. They are strongly influenced by the flow geometry and boundary conditions. The Kolmogorov hypothesis states that small-scale turbulent motions become approximately statistically isotropic at sufficiently high Reynolds numbers, even if the large-scale flow is anisotropic. This occurs because the cascade process transfers energy through many stages of eddy interactions, gradually removing directional biases introduced by the large scales. As a result, the smallest turbulent motions lose information about the specific geometry of the flow and exhibit universal statistical behaviour.

The small-scale statistics, therefore, become independent of the large-scale geometry and boundary conditions. The range of scales where viscosity has negligible influence and en-

ergy is transferred between scales at an approximately constant rate equal to the dissipation rate ϵ is called the inertial subrange. Using dimensional analysis, Kolmogorov argues that the characteristic scaling relation for eddies of wave number k in this region is

$$E(k) \propto \epsilon^{\frac{2}{3}} k^{-\frac{5}{3}}. \quad (2.14)$$

For turbulence modelling, this universality of small-scale turbulence allows the unresolved small scales to be represented through appropriate modelling approaches, where the turbulence spectrum is effectively separated into resolved large energy-containing motions and smaller scales whose influence is incorporated through models based on their assumed universal behaviour.

2.3 APPROACHES TO ATMOSPHERIC MODELLING

Due to the turbulent nature of the atmospheric flows, exhibiting a wide range of interacting spatial and temporal scales, it is generally not feasible to resolve all relevant scales explicitly in computational models. Instead, different modelling approaches have been developed that trade off physical fidelity and computational cost by either resolving or parameterising turbulent processes (Pope, 2001). These approaches range from highly idealised, fully resolved simulations to coarse-resolution models used for weather prediction and climate studies.

DIRECT NUMERICAL SIMULATION

Direct numerical simulations (DNS) directly solves the governing equations (Section 2.1) without introducing turbulence parameterisations. All turbulent scales are resolved, from large eddies down to the smallest dissipative motions where kinetic energy is converted into heat. DNS therefore provides a detailed representation of turbulence but quickly becomes extremely computationally expensive. Since the required computational effort increases rapidly with Reynolds number, DNS is generally limited to small domains and relatively low-Reynolds-number flows (Pope, 2001; Mellado, 2012).

GENERAL CIRCULATION MODEL

General circulation models (GCMs) simulate the large-scale atmospheric dynamics with coarser spatial resolution. The governing equations are solved on grids that are much larger

than the dominant turbulent eddies of the ABL. Consequently, turbulent motions are not explicitly resolved but represented through tuned parameterisations of specific subgrid-scale processes, such as cloud formation (Kochkov et al., 2024). Due to their low resolution, GCMs can perform simulations over large areas and long periods of time with comparatively little computational effort and are well suited to long-term climate simulations and operational weather forecasting (Holton and Hakim, 2013).

LARGE EDDY SIMULATION

Large eddy simulations bridge the gap between DNS and GCM approaches. They explicitly resolve the large, energy-containing turbulent motions in the ABL while parametrising smaller eddies. Due to the fact that most turbulent kinetic energy resides in large-scale structures influenced by boundary conditions, while smaller eddies exhibit a more universal behaviour, LES is designed to capture the essential turbulent dynamics at substantially reduced computational cost compared to DNS, while also reducing the impact of turbulence closure compared to GCMs. LES provides three-dimensional, time-dependent flow fields and is therefore widely used to study ABL processes and cloud dynamics and can be used to develop and evaluate parameterisations for large-scale atmospheric models (Pope, 2001).

Therefore, LES is adopted in this thesis because it resolves the dominant turbulent structures in MCAOs while remaining computationally feasible, allowing a physically consistent representation of boundary layer evolution under prescribed large-scale conditions. A comprehensive description of its formulation is given in Section 2.4.

2.4 LARGE EDDY EQUATIONS

The LES approach is based on a scale separation achieved through spatial filtering of the governing equations. This filtering operation decomposes the flow into resolved large-scale motions and unresolved subgrid-scale contributions. While the resolved scales are computed explicitly, the effect of the unresolved motions appears as additional terms in the filtered equations and must be parameterised. The following section will introduce the most important principles and parameterisations and is largely based on (Pope, 2001) and (Heus et al., 2010), unless stated otherwise.

2.4.1 FILTERING

The filtering operation was introduced by [Leonard \(1975\)](#) and is performed by convolving an arbitrary scalar field ξ with a filter function $G(\vec{r}, \vec{x})$

$$\tilde{\xi}(\vec{x}, t) = \int G(\vec{r}, \vec{x}) \xi(\vec{x} - \vec{r}, t) d\vec{r}, \quad (2.15)$$

under the condition that the filter function satisfies the normalisation condition

$$\int G(\vec{r}, \vec{x}) d\vec{r} = 1. \quad (2.16)$$

The filter width is typically related to the grid spacing and therefore determines the smallest scales that can be explicitly resolved within the LES. The true field of ξ can then be decomposed into a resolved component $\tilde{\xi}$ and a residual component ξ^r according to

$$\xi = \tilde{\xi} + \xi^r. \quad (2.17)$$

In contrast to Reynolds decomposition, both contributions represent instantaneous flow fields associated with different turbulent length scales rather than mean and fluctuating quantities. The resolved field contains motions larger than the filter width, while the residual field represents the unresolved subgrid-scale motions. A schematic example of a filtered spectrum is shown in Figure 2.2.

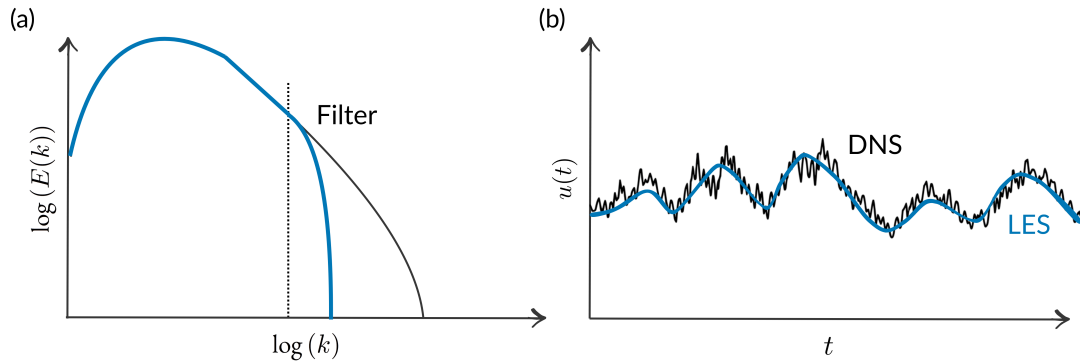


Figure 2.2: Schematic of filtering in LES (blue) for the energy spectrum $E(k)$ (grey; adapted from [Pope, 2001](#)); a) and for a filtered (blue) horizontal wind field $u(t)$ (grey; b).

2.4.2 FILTERED GOVERNING EQUATIONS

Considering spatially uniform filters, spatial derivatives commute with the filtering operation, such that

$$\frac{\partial \widetilde{\xi}}{\partial x_i} = \widetilde{\frac{\partial \xi}{\partial x_i}}, \quad \frac{\partial^2 \widetilde{\xi}}{\partial x_i^2} = \widetilde{\frac{\partial^2 \xi}{\partial x_i^2}}. \quad (2.18)$$

The filtered continuity equation Equation (2.7) is

$$\frac{\partial \widetilde{\varrho_0 u_i}}{\partial x_i} = 0. \quad (2.19)$$

Applying the filter to the momentum equation Equation (2.10) gives

$$\frac{\partial \widetilde{u_i}}{\partial t} = -\frac{1}{\varrho_0} \frac{\partial(\varrho_0 \widetilde{u_i u_j})}{\partial x_j} - \frac{\partial}{\partial x_i} \left(\frac{\widetilde{p}}{\varrho_0} \right) + \delta_{i3} g \frac{\widetilde{\theta_v}}{\theta_{v0}} - 2\epsilon_{ijk} \Omega_j \widetilde{u_k} + \frac{\partial^2 \widetilde{u_i}}{\partial x_j^2}. \quad (2.20)$$

The nonlinear advection term introduces correlations between resolved and unresolved scales. To represent these subgrid contributions, a residual stress tensor is defined as

$$\tau_{ij}^R \equiv \widetilde{u_i u_j} - \widetilde{u_i} \widetilde{u_j}. \quad (2.21)$$

This tensor can be decomposed into an isotropic τ_{ii}^R and an anisotropic τ_{ij}^r component. The residual kinetic energy is defined by

$$e = \frac{1}{2} \tau_{ii}^R. \quad (2.22)$$

The remaining anisotropic residual stress tensor is then given by

$$\tau_{ij}^r \equiv \tau_{ij}^R - \frac{2}{3} \delta_{ij} e. \quad (2.23)$$

Substituting this decomposition into Equation (2.20) results in the momentum equation

$$\frac{\partial \widetilde{u_i}}{\partial t} = -\frac{1}{\varrho_0} \frac{\partial(\varrho_0 \widetilde{u_i u_j})}{\partial x_j} - \frac{\partial}{\partial x_i} \left(\frac{\widetilde{p}}{\varrho_0} \right) + \delta_{i3} g \frac{\widetilde{\theta_v}}{\theta_{v0}} - 2\epsilon_{ijk} \Omega_j \widetilde{u_k} + \frac{\partial^2 \widetilde{u_i}}{\partial x_j^2} - \frac{1}{\varrho_0} \frac{\partial(\varrho_0 \tau_{ij}^r)}{\partial x_j}. \quad (2.24)$$

Finally, applying the filtering operation to the scalar conservation equation and defining the residual scalar flux

$$R_{u_i,\xi} \equiv \widetilde{u_i\xi} + \widetilde{u_i}\widetilde{\xi}, \quad (2.25)$$

yields

$$\frac{\partial \widetilde{\xi}}{\partial t} + \frac{\partial \widetilde{u_i}\widetilde{\xi}}{\partial x_i} = -\frac{\partial R_{u_i,\xi}}{\partial x_i} + \nu_\xi \frac{\partial^2 \widetilde{\xi}}{\partial x_i^2} + S_\xi. \quad (2.26)$$

The anisotropic residual stress tensor τ_{ij}^r and the residual scalar fluxes $R_{u_i,\xi}$ remain unresolved within the LES formulation and must therefore be parametrised.

2.4.3 SUBFILTER-SCALE CLOSURE

The closure of the residual stress tensor τ_{ij}^r and scalar fluxes $R_{u_i,\xi}$ can be modelled through an eddy-diffusivity approach as described by [Heus et al. \(2010\)](#), whose notation is followed throughout this section. The deviatoric subfilter stress and scalar flux are written as

$$\tau_{ij}^r = -K_m \left(\frac{\partial \widetilde{u_i}}{\partial x_j} + \frac{\partial \widetilde{u_j}}{\partial x_i} \right) \quad \text{and} \quad R_{u_i,\xi} = -K_h \frac{\partial \widetilde{\xi}}{\partial x_j}, \quad (2.27)$$

with the eddy viscosity coefficient K_m and the eddy diffusivity coefficient K_h ([Heus et al., 2010](#)).

According to [Smagorinsky \(1963\)](#) and later LES formulations ([Wyngaard, 2004](#); [Pope, 2001](#)), the eddy viscosity can be modelled algebraically using the Smagorinsky closure,

$$K_m = c_s^2 \lambda^2 (2S_{ij}S_{ij})^{\frac{1}{2}}, \quad (2.28)$$

where c_s denotes the Smagorinsky constant, λ a mixing length, and the resolved strain-rate tensor is defined as

$$S_{ij} = \frac{1}{2} \left(\frac{\partial \widetilde{u_j}}{\partial x_i} + \frac{\partial \widetilde{u_i}}{\partial x_j} \right). \quad (2.29)$$

In this algebraic closure, the subgrid turbulence is obtained from the local strain rate, such that the modelled subfilter dissipation depends solely on resolved velocity gradients ([Pope, 2001](#)). In contrast, the prognostic subfilter-scale turbulent kinetic energy (SFS-TKE) model by [Deardorff \(1972\)](#) determines the mixing coefficients from a transport equation for the subfilter energy e rather than from the instantaneous strain rate.

As a result, the turbulence intensity evolves dynamically and reflects transport effects. The eddy viscosity and eddy diffusivity are then defined as

$$K_m = c_m \lambda e^{1/2} \quad \text{and} \quad K_h = c_h \lambda e^{1/2}, \quad (2.30)$$

where c_m and c_h are constants and λ denotes the mixing length (Heus et al., 2010). In contrast to the Smagorinsky closure, this formulation introduces an additional prognostic equation for e ,

$$\frac{\partial e}{\partial t} = -\frac{\partial \tilde{u}_j e}{\partial x_j} - \tau_{ij} \frac{\partial \tilde{u}_i}{\partial x_j} + \frac{g}{\theta} R_{w,\theta_v} - \frac{\partial R_{u_j,e}}{\partial x_j} - \frac{1}{\bar{\rho}} \frac{\partial R_{u_j,\pi}}{\partial x_j} - \epsilon, \quad (2.31)$$

with ϵ the SFS-TKE dissipation rate (Heus et al., 2010). The additional buoyancy-production term explicitly represents the generation of turbulence by vertical heat fluxes, making the Deardorff closure particularly suitable for atmospheric boundary layer simulations where convection is important.

2.4.4 SURFACE MODEL

Atmospheric turbulence is strongly influenced by the lower boundary conditions. However, the viscous surface fluxes are not resolved in LES and need to be parametrised as boundary fluxes into the surface layer. Following Heus et al. (2010), the subfilter-scale surface flux of an arbitrary variable ξ is written as

$$F_{s\xi} = \widetilde{w\xi} - \widetilde{w}\widetilde{\xi}, \quad (2.32)$$

representing the unresolved turbulent exchange between the surface and the atmosphere (Heus et al., 2010). The theoretical basis of most atmospheric surface flux parameterisations is Monin–Obukhov similarity theory (MOST), introduced by Monin and Obukhov (1954), who consider horizontally homogeneous flow over a rough surface and assume a constant-flux surface layer, where the turbulent momentum flux (stress) is defined as

$$\tau = -\overline{\rho u'w'}, \quad (2.33)$$

with u' and w' denoting fluctuations of horizontal and vertical wind components. From this stress, a friction velocity is introduced

$$u_* = \sqrt{\frac{\tau}{\rho}}, \quad (2.34)$$

which provides the characteristic velocity scale governing turbulence in the surface layer (Monin and Obukhov, 1954).

A central result of MOST is that nondimensional turbulence statistics depend only on the stability parameter

$$\zeta = \frac{z}{\mathcal{L}}, \quad (2.35)$$

where z is the height above the surface and $\mathcal{L}$ is the Obukhov length

$$\mathcal{L} = -\frac{u_*^3 \overline{\theta_v}}{\kappa g (\overline{w' \theta_v'})_s}, \quad (2.36)$$

where $\overline{w' \theta_v'}_s$ is the surface buoyancy flux and κ is the von Kármán constant; this length expresses the relative importance of shear and buoyancy forcing (Monin and Obukhov, 1954; Heus et al., 2010). Under neutral stratification, the similarity arguments lead to a logarithmic wind profile (Monin and Obukhov, 1954).

Building on these similarity relations, Louis (1979) formulates a bulk transfer scheme by integrating the MOST flux-profile relations between the roughness length and the lowest model level. In modern LES implementations (e.g., Heus et al., 2010), these integrated similarity relations are expressed through drag coefficients,

$$C_\xi = \frac{u_{*0} \xi_{*0}}{\langle u_1 \rangle \langle \tilde{\xi}_1 - \xi_0 \rangle}, \quad (2.37)$$

with u_1 being the magnitude of the horizontal wind speed at the first model level, which yields local surface fluxes

$$F_{s\xi} = -C_\xi \langle u_1 \rangle (\tilde{\xi}_1 - \xi_0). \quad (2.38)$$

These fluxes act as a subfilter-scale boundary model, ensuring that horizontally averaged flux-gradient relationships remain consistent with MOST while coupling the resolved LES flow to the unresolved surface layer (Heus et al., 2010).

2.4.5 PARAMETRISATION OF MICROPHYSICS

The representation of cloud microphysics in LES is essential for simulating cloud processes and precipitation. These processes occur at very small spatial scales, far smaller than the typical modelling grid. Therefore, the creation and evolution of hydrometeors, such as cloud droplets and ice crystals, need to be parametrised. Generally, there are two main approaches for microphysical parametrisations to represent particle populations. Bin-microphysics schemes explicitly resolve the particle size distribution by dividing it into bins and predicting the mass or number of particles in each bin. This approach provides a detailed representation of hydrometeors and processes, but is computationally expensive. Alternatively, models use bulk schemes that represent the particle size distribution using a limited set of statistical moments, such as the mass and number densities of hydrometeors, to reduce computational costs (Morrison et al., 2020).

A commonly used microphysical model was developed by Seifert and Beheng (2006); a two-moment bulk scheme to represent the mass and number densities of five hydrometeors: cloud liquid, rain, cloud ice, snow and graupel. The following section is based on this scheme and follows the corresponding paper by Seifert and Beheng (2006).

In this two-moment parameterisation, the hydrometeor size distribution $f(x)$ is a function of the particle mass x and is characterised by its moments

$$M_{\xi}^k = \int x^k f_{\xi}(x) \, dx , \quad (2.39)$$

which corresponds to the number density for $k = 0$ and to the mass density for $k = 1$ for each hydrometeor class ξ of cloud droplets, rain, ice crystals, snow and graupel. The time evolution of each moment follows a transport equation of scalars (Equation (2.13)).

For liquid hydrometeors, droplet formation through activation of CCN is represented as a function of supersaturation and aerosol properties, consistent with the processes outlined in Section 1.3.3. Subsequent condensational growth is treated using a saturation-adjustment

approach, ensuring that excess water vapour is instantaneously converted into liquid water while keeping the droplet number constant. The droplet population is partitioned into cloud droplets and raindrops using a critical size threshold, reflecting the transition from diffusion-dominated growth to collision-driven processes. The formation of precipitation is parameterised through autoconversion, describing the initial growth of droplets into raindrops, and accretion, which represents the continued growth of raindrops by collecting smaller droplets. Interactions within and between hydrometeor classes are further accounted for through self-collection, which reduces particle number while conserving mass, and collisional breakup, which redistributes mass toward smaller sizes. The removal of liquid water from the cloud is represented by sedimentation via size-dependent fall velocities or evaporation.

For frozen hydrometeors, the parametrisation is extended to include cloud ice, snow, and graupel, whose evolution is likewise described by their number and mass moments. Ice formation is represented by adjusting the ice particle number concentration towards the available ice nuclei as a function of temperature and supersaturation, as introduced in Section 1.3.3, and by including freezing processes of supercooled droplets. The subsequent growth of ice particles occurs through vapour deposition and through collection processes that govern interactions between different hydrometeor types. These include aggregation between ice particles, leading to snow formation, and riming, in which ice particles collect supercooled droplets and may be converted into graupel depending on the accumulated rime mass. Collisions between rain and ice are also considered, contributing to mixed-phase interactions and phase transitions. The parametrisation further accounts for the melting of ice particles in warmer conditions and for their sedimentation using species-dependent fall velocities.

2.5 LAGRANGIAN LARGE EDDY SIMULATION

The atmosphere can be modelled within different reference frames. In an Eulerian frame (Figure 2.3 a), the model domain is fixed in space while the atmospheric flow moves through it. Thus in such LES, air masses pass through a stationary computational domain and upstream influences must be represented through boundary conditions and large-scale forcings.

When studying air mass transformations, it can be advantageous to adopt a Lagrangian perspective (Figure 2.3 b), which allows following the evolution of an air mass over longer

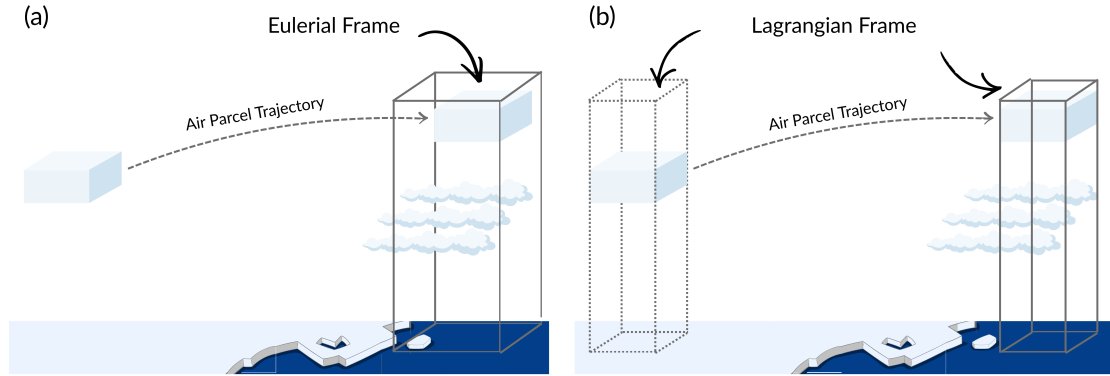


Figure 2.3: Schematic of different modelling approaches for LES, Eulerian frame fixed in space (a) and Lagrangian frame following an air parcel along the trajectory (b).

periods without requiring a substantial increase in domain size. Furthermore, the strong dependence of air mass transformations on the upstream history of the air mass can be represented more consistently.

In a Lagrangian LES, the simulation domain follows the movement of an air mass along its trajectory. The model domain effectively moves with the mean flow, allowing the internal processes within the air mass to evolve continuously along its path. In this configuration, the large-scale forcings are prescribed relative to the moving frame of reference. Consequently, the horizontal wind within the simulation domain is approximately zero at the trajectory height, while the corresponding Eulerian wind represents the propagation speed of the air mass trajectory. This approach makes Lagrangian sampling particularly well-suited for studying MCAOs, where the thermodynamic and dynamical evolution of an air mass strongly depends on its history along the trajectory. An introduction to the calculation of these trajectories will be given in Section 3.3.1.

2.6 THE DUTCH ATMOSPHERIC LARGE EDDY SIMULATION MODEL

To perform the LES experiments in this thesis, the Dutch Atmospheric Large Eddy Simulation (DALES) model is used. DALES is a turbulence-resolving model developed for studies of the atmospheric boundary layer. It was introduced by [Heus et al. \(2010\)](#), based on a model developed by [Nieuwstadt and Brost \(1986\)](#).

DALES solves the prognostic equations (see Section 2.1) for the three wind components u_i , the liquid water potential temperature θ_l and the subfilter-scale turbulent kinetic energy e , as well as the total humidity q_t , the CCN concentration and the number and mass densities of five hydrometeor species. It uses a SFS-TKE model by (Deardorff, 1972) (see Section 2.4.3) and applies MOST to calculate fluxes and gradients at the surface boundary. The model is run on an Arakawa C-grid (Arakawa and Lamb, 1977), with scalars defined at the cell centre and velocity components at cell faces (Heus et al., 2010).

For the representation of mixed-phase cloud in the Arctic, DALES has been extended beyond its original warm-cloud formulation to a mixed-phase double-moment bulk microphysics scheme based on Seifert and Beheng (2006) (see Section 2.4.5), predicting the mass and number concentrations of cloud droplets, rain, cloud ice, snow and graupel, adapted for high Arctic clouds (Chylik et al., 2023; Schnierstein et al., 2024). DALES can be applied in a Lagrangian framework, as demonstrated for Arctic conditions by Neggers et al. (2019), making it well suited for investigating Arctic air mass transformations such as MCAOs.

3

Observing Marine Cold Air Outbreaks

Marine cold air outbreaks in the Arctic often occur in remote regions, where sparse observations complicate the collection of high-quality measurement data characterising these events (Geerts et al., 2022). Observations of MCAOs ideally require coordinated measurements of surface fluxes, cloud properties, and microphysics from multiple platforms rather than single instruments (Renfrew and Moore, 1999; Geerts et al., 2022).

Early observational work focused primarily on turbulent convection and the roll structure of cloud streets during MCAOs. Aircraft observations made it possible to directly sample MCAOs using in-situ and remote sensing measurements and obtain vertical profiles from the deployment of dropsondes. Early research campaigns analysed roll vortices and boundary layer stability using observations from towers and aircraft, and combined aircraft and ship observations to investigate boundary layer development and organised convection (LeMone, 1973; Hein and Brown, 1988; Kellner et al., 1987). Further aircraft campaigns increasingly couple aircraft measurements with satellite observations to connect cloud structures to the evolving boundary layer (Hartmann et al., 1997; Brümmer, 1996, 1997; Renfrew and Moore,

1999). However, most early aircraft studies relied on localised transects and were therefore only able to cover small areas of MCAOs.

More recent campaigns increasingly rely on coordinated multi-platform observations combining aircraft, ground-based remote sensing, ships, and balloons. For example, the Cold air Outbreaks in the Marine Boundary Layer Experiment (COMBLE) provided sustained ground-based remote sensing observations of MCAO clouds in northern Scandinavia, enabling detailed studies of shallow convective clouds and boundary layer structure (Mages et al., 2023; Geerts et al., 2022; Wu et al., 2025; Juliano et al., 2024; Lackner et al., 2023). Observations from the Multidisciplinary drifting Observatory for the Study of Arctic Climate (MOSAiC) expedition have been combined with COMBLE measurements to analyse transport processes (Kirbus et al., 2023a; DeMott et al., 2025). Multiple aircraft campaigns have provided further observations of Arctic clouds and boundary layer processes during cold air outbreak conditions (e.g. Solomon et al., 2009; Verlinde et al., 2007; Knudsen et al., 2018; Seethala et al., 2021; Tornow et al., 2021; Michaelis et al., 2022; Tornow et al., 2023). Recent campaigns increasingly aim to follow the air masses along their path. For example, the Cold-Air outbreak Experiment in the Sub-ARctic (CAESAR), aimed to sample MCAOs from the sea-ice edge along the downstream flow in the North Atlantic (Zuidema et al., 2026). The HALO-(AC)³ campaign, which this thesis is based on, utilised multiple research aircraft at different altitudes to investigate air mass transformations during MCAOs and WAIs, including observations in the high Arctic above sea-ice (see Section 3.1; Klingebiel et al., 2025; Kirbus et al., 2024; Wendisch et al., 2024; Schirmacher et al., 2024; Walbröl et al., 2024). This setup is an ideal basis to study the evolution of MCAOs, providing upstream information and comprehensive remote-sensing and in-situ observations. The campaign setup is introduced further in the following Section 3.1.

3.1 HALO-(AC)³ CAMPAIGN

The HALO-(AC)³ campaign aimed to improve the understanding and representation of air mass transformation processes occurring during meridional transport between the Arctic and the mid-latitudes (Wendisch et al., 2021). The campaign was conducted in March and April 2022 and focused on the Fram Strait region and the adjacent Norwegian and Greenland Seas. Particular emphasis was placed on the investigation of northward WAIs and southward MCAOs, as outlined in the comprehensive campaign overview presented by Wendisch et al.

(2024). A detailed description of the instrument configuration and resulting datasets is provided by Ehrlich et al. (2025).

To observe both the large-scale evolution of air masses and small-scale boundary layer processes, three coordinated research aircraft were deployed and operated in partially co-located flight patterns at different altitudes. The HALO aircraft, operated by the German Aerospace Center (DLR), has a range of approximately 9000 km and a maximum flight altitude of about 15 km. It operated from Kiruna, Sweden (67.85 °N, 20.22 °E), and performed 17 research flights covering the Norwegian Sea, Greenland Sea, Fram Strait, and the central Arctic Ocean (Wendisch et al., 2024).

HALO was equipped with a suite of active and passive remote sensing instruments designed to observe the vertical structure of the troposphere, including clouds, water vapour, aerosols, precipitation, and surface properties (Wendisch et al., 2024). From this suite, this thesis incorporates observations from the Airborne Vertical Atmospheric Profiling System (AVAPS) system, which deploys Vaisala RD-41 dropsondes to obtain high-resolution vertical profiles of temperature, humidity, pressure, and horizontal wind. Furthermore, measurements from the HALO Microwave Package (HAMP) Ka-band Doppler cloud radar were used to observe cloud structures and precipitation profiles, while the accompanying multi-channel microwave radiometer retrieved atmospheric temperature, humidity, and liquid water content from measured microwave emissions (Mech et al., 2014). Additionally, the Video airborne Longwave Observations within siX channels (VELOX) thermal infrared camera system was employed and is used in this thesis to constrain the surface boundary conditions in LES; more information on this instrument is presented in Section 3.3.3.

In addition, the Polar 5 (P5) and Polar 6 (P6) aircraft, operated by the Alfred-Wegener-Institut (AWI), were based in Longyearbyen on Svalbard (78.24 °N, 15.49 °E). These aircraft typically flew at lower altitudes with a range of approximately 2300 km and a maximum altitude of around 6 km.

P5 primarily conducted active and passive remote sensing observations of clouds, precipitation, aerosols, and surface properties (Mech et al., 2019; Schirmacher et al., 2023). It was also equipped with an AVAPS dropsonde system (Vaisala, 2023). In this thesis, observations from the Microwave Radar/Radiometer for Arctic Clouds (MiRAC) microwave cloud radar and radiometer system were used to retrieve cloud properties (Mech et al., 2019). Additionally,

the Airborne Mobile Aerosol Lidar (AMALi), a multi-wavelength lidar system, provided vertical profiles of clouds and aerosols by detecting laser backscatter from atmospheric particles (Stachlewska et al., 2010). The P6 aircraft was primarily dedicated to in-situ measurements of aerosols, cloud microphysics, and trace gases in the lower troposphere. However, these observations are not utilised in this thesis.

3.1.1 QUASI-LAGRANGIAN OBSERVATIONS

Previous studies of Arctic air mass transformations primarily relied on Eulerian observations obtained at fixed locations, such as ground-based stations or ships, sometimes complemented by aircraft or satellite measurements (Uttal et al., 2002; Shupe et al., 2022; Kirbus et al., 2023a; Svensson et al., 2023). However, they cannot directly capture the temporal evolution of individual air masses. To overcome this limitation, the HALO- $(\mathcal{AC})^3$ campaign implemented a quasi-Lagrangian observation strategy that repeatedly sampled the same moving air mass along its trajectory in order to quantify transformation processes during transport (Wendisch et al., 2024).

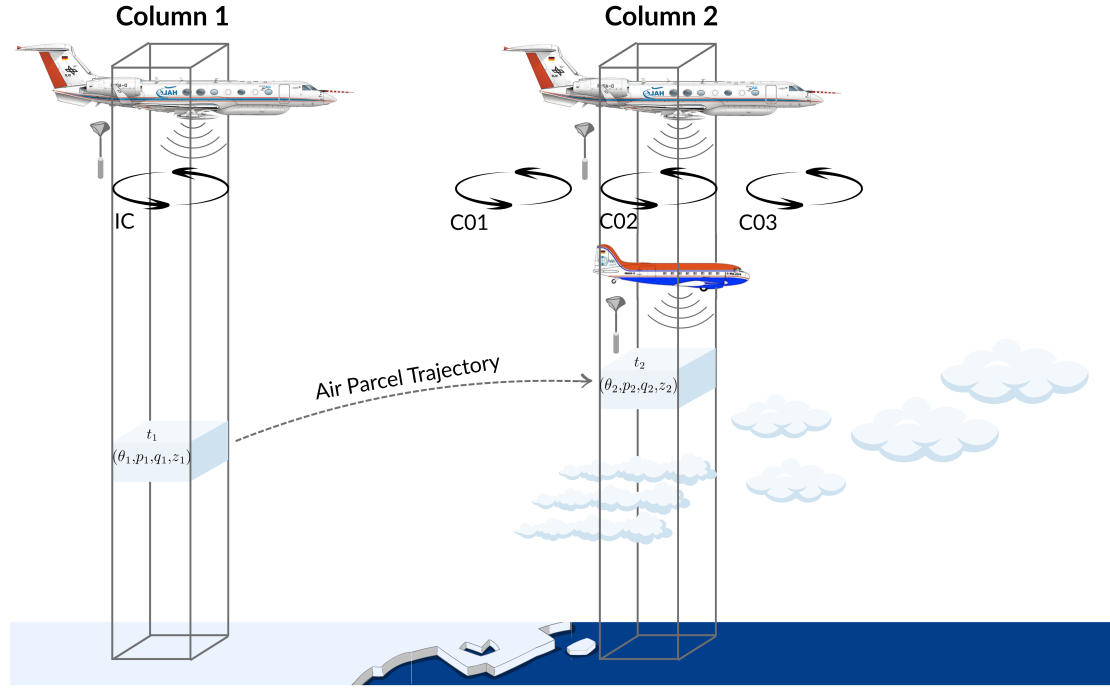


Figure 3.1: Schematic of HALO- $(\mathcal{AC})^3$ quasi-lagrangian flight strategy, with mesoscale dropsonde observations at locations IC of RF10 and C01, C02 and C03 of RF11, adapted from Wendisch et al. (2024).

Within the quasi-Lagrangian framework, air masses were first identified using trajectory calculations derived from atmospheric forecasts. The aircraft then sampled the air mass at an initial location, indicated as Column 1 in Figure 3.1, and attempted to intercept it again later along its predicted trajectory. However, the column air mass is not transported uniformly, as the trajectories may vary with height. Subsequent observations thus can only capture specific layers of the originally sampled air mass (illustrated by the blue boxes in Figure 3.1). The flight planning was supported by trajectory calculations using the Lagrangian Analysis Tool (LAGRANTO) (Sprenger and Wernli, 2015), driven by European Centre for Medium-Range Weather Forecasts (ECMWF) IFS Cy43r1 forecasts. For the subsequent analysis of the air mass trajectories, the ECMWF Reanalysis v5 (ERA5) reanalysis was used (Hersbach et al., 2020).

For the specific case analysed in this thesis, HALO performed circular flight patterns, indicated as circles in Figure 3.1. The initial circle IC was observed during research flight RF10 on 29th March 2022 (flight pattern shown in Figure 3.2), capturing the initial state t_1 of the observed MCAO through the release of dropsondes. On the following day, the same air mass was intercepted again by HALO at time t_2 in Column 2 using the second circle leg Co2 of the flight pattern during RF11.

Two additional circular patterns (Co1 and Co3) were flown along the predicted trajectory. However, these did not sample the same air mass, due to the slow propagation speed of the air mass relative to the aircraft's flight speed. Additional observations in Column 2 at lower altitudes were obtained by the P5 aircraft, which sampled the air mass shortly after it had passed the location (see Figure 3.1).

3.1.2 ADDITIONAL OBSERVATIONS

Parallel to the HALO-(AC)³ campaign, several other field campaigns were conducted in the same region. One of these was the Isotopic Links to Atmospheric Water's Sources (ISLAS2022) field campaign (ISLAS2022, 2023), which utilised controlled meteorological balloon (CMET) balloons. These balloons use a reversible lift-gas compression technique to regulate their altitude, allowing them to travel with an air mass along its trajectory for several days (Voss et al., 2010). In this thesis, observations of temperature and humidity from these platforms are used to verify the simulation setup with an independent dataset. The instruments and data are introduced in more detail in Chapter 5.

3.2 CASE DESCRIPTION

This thesis is based on an MCAO observed on 29th March 2022 (RF10) and 30th March 2022 (RF11) over the Fram Strait. During this event, cold Arctic air originating over the sea-ice-covered central Arctic was advected southward across the MIZ and over the open ocean. The synoptic-scale circulation during this case was characterised by a high-pressure system located near eastern Greenland and a low-pressure system over the Barents Sea. The pressure gradient between the two systems generated a persistent northerly flow in this region, which favoured cold advection from the Arctic Ocean to the Fram Strait, with near-surface wind speeds of approximately 6 m s^{-1} to 10 m s^{-1} . The synoptic situation is shown in Figure 3.2.

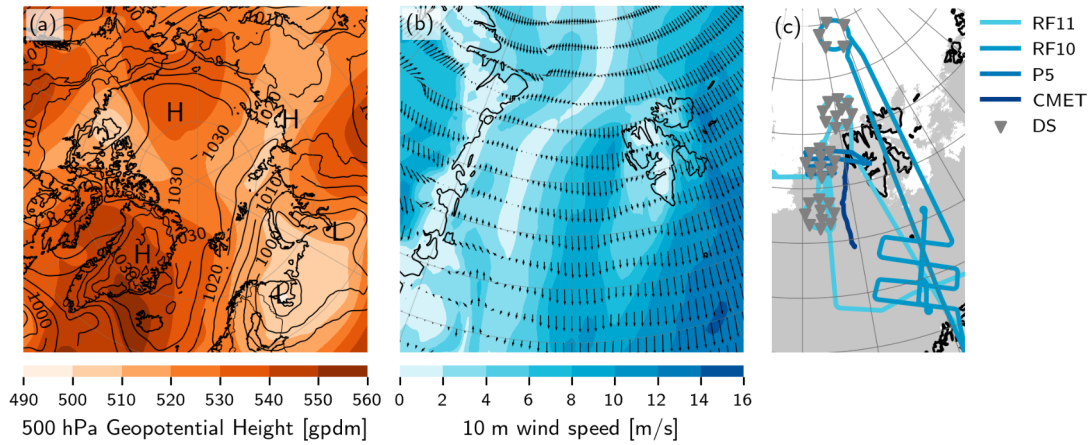


Figure 3.2: Maps of geopotential height at 500 hPa (shading) and mean sea level pressure (black contour lines; a) and wind speed at 10 m (shading) and wind direction (arrows) on 30th March 2022 at 12:00 UTC during HALO- $(\mathcal{AC})^3$ from ERA5 (b). Flight paths of HALO flights RF10 and RF11 and P5 on 30th March 2022, together with CMET, Dropsonde (DS) locations ($\blacktriangledown$) and the sea-ice concentration retrieved from satellite observations (Spreen et al., 2008) indicated in grey and white (c).

As described in Section 3.1.1, HALO followed a quasi-lagrangian sampling strategy observing the evolution of an air mass along its southward trajectory. During RF10 on 29th March 2022 and RF11 on 30th March 2022 circular mesoscale flight patterns with a diameter of about 150 km were flown while releasing several dropsondes along the circle (Figure 3.2 c). These dropsondes provided vertical profiles of thermodynamic and dynamical variables and allowed the derivation of divergence and subsidence from horizontal wind gradients (Section 3.3.4). Additional independent observations were taken with the P5 aircraft and the CMET balloon. These have been used as independent validation in this thesis.

MCAO INDEX

The thermodynamic instability of an MCAO can be quantified using the MCAO Index, which is defined as the difference in potential temperature θ between the sea surface and 800 hPa (Papritz and Spengler, 2017; Dahlke et al., 2022),

$$M = \theta_{SKT} - \theta_{800 \text{ hPa}} . \quad (3.1)$$

This index provides a measure of the lower-tropospheric instability associated with MCAOs. Positive values indicate lower-tropospheric instability, favouring convective overturning and cloud formation, whereas negative values correspond to stable stratification. Figure 3.3 a shows the MCAO index from dropsonde observations and a Lagrangian LES (Ctrl; see Chapter 5). At the initial stage, when the air mass moves over the open ocean, M ranges between two and four, classifying the event as a weak MCAO according to Papritz and Spengler (2017). With increasing distance to the sea-ice edge, M decreases, as the ABL grows and decouples (Figure 5.12).

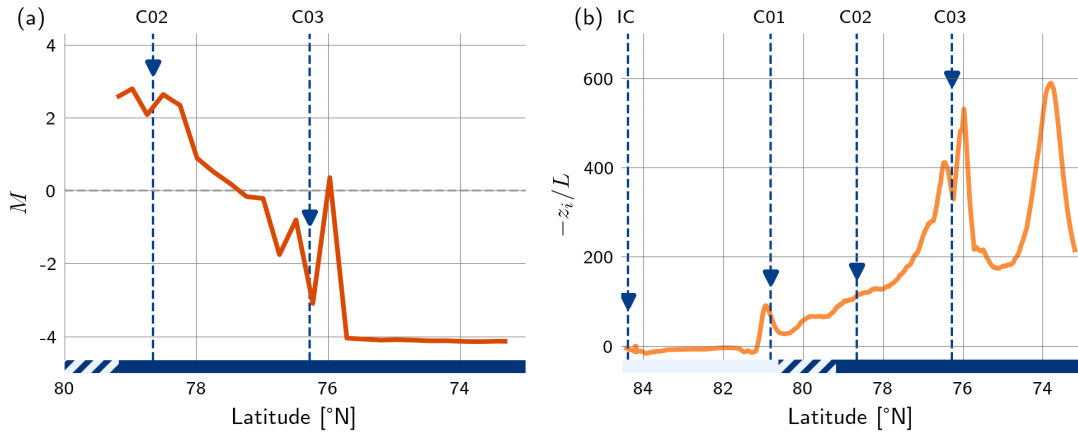


Figure 3.3: Latitude dependence of the MCAO Index M (a) and the stability parameter $\frac{z_i}{L}$ (b) from a 45 h semi-Lagrangian DALES simulation (Ctrl), along the trajectory of an MCAO observed on 30th March 2022, compared with dropsonde observations ($\blacktriangledown$). Dashed lines indicate the locations of dropsonde circles, similarly published in Paulus et al. (2026) and revisited in Chapter 5.

STABILITY PARAMETER

The atmospheric stability during the event is further characterised by the stability parameter $-\frac{z_i}{L}$, where z_i denotes the boundary layer height and L is the Obukhov length, which

characterises the role of buoyancy in turbulent flows. $-\frac{z_i}{\mathcal{L}}$ therefore provides a measure of the degree of convective instability relative to the depth of the boundary layer (Stull, 1988). Following Klingebiel et al. (2025), $\frac{z_i}{\mathcal{L}}$ can be calculated from dropsonde observations based on a method developed by Brümmer (1999):

$$-\frac{z_i}{\mathcal{L}} = z_i \frac{\kappa g C_H \Delta\theta_{a,s} u_{90}}{\theta_0 (C_D u_{90}^2)^{3/2}}, \quad (3.2)$$

with u_{90} the 90 m wind speed, $\Delta\theta_{a,s}$ the potential temperature difference between the air in 90 m height and the water surface and C_D , C_H the dimensionless drag and heat transfer coefficients, which are assumed to be $C_D = C_H = 1.3 \cdot 10^{-3}$ (same as Brümmer, 1999). In Figure 3.3 b, the stability parameter from the LES (Ctrl; see Chapter 5), together with dropsonde observations. As the air mass travels southward over open water, the surface heat fluxes increase and the boundary layer becomes progressively more unstable, similar to observations by previous studies for other MCAOs (e.g. Klingebiel et al., 2025). Thus, the studied case represents a comparatively weak MCAO event with moderate northerly winds and relatively small thermodynamic instability compared to more intense outbreaks (Fletcher et al., 2016a).

3.3 SETUP OF LAGRANGIAN LARGE EDDY SIMULATION

In order to realistically represent the atmospheric state in a Lagrangian LES, it is beneficial to include observed variables as input in the initialisation, boundary conditions and larger-scale forcings of the model. Lagrangian models follow a predefined trajectory of an air parcel, which can be generated from a global model or reanalysis. The following section will discuss examples of how such simulations can be initialised and run almost entirely with observations from an aircraft campaign such as HALO-(AC)³, a more detailed description of the concrete setup used in this thesis will be given in Chapter 5.

3.3.1 TRAJECTORY CALCULATION

The trajectory of an air parcel can be estimated from the evolution of the three-dimensional wind field from a model or reanalysis output by numerically solving the trajectory equation (Wernli and Davies, 1997; Sprenger and Wernli, 2015):

$$\frac{D\vec{x}}{Dt} = \vec{u}(x) , \quad (3.3)$$

where the position of an air parcel in geographical coordinates is $\vec{x}$ and $\vec{u}$ is the 3-D wind vector. Tools like LAGRANTO take a three dimensional wind field from a model or reanalysis output and track air parcels from a defined three-dimensional starting point at a specific location and height level and integrate them forward or backward along the wind direction to find the location and height of the air mass in the next time step, interpolating the wind between the grid points and time steps (Sprenger and Wernli, 2015). This gives the trajectory of the mean flow an air parcel follows, with the resolution of the input model data, not resolving turbulence (Kirbus et al., 2024), and can be used as a simulation trajectory for a Lagrangian LES.

3.3.2 INITIALISATION

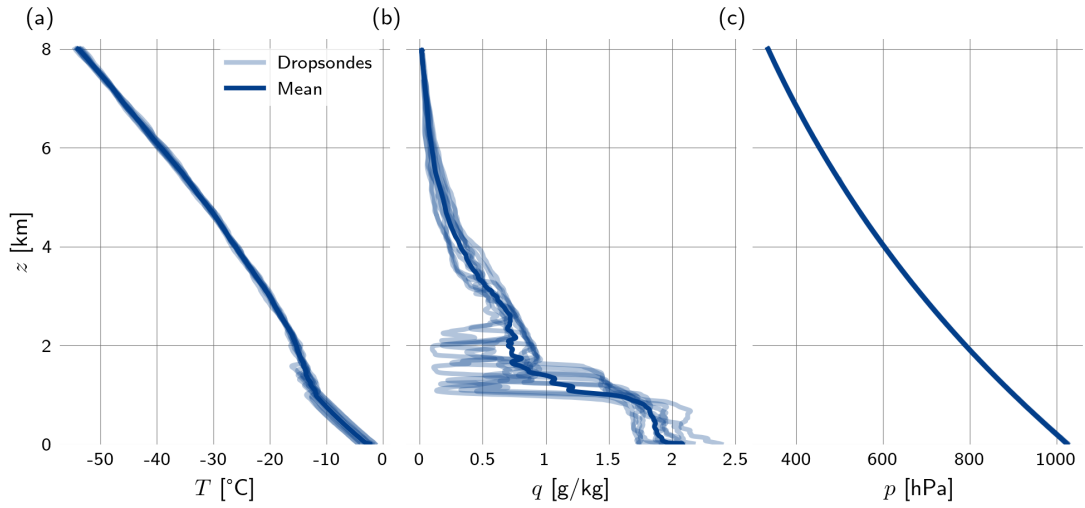


Figure 3.4: Dropsonde profiles recorded during HALO- $(\mathcal{A})^3$ RF10 in the area of IC used to initialise the Lagrangian LES. Shaded profiles show individual dropsonde measurements, while opaque lines indicate the averaged profiles over five dropsondes released in the initialisation area for temperature T (a), specific humidity q (b), and pressure p (c).

At the start of the trajectory, the simulation needs to be initialised by describing the atmospheric state. In this thesis, the model domain is assumed to be horizontally homogeneous at the start of the simulation and is therefore initialised with averaged vertical profiles of pressure p , temperature T , and humidity q observed during the HALO-($\mathcal{AC}$)³ research flight RF10. The initial dropsonde profiles used in this thesis are shown in Figure 3.4. While the pressure and temperature profiles show very little variation across different recordings, the humidity profiles exhibit noticeable variability between individual dropsondes, thus averaging over the five available soundings accounts for local features and ensures consistency with the assumption of horizontal homogeneity in the model domain.

3.3.3 BOUNDARY CONDITIONS

DALES uses periodic horizontal boundary conditions, feeding the outgoing flow back into the domain on the opposite side. The surface boundary conditions must be prescribed along the trajectory of the model, in particular the skin temperature T_{skin} and the sea-ice fraction. These quantities can be retrieved from the VELOX airborne thermal infrared imager on board HALO. VELOX measures thermal infrared radiation emitted by clouds, sea-ice, and ocean in several wavelength channels between 7.7 and 12 μm (Schäfer et al., 2022). From these measurements, cloud-top temperatures as well as surface temperatures over cloud-free regions can be derived. In addition, the thermal contrast between ocean and sea-ice allows the retrieval of sea-ice cover at high spatial resolution. The instrument provides two-dimensional thermal images with a spatial resolution of approximately 10 m $\times$ 10 m at a target distance of 10 km (Müller et al., 2025). For this study, images obtained during the circular flight legs of the RF10 and RF11 HALO flights are longitudinally averaged and interpolated along the simulation trajectory. A more detailed explanation and visualisation of this procedure is given in Section 5.3.3. Additional surface parameters required by the model, such as the surface albedo, the aerodynamic roughness length for momentum, and the thermal roughness length for heat, could not be observed directly during the campaign. These quantities were therefore obtained from the ERA5 reanalysis (Hersbach et al., 2020).

3.3.4 LARGE-SCALE FORCINGS FROM OBSERVATIONS

Large-scale forcings are a critical component of LES because they provide the environmental constraints that drive the evolution of ABL structures. Observing mesoscale gradients

required for the derivation of these forcings is difficult; therefore, observational datasets sufficient to run LES with observationally derived forcings in the Arctic are rare. However, they can provide valuable insights into the development of air masses during their transformation. As discussed in Chapter 1, mesoscale gradients in the atmosphere, particularly horizontal divergence and subsidence, play a fundamental role in the development of Arctic MCAOs (e.g. Schirmacher et al., 2024). In LES models, the specification of large-scale vertical motion strongly controls the thermodynamic structure of the ABL and was found to significantly influence simulated cloud cover and boundary layer evolution (e.g., van der Dussen et al., 2016).

To derive large-scale forcings such as advection and subsidence, horizontal gradients of humidity, temperature, and wind must be observed. Early approaches relied on ground-based radiosonde networks, where simultaneous soundings at multiple stations were used to estimate horizontal gradients between vertical profiles at different locations (e.g., Mapes et al., 2003). To improve the temporal coverage of wind observations, Mapes and Lin (2005) introduced Doppler radar techniques to derive temporally continuous wind and divergence fields. However, these methods remain geographically limited to accessible regions. To overcome these spatial limitations and enable observations in remote regions, Lenschow et al. (1999) estimated horizontal wind divergence D from aircraft measurements of the horizontal wind $\vec{v}_h$ along a closed circular flight track of area A using

$$D = \frac{1}{A} \oint_{\partial A} \vec{v}_h \cdot \vec{n}, dl . \quad (3.4)$$

Alternatively, the divergence can be defined from the horizontal wind vector as

$$D = \nabla \cdot \vec{v}_h . \quad (3.5)$$

To solve this, Lenschow et al. (2007) established a regression-based approach, calculating horizontal gradients from airborne measurements over a mesoscale area. This method was further developed by Bony and Stevens (2019), who used dropsonde releases along aircraft flight patterns together with a linear regression method to estimate vertical profiles of gradients across the mesoscale field between the observed profiles. This approach has been applied during measurement campaigns in the subtropics and midlatitudes (George et al., 2021; Li et al., 2021; Dorff et al., 2024).

These studies estimate the horizontal wind $\vec{v}_h$ using a first-order Taylor expansion around the centre point of the flight pattern:

$$\vec{v}_h = \vec{v}_{h0} + \frac{\partial \vec{v}_h}{\partial x} \Delta x + \frac{\partial \vec{v}_h}{\partial y} \Delta y + \frac{\partial \vec{v}_h}{\partial t} \Delta t + \mathcal{O}(\vec{v}_h'') , \quad (3.6)$$

where (x, y, z) are Cartesian coordinates, Δx is the eastward distance, Δy the northward distance from the centre of the flight circle, and $\vec{v}_{h0}$ is the mean wind averaged over all dropsondes. [Bony and Stevens \(2019\)](#) assumed a stationary flow, neglecting the time dependence and ignoring higher-order terms, reducing the expression

$$\vec{v}_h = \vec{v}_{h0} + \frac{\partial \vec{v}_h}{\partial x} \Delta x + \frac{\partial \vec{v}_h}{\partial y} \Delta y , \quad (3.7)$$

which represents an overdetermined linear system of equations. A least-squares solution can therefore be obtained for the horizontal gradients $\frac{\partial \vec{v}_h}{\partial x}$ and $\frac{\partial \vec{v}_h}{\partial y}$ through linear regression.

In previous studies conducted in the subtropics, the geographic coordinate system (latitude and longitude) was approximated as locally equidistant, allowing Cartesian coordinates (x, y, z) to represent horizontal gradients with sufficient accuracy. However, at high latitudes, the horizontal distances between longitudes become increasingly distorted, which can lead to significant errors in gradient calculations. Therefore, the regression method can instead be transformed to spherical coordinates with radial distance r from Earth's centre, polar angle ϑ , and azimuthal angle ϕ . This formulation naturally accounts for Earth's curvature and allows more accurate gradient estimation in polar regions.

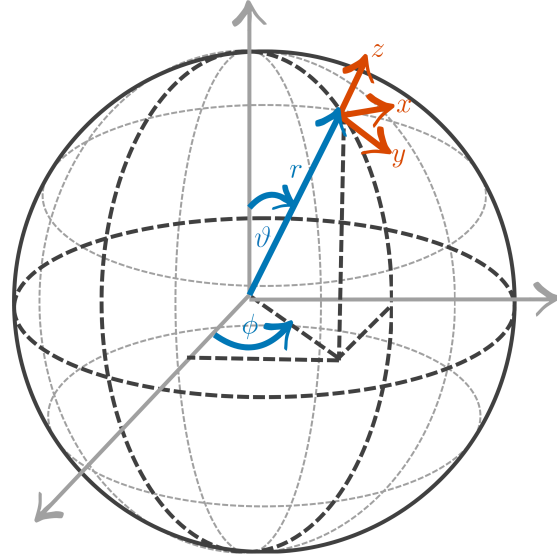


Figure 3.5: Schematic of Cartesian and spherical coordinate systems used to describe locations in the atmosphere.

The relationship between Cartesian and spherical coordinates (illustrated in Figure 3.5) is

$$x = r \cos(\phi) \vartheta \quad (3.8)$$

$$y = r \phi \quad (3.9)$$

$$z = r - R_{\text{Earth}} , \quad (3.10)$$

and the corresponding differentials are

$$dx = r \cos(\phi) d\vartheta \quad (3.11)$$

$$dy = r d\phi \quad (3.12)$$

$$dz = dr . \quad (3.13)$$

Following Equation (3.7) and assuming stationarity, any variable $\xi \in [u, v, q, T, p]$ can be approximated as

$$\xi = \xi_0 + \frac{\partial \xi}{\partial \vartheta} \Delta \vartheta + \frac{\partial \xi}{\partial \phi} \Delta \phi . \quad (3.14)$$

The horizontal advection of ξ can be calculated from the horizontal gradients as

$$- \vec{v}_h \cdot \nabla \xi = - \left[u \frac{\partial \xi}{\partial x} + v \frac{\partial \xi}{\partial y} \right] = - \left[u \frac{1}{r} \frac{\partial \xi}{\partial \vartheta} + v \frac{1}{r \sin \vartheta} \frac{\partial \xi}{\partial \phi} \right] . \quad (3.15)$$

The horizontal divergence is given by

$$D = \nabla \cdot \vec{v}_h = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = \frac{1}{r \sin \vartheta} \left[\frac{\partial u}{\partial \phi} + \frac{\partial \sin \vartheta v}{\partial \vartheta} \right] . \quad (3.16)$$

Using the continuity equation (Equation (2.6)) and neglecting density fluctuations, the horizontal divergence can be related to the vertical velocity w :

$$D = - \frac{\partial w}{\partial z} = - \frac{\partial w}{\partial r} . \quad (3.17)$$

Since the radial coordinate r and the Cartesian height z are linearly related, vertical gradients can be expressed either in z or in r without further conversion. Integrating from the surface yields the vertical velocity

$$w = w(0) - \int_0^z D \, dz = - \int_0^z D \, dz , \quad (3.18)$$

with the vertical velocity being zero at the surface by definition ($w(0) = 0$). Depending on the LES model input and vertical coordinates, the vertical velocity can be expressed in pressure coordinates as

$$\omega = \frac{dp}{dt} . \quad (3.19)$$

Using the chain rule, this can be related to the vertical velocity in height coordinates as

$$\omega = \frac{\partial p}{\partial z} \frac{dz}{dt} = \frac{\partial p}{\partial z} w . \quad (3.20)$$

Under the hydrostatic approximation, valid for large-scale flows with negligible vertical acceleration, the vertical pressure gradient balances gravity:

$$\frac{\partial p}{\partial z} = -\rho g . \quad (3.21)$$

The relationship between ω and w becomes

$$\omega = -\rho g w . \quad (3.22)$$

The horizontal pressure gradients are prescribed in LES through the geostrophic wind. The geostrophic wind represents the large-scale pressure gradient force balanced by the Coriolis force and can be written as

$$\vec{v}_g = -\frac{1}{\rho 2\Omega \sin \phi} \nabla p = -\frac{1}{\rho 2\Omega \sin \phi} \begin{pmatrix} \frac{\partial p}{\partial y} \\ \frac{\partial p}{\partial x} \end{pmatrix} = -\frac{1}{\rho 2\Omega \sin \phi} \begin{pmatrix} \frac{1}{r} \frac{\partial p}{\partial \phi} \\ \frac{1}{r \cos \phi} \frac{\partial p}{\partial \theta} \end{pmatrix} \quad (3.23)$$

Together, the derived horizontal advection, geostrophic wind, and subsidence represent the large-scale flow, which influences the evolution of the simulated boundary layer and cloud field and provides a complete set of large-scale forcings required to drive Lagrangian LES.

To evaluate the robustness of the methodology under Arctic conditions, independent estimates of horizontal divergence were calculated using the two approaches introduced above: the line-integral method (Equation (3.4)) and the regression method (Equation (3.5)). This thesis uses the measurements from dropsondes, which were released from two different research aircraft (HALO and P5), to provide independent observational datasets. The comparison between both datasets allows an assessment of the consistency and uncertainty in the derived large-scale forcings, with the results presented in detail in Chapter 4.

4

Airborne Measurements of Mesoscale Divergence at High Latitudes during HALO- $(\mathcal{AC})^3$

The following chapter addresses the first research question:

HOW CAN ATMOSPHERIC SUBSIDENCE BE QUANTIFIED IN THE
ARCTIC USING OBSERVATIONS?

Using airborne dropsonde measurements from HALO- $(\mathcal{AC})^3$, subsidence is derived from horizontal divergence based on mass continuity, and the robustness of this approach under Arctic conditions is assessed.

This work has been published in the Journal of the Atmospheric Sciences of the American Meteorological Society (AMS), and its format has been changed to match this thesis.

Paulus, F. M., M. Karalis, G. George, G. Svensson, M. Wendisch, and R. A. J. Neggers, 2024: Airborne Measurements of Mesoscale Divergence at High Latitudes during HALO- $(AC)^3$. J. Atmos. Sci., 81, 2051–2067, <https://doi.org/10.1175/JAS-D-24-0034.1>.

© American Meteorological Society. Used with permission.

AUTHOR CONTRIBUTION

FP processed the data and implemented and evaluated the regression method together with RN. GG provided the basis for the data processing of dropsonde measurements and the regression method. MK and GS performed the line-integral method. RN, GG, and GS advised on the interpretation of the results. FP prepared the manuscript, MK wrote several sections, and RN, GS, GG, and MW revised various intermediate versions.

DATA AVAILABILITY

The pre-processed dropsondes data are available at [George et al. \(2024\)](#) and can be accessed via the ac3airborne module ([Mech et al., 2022](#)). The ERA5 data sets can be freely retrieved from the Copernicus Climate Change Service (C3S) Climate Data Store (CDS) at <https://cds.climate.copernicus.eu/>, last accessed on 6 November 2023. The code used to process the data is published at [Paulus and Karalis \(2023\)](#) (<https://zenodo.org/doi/10.5281/zenodo.10402060>).

AIRBORNE MEASUREMENTS OF MESO-SCALE DIVERGENCE AT HIGH LATITUDES DURING HALO- $(\mathcal{AC})^3$

ABSTRACT

Boundary layer cloud transformations at high latitudes play a key role for the Arctic climate and are partially controlled by large-scale dynamics such as subsidence. While measuring large- and mesoscale divergence on spatial scales on the order of 100 km has proven notoriously difficult, recent airborne campaigns in the subtropics have successfully applied measurement techniques using multiple dropsonde releases in circular flight patterns. In this paper it is shown that this method can also be effectively applied at high latitudes, in spite of the considerable differences in atmospheric dynamics compared to the subtropics. To show the applicability, data collected during the airborne HALO- $(\mathcal{AC})^3$ field campaign near Svalbard in Spring 2022 were analysed, where several flight patterns involving multiple dropsonde launches were realised by two aircraft. This study presents a first overview of the results. We find that the method indeed yields reliable estimates of mesoscale gradients in the Arctic, producing robust vertical profiles of horizontal divergence and consequently, subsidence. Sensitivity to aspects of the method is investigated, including dependence on sampling area and the divergence calculation.

SIGNIFICANCE STATEMENT

The aim of this work is to report encouraging results with a recently proposed aircraft-based method for measuring mesoscale vertical motions at high latitudes. Dropsonde data from the recent HALO- $(\mathcal{AC})^3$ campaign is used to this purpose. The method has so far mainly been applied at subtropical to mid-latitudes, but not in the Arctic, where the weather and climate conditions are very different. Gaining insight is significant because vertical winds play a key role in the Arctic climate system, including air mass transformations and the behaviour of

clouds and precipitation. The results motivate use of the measured vertical winds in follow-up studies with process models.

4.1 INTRODUCTION

The accurate representation of vertical motion, including divergence and subsidence, is essential for model studies, particularly to represent the boundary layer and study clouds. Divergence and related vertical motions (including subsidence) change the vertical structure of the atmosphere, influencing thermal stability and relative humidity, which in turn affect radiation and cloud dynamics. [Neggers et al. \(2019\)](#) established a direct link between subsidence and the life cycle of low-level clouds in the Arctic, where strong and sudden subsidence events were responsible for cloud collapses. This discovery, based on Lagrangian large eddy simulation (LES) experiments of transitioning warm air mass intrusions into the high Arctic, holds significant implications for the local energy budgets in the Arctic.

Therefore, the study of subsidence is of premiere importance. However, measuring subsidence presents a classic challenge in meteorology. The small signal and considerable fluctuation of vertical motion pose obstacles to direct observation. This challenge has led to the development of various methods to estimate and quantify subsidence. [Lenschow et al. \(1999\)](#) pioneered methods to quantify horizontal divergence through aircraft-based measurements, while [Mapes and Lin \(2005\)](#) developed a method to derive divergence, and thereby subsidence, using ground-based Doppler-radar observations. These methods rely on mass conservation for deriving the divergence. While surface-based methods have been used widely to measure subsidence ([Mapes et al., 2003](#); [Mapes and Lin, 2005](#); [Li et al., 2018](#)), they can be challenging to apply in the Arctic, particularly in the central Arctic region. The remote and harsh conditions of the Arctic, along with limited accessibility, imply serious difficulties for deploying and maintaining land-based measurement systems. As a result, aircraft-based methods offer a viable alternative for obtaining subsidence information in these challenging environments.

Recent advancements in atmospheric research have introduced a new technique for measuring vertical motion, through dropsondes released along horizontal, circular flight patterns by aircraft ([Bony and Stevens, 2019](#)), inspired by techniques based on [Lenschow et al. \(2007\)](#). Notably, campaigns like the Next Generation Remote Sensing for Validation Studies

(NARVAL2) (Bony and Stevens, 2019) and Elucidating the role of clouds-circulation coupling in climate (EUREC⁴A) (Stevens et al., 2021) have leveraged the High Altitude and Long Range Aircraft (HALO) to execute these horizontal circles in the subtropics. While the method has already shown promise in subtropical regions (Bony and Stevens, 2019; George et al., 2021) and cold air outbreaks in the western North Atlantic Ocean (Li et al., 2021), its applicability to the Arctic cannot be assumed a priori. The theoretical suitability of this method for Arctic studies opens novel pathways to probe the complexities of vertical motion and its implications. The distinct atmospheric conditions in the Arctic, however, pose unique challenges that differ from those encountered in subtropical areas. The Rossby radius and time scale of deformation decreases with latitude, causing the dynamical features in the Arctic to be more transient and spatially constrained. This calls into question the soundness of the stationarity assumption necessary for this observational framework, as well as the representativity of the column measurements for the general divergence fields, which was previously established for the subtropics (Bony and Stevens, 2019). Therefore, exploring the viability of this method in the Arctic context is of paramount importance, as it holds the potential to unravel crucial insights into subsidence dynamics and their implications in high-latitude regions.

The High Altitude and Long Range Aircraft - Trans-regional Collaborative Research Center TRR 172 - Arctic Amplification: Climate Relevant Atmospheric and Surface Processes and Feedback Mechanisms (HALO-(AC)³) campaign in March and April 2022, was partly designed to evaluate the method within the Arctic context. The central goals of the campaign include the investigation of key aspects of atmospheric circulation such as air mass transformations, and their implications for climate dynamics in the Arctic through quasi-Lagrangian observations (Wendisch et al., 2021). One of the primary objectives was to study the transformations of air masses during warm air intrusions and cold air outbreaks in the Arctic region. A key focus of the campaign involved flying mesoscale ($\mathcal{O}(100 \text{ km}, 1 \text{ h})$) horizontal circles with HALO, during which dropsondes are released at regular intervals along the circle. These circles can capture the area-averaged atmospheric divergence and subsidence, allowing further investigations on air mass dynamics.

This study undertakes a dual-pronged investigation with the primary goals of determining the effectiveness of the horizontal circle method and comprehensively substantiating the feasibility of employing dropsondes for measuring the vertical profile of mass divergence and

vertical motion in the Arctic. By integrating these objectives, the study aims to document the circle method's applicability in the Arctic environment.

Section 4.2 delves into the methodology, outlining how the vertical profiles of divergence and vertical velocity can be derived from horizontal wind measurements. Additionally, the airborne field campaign is introduced, along with the flight strategy adopted to test the technique. Section 4.3 offers the resulting vertical profiles and discusses the credibility of them along with an error analysis. Lastly, Section 4.4 provides a forward-looking perspective on the implications of these measurements for future airborne campaigns.

4.2 DATA AND METHODS

4.2.1 HALO- $(\mathcal{AC})^3$ CAMPAIGN

The HALO- $(\mathcal{AC})^3$ campaign in March and April 2022 in the Norwegian sector of the Arctic was a joint effort by two collaborative programs: (i) the Trans-regional Collaborative Research Center TRR 172 - Arctic Amplification: Climate Relevant Atmospheric and Surface Processes and Feedback Mechanisms ($(\mathcal{AC})^3$) (Wendisch et al., 2023), and (ii) the Priority Program 1294 for the Atmospheric and Earth System Research program with the High Altitude and Long Range Aircraft (HALO). Both programs are funded by the German Research Foundation. Three airborne platforms were used to probe and understand transforming air masses in the region. A Lagrangian sampling approach was adopted to track and sample air masses as efficiently and frequently as possible during their transformation (Wendisch et al., 2024). The campaign leveraged the unique capabilities of multiple distinct aircraft. This study relies on the dropsonde measurements launched by:

- **HALO:** A long-range research aircraft operated by the German Aerospace Center (DLR), operating from Kiruna, Sweden as a remote sensing platform (Stevens et al., 2019). HALO conducted multiple circular flight patterns (150 km in diameter) at different locations releasing 5 to 10 dropsondes with even spacing along the flight path (Figure 4.1). All circular flight pattern took between 35 min to 45 min. The platform has an Airborne Vertical Atmospheric Profiling System (AVAPS) dropsonde system (NSF NCAR Earth Observing Laboratory, 1993; Hock and Franklin, 1999) and uses Vaisala RD-41 sondes (Black et al., 2017).

- **Polar 5 Aircraft (P5):** A low-flying aircraft operated by the Alfred-Wegener-Institut (AWI), based in Longyearbyen, Svalbard (Wesche et al., 2016), also operated as a remote sensing platform. This aircraft conducted active and passive remote sensing measurements, focusing on investigating smaller-scale processes in the lower troposphere (at altitudes below 3 to 5 km). The platform operates an AVAPS Lite dropsonde system (Michaelis et al., 2022) also using Vaisala RD-41 dropsondes (Vaisala, 2023).

Vaisala RD-41 dropsondes measured temperature with a resolution of 0.01 °C, humidity with a resolution of 0.1 % RH and pressure with a resolution of 0.01 hPa (Vaisala, 2023). The horizontal wind velocity was derived from the Global Positioning System (GPS) with an estimated accuracy of 0.1 m s⁻¹ (Hock and Franklin, 1999; Wang et al., 2015).

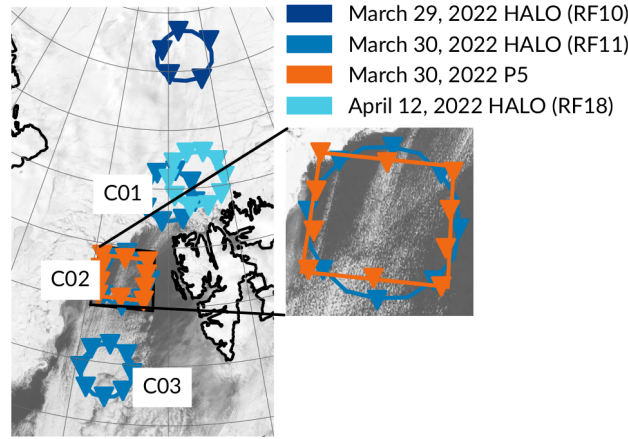


Figure 4.1: HALO-(AC)³ flight track patterns for three HALO flights (RF10, RF11 and RF18, blue) and Polar 5 (P5, red) with the locations of the dropsonde launches (▼). Satellite image at the time of RF11 30th March 2022 from Terra MODIS (NASA, 2022) in the background, adapted in modified form from Fig.17a in Wendisch et al. (2024).

This study is based on three HALO-(AC)³ Research Flights (RFs) of HALO, conducted on 29th March 2022 (RF10), 30th March 2022 (RF11) and 12th April 2022 (RF18). On all three days, mesoscale circles (150 km diameter) were flown west of Svalbard. Figure 4.1 shows their locations and a satellite image of the day of RF11 for reference. RF10 included a mesoscale horizontal circle close to the North Pole but was subject to strict air space restrictions, by which only five dropsondes could be released. On the next day during RF11, three circles were flown (C01 north of Svalbard, C02 west of Svalbard and C03), each situated

to match the most estimated southbound Lagrangian trajectories of the air masses sampled during RF10. A total of eight dropsondes were deployed along the perimeter of each circle, and two additional sondes were released in their interior. The third flight with a circular flight pattern investigated here (RF18), featured a single circle with 10 sondes launched on its perimeter. Each HALO- $(\mathcal{AC})^3$ flight pattern was conducted at a height of ~ 8 km and took up to 45 min time.

Notably on 30th March 2022, the Polar 5 executed a gridded flight pattern with regularly spaced dropsonde releases, co-located with HALO RF11's second mesoscale circle (RF11 Co2) (Figure 4.1, green) at a height of ~ 2.5 km. This strategic spatial co-location of flight patterns had the aim of obtaining two independent sets of soundings for the same area, enabling a critical assessment of the divergence calculations. For operational reasons, P5 conducted its pattern 45 min after the co-located RF11, and the pattern took 3 h in total, because of a more complex flight pattern, but all dropsondes on the outer edge of the pattern were dropped within 2 h. To avoid confusion, with different numbering of the Research flights of both aircraft, this flight will not be referred to by its flight number, but by the platform name (P5).

4.2.2 LINE-INTEGRAL METHOD

Given the dropsonde data, the first method for calculating divergence relies on a line integral. Vertical wind velocity in the atmosphere is related to horizontal wind divergence using the continuity equation under the incompressibility assumption. In practice, estimating the mean divergence D , and thus subsidence, of an area A , is achieved by measuring the integrated horizontal velocity $\vec{v}_h$ at its boundary ∂A , where ∂A is a closed curve surrounding A , according to the Gauss theorem:

$$D = \frac{1}{A} \oint_{\partial A} \vec{v}_h \cdot \vec{n} \, dl . \quad (4.1)$$

This is the conceptual basis behind the Line-Integral Method, which utilizes the perimeter's closed nature of measuring horizontal wind either at flight level (Lenschow et al., 1999, 2007) or with sounding profiles (Bony and Stevens, 2019).

The number of observations attainable along the flight pattern is finite, therefore, it is more appropriate to write Equation (4.1) in the form of a sum:

$$D = \frac{1}{A} \sum_N v \delta l , \quad (4.2)$$

where, N is the number of dropsondes considered, A is the area enclosed by the flight track and v is the radial component to the curve. δl represents the length element computed as L/N where L is the length of the perimeter, assuming the sounding points are approximately uniformly distributed along the curve. The horizontal displacement of the dropsondes from the mean coordinates of their corresponding profiles during their descent is small and the sondes drift in the same direction, causing only small deformations of the pattern (see Figure 4.6). We estimate that taking the deformation of the pattern in the vertical into account would lead to relative changes smaller than 1 % and 10 % in the divergence and pressure velocity signal respectively (Figure 4.7). For simplicity of the method, we decided to disregard the horizontal displacement of the dropsondes from the mean coordinate of their corresponding profiles for this analysis. Vertical motion, expressed from now on in terms of pressure velocity $\omega = \frac{dp}{dt}$, with the pressure p , can then be derived through the vertical integration of the D profiles over the column and accounting for the change in air density ρ

$$\omega = -g\rho \int_0^z D \, dz , \quad (4.3)$$

where $g = 9.81 \, \text{m s}^{-2}$ is the gravitational acceleration.

4.2.3 REGRESSION METHOD

An alternative method for calculating mesoscale divergence D makes use of its dependence on the spatial gradients in the horizontal wind field $\vec{v}_h = (u, v)$,

$$D = \nabla \cdot \vec{v}_h = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} . \quad (4.4)$$

Under the assumption that wind fluctuations in both longitude and latitude follow a linear pattern at every vertical level z , the wind at any point at the circumference can be

approximated by a first-order Taylor expansion from the centre point of the flight pattern (Lenschow et al., 2007):

$$\vec{v}_h = v_{h,0} + \frac{\partial \vec{v}_h}{\partial x} \Delta x + \frac{\partial \vec{v}_h}{\partial y} \Delta y + \frac{\partial \vec{v}_h}{\partial t} \Delta t + \mathcal{O}(v_h'') , \quad (4.5)$$

with the mean wind velocity over all dropsondes being $v_{h,0}$, the eastward displacement Δx and the northward displacement Δy from the centre of the circle. We can justify omitting the time dependence (stationarity) in Equation (4.5) by the significant difference in aircraft speed compared to wind speed, making the sampling time scale much smaller than the advective time scale. Employing stationarity also resolves concerns arising from the non-independence of the aircraft's time and space sampling (Bony and Stevens, 2019).

Inserting measured data from deployed dropsondes into Equation (4.5), after removing time-derivatives and higher-order terms, yields an over-determined linear system of equations, for which the least squares solution can be found through linear regression. Unlike the Line-Integral Method, this method does not require a closed flight path, it can utilise all dropsondes deployed in a constrained area during a time frame, where stationarity is justifiable (Lenschow et al., 2007).

Most previous efforts using the Regression Method to calculate mesoscale divergence from sonde patterns were conducted in subtropical regions (NARVAL₂, EUREC⁴A) and twice in the mid-latitudes during the Aerosol Cloud Meteorology Interactions over the Western Atlantic Experiment (ACTIVATE) (Li et al., 2021). At those locations, the geographical coordinate system (latitude, longitude) can be approximated equidistant and Cartesian coordinates (x,y,z) are appropriate to express horizontal gradients. In contrast, the HALO-(AC)³ campaign target area is much closer to the North Pole, where horizontal distances associated with Cartesian coordinates from latitudes and longitudes become increasingly distorted, which would introduce substantial errors in gradient calculations.

To effectively address this problem, we introduce the Regression Method in a spherical coordinate system, with the distance r from the Earth's centre, the polar angle ϑ and the azimuth angle ϕ . This enables more accurate calculations by inherently accommodating for the curvature of the Earth.

By applying a coordinate transformation from Cartesian to spherical coordinates (Bronshstein et al., 2007), the Regression Method can then be written as:

$$\vec{v}_h = v_{h,0} + \frac{\partial \vec{v}_h}{\partial \vartheta} \Delta \vartheta + \frac{\partial \vec{v}_h}{\partial \phi} \Delta \phi \quad (4.6)$$

$$D = \nabla \cdot \vec{v}_h = \frac{1}{r \sin(\vartheta)} \cdot \left(\frac{\partial u}{\partial \phi} + \frac{\partial \sin(\vartheta) v}{\partial \vartheta} \right) . \quad (4.7)$$

4.3 RESULTS

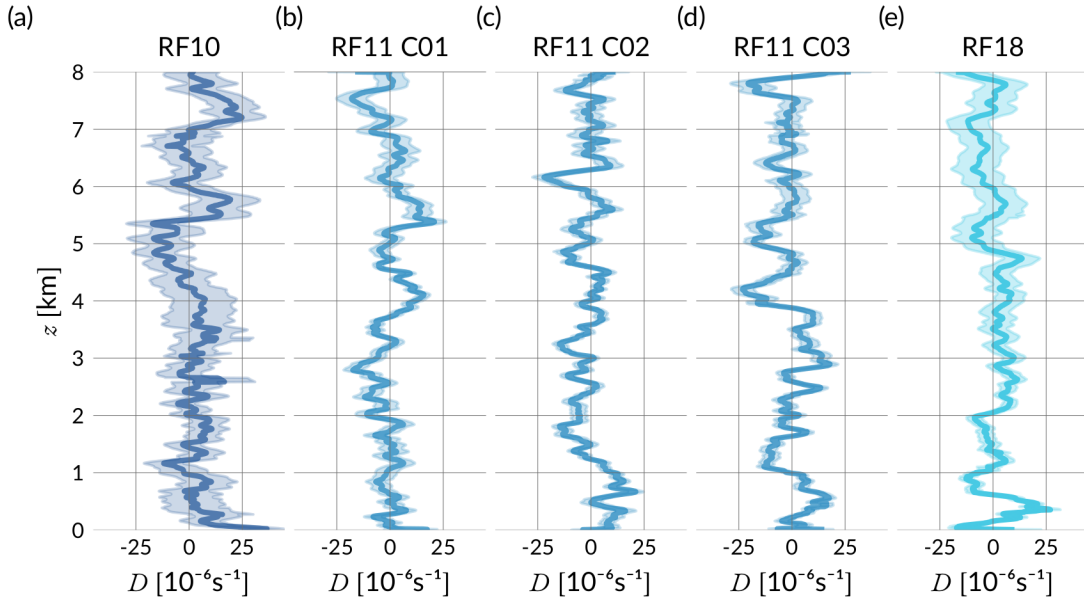


Figure 4.2: Vertical profiles of wind divergence D obtained by the Regression Method from HALO pattern flights RF10 (5 sondes) (a), RF11 C01 (10 sondes), (b) RF11 C02 (10 sondes) (c), RF11 C03 (10 sondes) (d) and RF18 (10 sondes) (e) with estimated error (shaded).

For all available circular flight patterns of HALO (Figure 4.1), we calculated divergence as a function of altitude z up to a height of 8 km using the Regression Method, see Figure 4.2. Of all five circles, the three central ones were flown on the same day, sampling the low-level air mass at multiple locations. The profile of RF10 (Figure 4.2 a) showcases the largest ranges of the estimated error, which we obtained through Gaussian Error Propagation of the standard errors of the wind direction and distance from the centre of the flight pattern, obtained through the regression analysis (Section 4.5.3). This can be attributed to the reduced num-

ber of dropsonde launches for this specific circle (only 5 due to airspace restrictions). Consequently, the reduced sample size contributes to the higher error range, highlighting the potential impact of limited data on the precision of observed fluctuations.

The divergence profiles fluctuate with height around zero. Compared to the divergence profiles, sampled in the subtropics during NARVAL2 and the mid-latitudes during ACTIVATE, we find a narrower range of variation; $(-25 \cdot 10^{-6} \text{ s}^{-1} \text{ to } 25 \cdot 10^{-6} \text{ s}^{-1})$ compared to up to $(-40 \cdot 10^{-6} \text{ s}^{-1} \text{ to } 40 \cdot 10^{-6} \text{ s}^{-1})$ in the previous campaigns (Bony and Stevens, 2019; Li et al., 2021). A second characteristic is the lack of coherent features in horizontal divergence profiles in the lower layers compared to the subtropics (George et al., 2023). The divergence and convergence layers are much thinner than in the subtropics, where Bony and Stevens (2019) find layers of about 2 km, while the observed profiles in the Arctic show smaller layers below 1 km in depth. Combined with the generally lower values of D , this could be due to the more stable atmosphere compared to the subtropics. More pronounced features in the lowest levels visible in RFI1 Co2, RFI1 Co3 and especially RFI8 coincide with deeper boundary layers of 0.43 km, 1.04 km and 0.39 km respectively, while the boundary layer height of RFI0 and RFI1 Co1, which were both over sea-ice, is very close to the ground with 0.065 km.

To test the robustness of the horizontal circle method, and to compare the two alternatives for calculating horizontal divergence as described in Section 4.2.2, we subsequently focus on RFI1 Co2, before coming back to all HALO-flights to study the dependence of the significance of the measurements on the number of dropsondes. The results are shown in Figure 4.3, covering a shallower vertical range up to 2.5 km, slightly below the flight height of the P5 aircraft. Zooming in on this range highlights the existence of a well-defined layer with a marked positive divergence, peaking at about 600 m height, and allows for a more detailed comparison between the profiles from the two aircraft. Also shown are the associated vertical profiles of the pressure velocity ω , peaking at about 1300 m height. For reference, mean profiles from the ERA5 Reanalysis from the European Centre for Medium-Range Weather Forecasts (ECMWF) (Hersbach et al., 2020), averaged over all grid points inside the sampling area (Figure 4.8 a and b) are also shown, at 11:00 UTC (closest time to RFI1 Co2) and 14:00 UTC (time central within P5 flight), together with the boundary layer height calculated from the Bulk-Richardson Number for both flights (Section 4.5.5). It needs to be noted, that the dropsondes used in this study were assimilated in the ERA5 reanalysis and

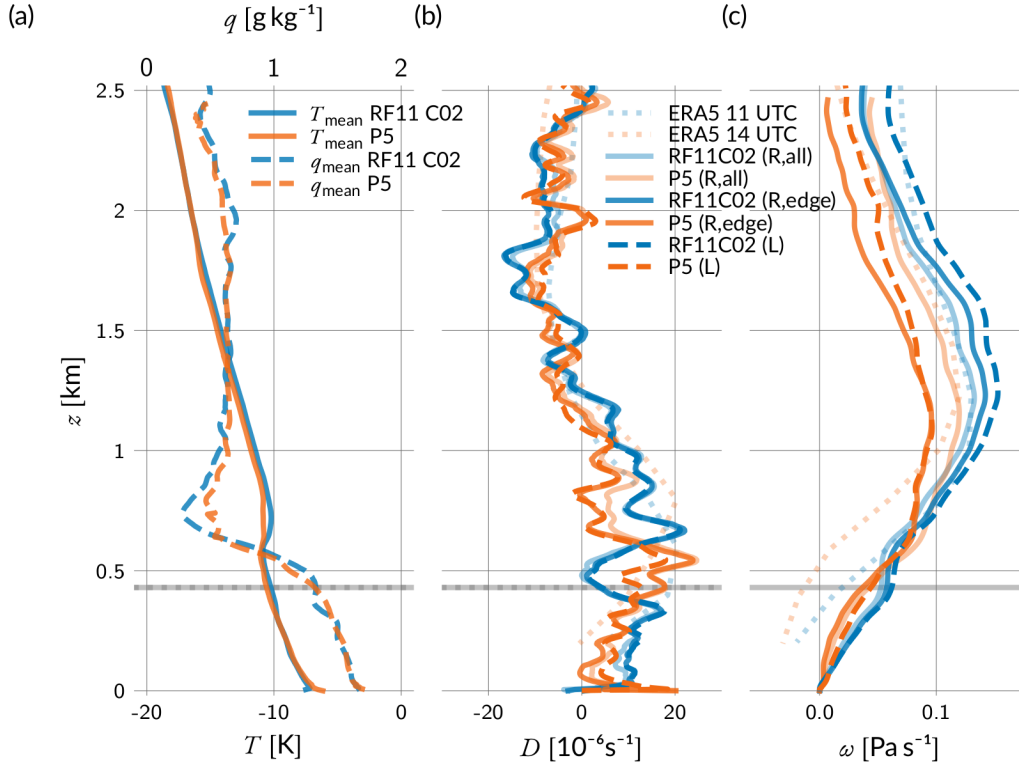


Figure 4.3: Vertical profiles of the mean temperature T_{mean} and specific humidity q_{mean} , averaged over all dropsonde measurements deployed during each flight (a), mass divergence D (b) and pressure velocity ω (c) obtained from dropsonde measurements from co-located flight of HALO(RF11 C02) (blue) and P5 (red) aircraft on 30th March 2022. Regression Method using all sondes (R,all) (semitransparent), Line-Integral Method (dashed), and Regression Method using only sondes which can be included in the Line-Integral Method (R,edge) (solid), mean Divergence D and pressure velocity ω from ERA5 reanalysis (Hersbach et al., 2020) on 30th March 2022 at 11:00 UTC (approx. time of RF11 C02) and 14:00 UTC (approx. time of P5), averaged over the flight areas (Figure 4.8 (a) and (b)) and height of the boundary layer $z_{\text{BL}} = 0.43$ km obtained from both flights (Section 4.5.5).

therefore this comparison is not representative of the performance of divergence and subsidence in the Arctic but is only shown as a reference to a widely used reanalysis dataset for modelling applications.

The presence of the well-defined low-level D anomaly facilitates (i) comparing the two calculation methods and (ii) assessing the robustness of the divergence measurements. It has been shown with subtropical sounding data that the Line-Integral Method and Regression Method yield similar results (Bony and Stevens, 2019). To revisit this question for the Arctic, we applied both methods to the co-located measurements collected by HALO and P5 on 30th March 2022. Note that the Line-Integral Method can only be applied to sondes

on the outer perimeter of the flight pattern (8 sondes for HALO and 10 sondes for P5). To achieve a fair comparison at the same statistical level, the Regression Method was applied to the same sondes used in the Line-Integral Method calculation. Comparing the profiles from the Regression Method using all sondes to the Regression Method using a reduced number of sondes shows slight differences in Divergence, which appear not to be accounted for when using only perimeter sondes. This suggests that deploying sondes in the centre of a pattern might be beneficial. A more in depth analysis of the statistical significance of the results using different numbers of sondes is performed later in this study.

A comparison between the results obtained from the two flight patterns (RF11 Co2 and P5) reveals satisfactory alignment in divergence measurements above 1 km. However, below this level (around 500 m), a notable disparity emerges in the profiles obtained from P5 measurements, diverging significantly from those extracted from HALO data. This shift cannot be seen in the ERA5 profiles, which seems to not capture this variability in time, but only shows a slight lifting of the divergence profile. The observed shift could be due to changes in the horizontal wind field during the time between the flights, affected by the island of Svalbard in the east of the measurement area, seen in the mean wind profiles of u and v (Figure 4.14). It is possible that these differences could be linked to lee waves, as the flight area is within reach of the observed lee waves (see Figure 4.1). The altitude of the differences in the profiles is similar to that of the Svalbard mountains, which hints an influence of lee waves. Small changes in the flow pattern could cause shifts in the lee waves, potentially explaining the differences. However, this remains speculative, and we cannot confirm it based on our current data.

Only minor discrepancies are apparent in the divergence profiles derived from the Line-Integral and the Regression Methods with both sonde samples, indicating the potential for equal applicability of both methods. This finding confirms the results of [Bony and Stevens \(2019\)](#), who concluded from the similarity, that the divergence measurements are not influenced by sampling errors from small-scale velocity fluctuations, but are reflective of the mesoscale wind.

The pressure velocity profiles diverge with growing height as a result of the upward integration of divergence profiles including small differences. At higher altitudes, the profiles diverge even more, causing differences in the vertical velocity profiles of 0.05 Pa s^{-1} to 0.1 Pa s^{-1} between 3 km and 8 km for the RF11 Co2 profile. [Van der Dussen et al. \(2016\)](#) have shown

that subsidence variations of similar order have a potential impact on simulated atmospheric boundary layer and cloud properties with LES of stratocumulus transitions, and cannot be neglected. The divergence profiles derived from different aircraft data and methods tend to align reasonably well with the average divergence from ERA5, but the profiles at the individual grid points show a high variability (Figure 4.8). All dropsondes deployed during HALO RF11 (32) have been assimilated into ERA5, making a direct comparison difficult and an evaluation of ERA5 based on these results impossible. When compared to data from the ACTIVATE campaign, ERA5 profiles show less agreement and significant temporal variations in the divergence profile (Li et al., 2021), which is in line with a comparison for the other flights (Figure 4.9, Figure 4.10, Figure 4.12 and Figure 4.13), during which less dropsondes have been deployed and not all have been assimilated (10 out of 18 for RF10 and 5 out of 20 for RF18). However, compared to the EUREC⁴A campaign, ERA5 and the observations for all flights showed good agreement (George et al., 2021).

To further characterise the agreement on divergence amplitude and vertical structure between independent datasets, a correlation analysis is performed. HALO data is binned by height and correlated against P5 data for this location (Figure 4.4). As the previous analysis shows similar results for the Regression- and the Line-Integral Methods, the Regression Method will be used for the remainder of the study, as it can utilise all dropsondes and works for varying numbers of dropsondes that do not have to be arranged along a closed line. The data are classified into height bins of 50 m (about 10 data points per bin). The associated correlation coefficient is 0.734. The correlation fit is slightly tilted compared to the one-to-one line, which could be caused by small changes in the wind flow in between measurements or originate from a longer flight time of the P5 aircraft, but lies within the errors of the measurements and the fit.

As described in Section 4.2.1 we performed different flight patterns with different numbers of sondes. In accordance with the outcomes of preliminary model analyses conducted in advance of the campaign, a decision was made to employ 10 sondes for each circular flight pattern, except for RF10 for reasons of operational limitations. In the case of the rectangular flight pattern executed by P5, the strategic decision was made to launch 15 sondes to enhance spatial coverage and comprehensiveness. In Figure 4.5, we present the estimated errors resulting from the linear regression across varying quantities of sondes. This exercise aims to examine the relationship between the number of sondes and the corresponding uncertainty to

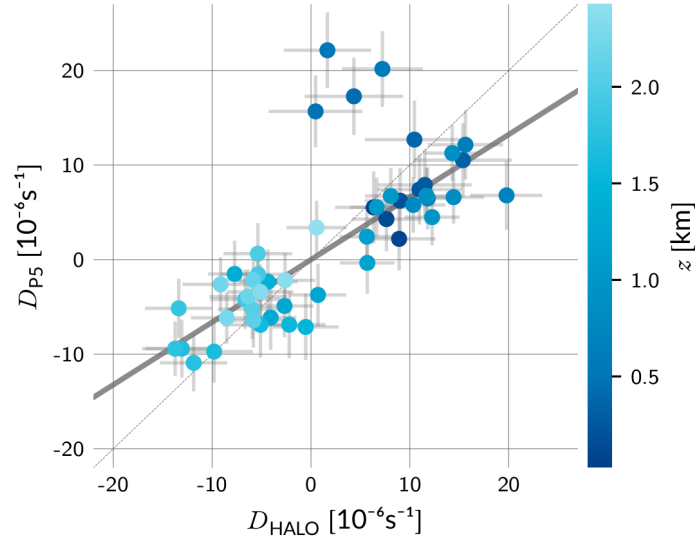


Figure 4.4: Correlation between divergence profiles from HALO (RF11 C02) and P5 measurements on co-located flights for different heights (binned into bins of 50 m) (colour scale) on 30th March 2022 obtained via Regression Method, correlation coefficient $r = 0.734$ and correlation fit $D_{P5} = (0.662 \pm 0.008) \cdot D_{HALO} + (7.826 \cdot 10^{-7} \pm 6.55 \cdot 10^{-13}) \text{ s}^{-1}$ (grey line), both excluding 4 outliers at 500 m (shaded).

obtain the optimal number of sondes required for the method with the potential to streamline costs by minimizing the number of deployed sondes while maintaining sufficiently good results for forthcoming campaigns. For each distinct flight pattern, divergence profiles were computed for every feasible combination of sondes, with a minimum threshold of five sondes. The mean vertical estimated error of all profiles sharing the same sonde count is depicted using a boxplot representation. Given the substantial similarity in results, the data from all three circles of RF11 were combined in one boxplot. Analysing the P5 measurements, we identified a power-law relationship governing the dependence of the estimated error on the number of sondes N :

$$\overline{\text{Err}(D)} = a \cdot N^{-b} + c. \quad (4.8)$$

The corresponding fit parameters are shown in Table 4.1. The displacement of the power-law curve appears to correlate with the synoptic conditions, which were different between RF18 and RF11, with a weak cold air outbreak present during RF10 and RF11, and a strong inversion above the surface in the lowest 500 m and Arctic cirrus on RF18. The estimated error derived from both the P5 and HALO measurements on these specific dates exhibited a notable degree of similarity. The first derivative of the fit functions for all cases reduces by

80 % between $N = 5$ and $N = 10$ and even 85 – 90 % for $N = 12$ sondes, showing a significant convergence in the estimated error of the divergence measurement. Based on the results of this analysis, we recommend using 10–12 sondes per pattern to estimate divergence on a scale of 150 km for Arctic research campaigns. This takes into account the possibility of sonde failures and aligns with the suggestions made by [Bony and Stevens \(2019\)](#) for subtropical regions.

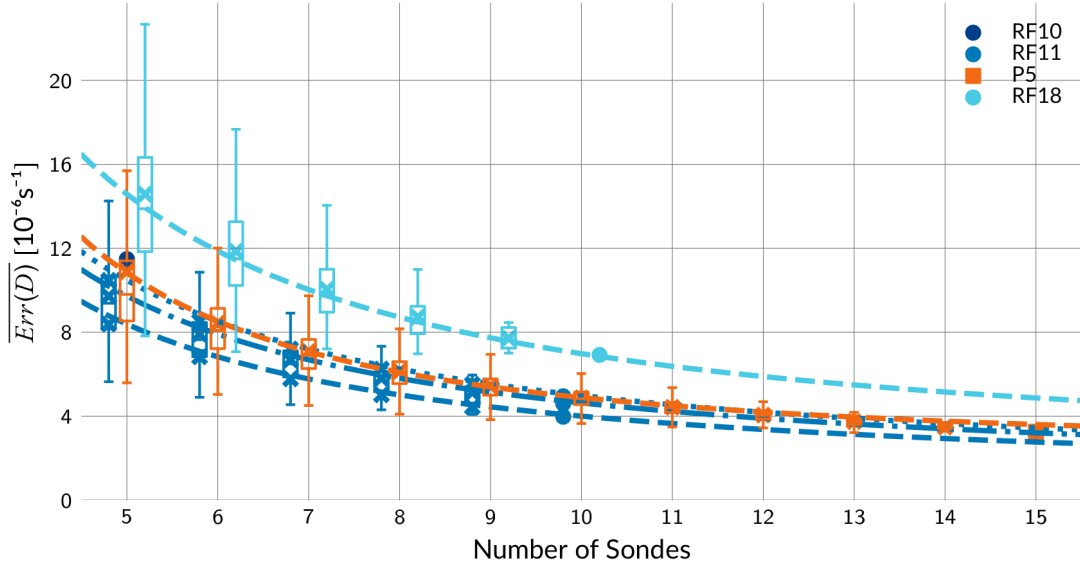


Figure 4.5: Vertically averaged estimated error of divergence $\overline{\text{Err}(D)}$ for all possible combinations of sondes N (min. 5 sondes) from Regression Method (boxplots, outliers not shown, RF11 and RF18 displaced for better visualisation), data from all circles of RF11 combined in one boxplot for better readability. Mean estimated error of all combinations (x), estimated error for maximum number of sondes (bold marker) and power law fit $\overline{\text{Err}(D)} = a \cdot N^{-b} + c$ for RF11 C01 (dotted), RF11 C02 (dashed), RF11 C03 (dash-dotted), RF08, and RF18, see Table 4.1 for fit parameters.

4.4 CONCLUSIONS

In this study data from dropsondes launched in mesoscale synoptic situations during multiple flights of the HALO- $(\mathcal{AC})^3$ campaign in the Norwegian sector of the Arctic were used to calculate divergence profiles and derive mesoscale vertical velocity (subsidence). The twofold objective is to test the robustness of this method at high latitudes and to compare alternative calculation methods. The Regression Method and the Line-Integral Method have demon-

strated an almost equivalent effectiveness at high latitudes, similar to findings from previous campaigns in the subtropics. We conclude a remarkable agreement observed in both amplitude and vertical structure of mesoscale divergence as obtained from disparate and independent sonde datasets collected by different aircraft over the same target area. Minor differences in divergence between the two datasets are likely caused by small changes in the flow between the sampling time of the two flights (approx. 45 min).

An error analysis confirms a non-linear decrease in the estimated error $\overline{\text{Err}(D)}$ with increasing dropsonde numbers explained well by a power-law relationship, similar to results from previous studies in the subtropics. Based on these results, we recommend an optimal range of ten to twelve dropsondes per mesoscale pattern, striking a balance between statistical significance and cost-efficiency, while also accounting for potential sonde failures. This number could be useful for planning future airborne campaigns for measuring mesoscale divergence.

The importance of subsidence as a forcing factor for LES and single column model (SCM) (Mirocha and Kosović, 2010; Neggers, 2015; Young et al., 2018; van der Linden et al., 2019), makes this dataset relevant for future modeling activities. LES and SCMs are frequently used to test or develop physical parameterization schemes while requiring the prescription of a reliable representation of the mesoscale dynamics. Information about the subsidence conditions is fed into these models in the form of divergence or vertical velocity profiles that are either constructed in an idealized manner, or, more frequently simply acquired from reanalysis products, at times in combination with crude estimates from in situ observations. Models tend to be highly sensitive to small changes in the imposed subsidence conditions (Mirocha and Kosović, 2010; van der Dussen et al., 2016; Young et al., 2018) which calls for a more carefully considered source for this kind of forcing information. Even though the ERA5 divergence vertical profile in the area of the collocated patterns is in overall agreement with its observationally constrained counterpart - which in itself could be the result of the assimilation of the abundant dropsonde profiles available for that time period - this is not always the case. Therefore, the use of reanalysis products for forcing SCM or LES models is less than optimal. The dataset presented in this study offers a more realistic depiction of the mesoscale vertical motion, while the use of different derivation methods, flight patterns and amounts of dropsondes provides a meaningful uncertainty range, within which the sensitivity of the models could be transparently investigated.

The results obtained in this study motivate various future research efforts. While the general circulation during the selected flights was relatively stable and slow-changing, the encouraging results obtained for these conditions invite exploring different weather regimes in the Arctic further. Investigating different flight patterns and assessing the impact of non-uniformly spaced dropsondes could further facilitate the broader application of this method for flights with more diverse research objectives. Finally, expanding our knowledge on vertical motion at high latitudes furthers our general understanding of the Arctic's atmospheric dynamics and climate, could guide the coordination of future airborne campaigns in the region, and contributes to the advancement of atmospheric science in general.

4.5 APPENDICES

4.5.1 APPENDIX A - HORIZONTAL DISPLACEMENT OF DROPSONDES

Figure 4.6 shows vertical profiles of the horizontal displacement of the dropsondes from the location of their release, to quantify the drift.

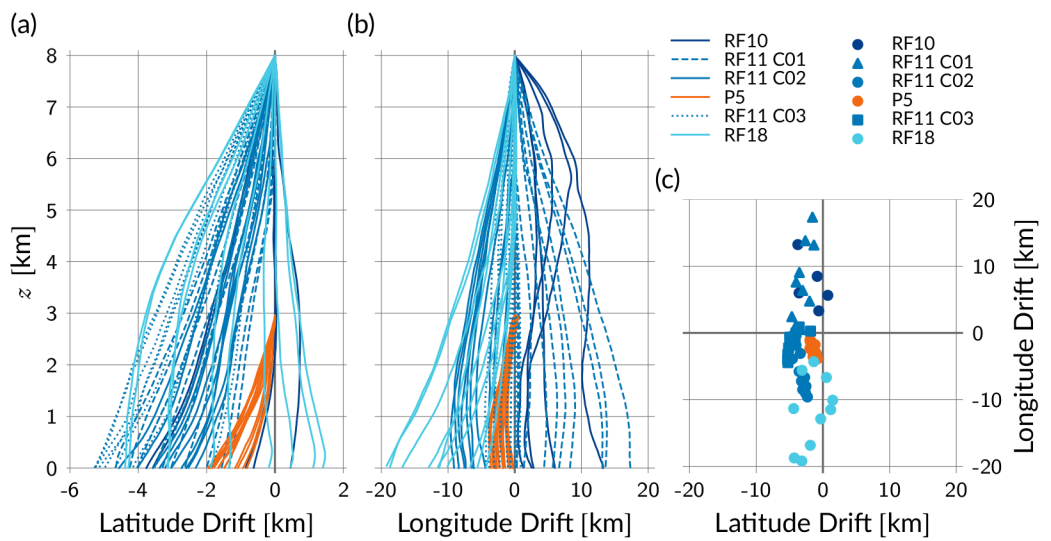


Figure 4.6: Vertical profiles of horizontal displacement of dropsondes from the location of release at t_0 in Latitude $\text{lat} - \text{lat}(t_0)$ (a) and Longitude $\text{lon} - \text{lon}(t_0)$ (b), maximum horizontal displacement (c).

4.5.2 APPENDIX B - VARIABILITY IN THE LINE-INTEGRAL METHOD

Figure 4.7 shows the relative frequency distribution of the absolute differences in the divergence and pressure velocity profiles calculated with the line-integral method, when disregarding or accounting for potential deformation of the patterns with height.

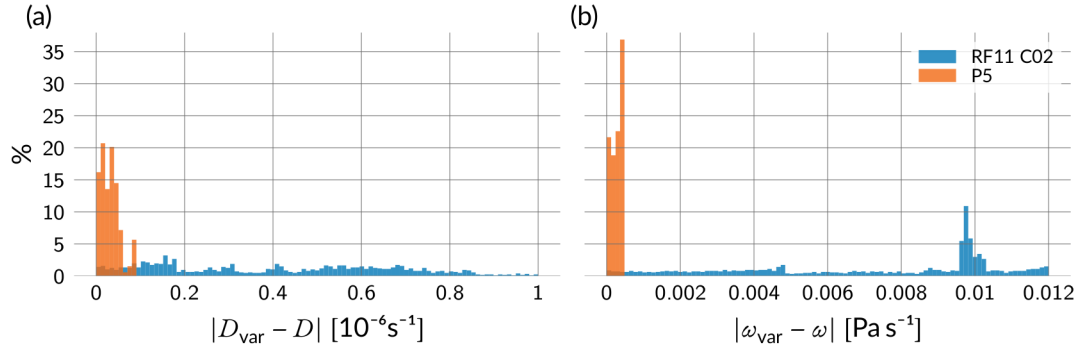


Figure 4.7: Relative frequency distribution of the absolute differences in the divergence (left panel) and pressure velocity (right panel) profiles calculated with the line-integral method, when disregarding or accounting for potential pattern deformation with height (D , ω and D_{var} , ω_{var} respectively). For the RF11 C02 flight (blue), pattern deformation includes changes in both the circle radius and relative dropsonde distances; for P5 (orange) only changes in the area enclosed by the flight track are considered.

4.5.3 APPENDIX C - ERROR ESTIMATION

The error of Divergence $\text{Err}(D)$ and vertical velocity $\text{Err}(\omega)$ is obtained through Gaussian Error Propagation (Bronshtein et al., 2007) of the standard error of the velocity measurements and the distance from the centre of the flight pattern:

$$\text{Err}(\xi(z)) = \frac{\sqrt{\sum_{k=1}^{N_{\text{Sondes}}} (\xi(z, k) - \bar{\xi}(z))^2}}{N_{\text{Sondes}}} \quad (4.9)$$

for $\xi \in \{u, v\}$. The error estimation of the horizontal gradient ($\chi \in \{\vartheta, \phi\}$) is then given by

$$\text{Err}\left(\frac{\partial \xi}{\partial \chi}\right) = \frac{\text{Err}(\xi)}{\sqrt{\sum_{k=1}^{N_{\text{Sondes}}} (\chi - \bar{\chi})^2}} \quad (4.10)$$

and the error on divergence $\text{Err}(D)$ is calculated as

$$\text{Err}(D) = \frac{1}{\sin(\bar{\vartheta})} \sqrt{\left(\text{Err}\left(\frac{\partial u}{\partial \phi}\right)\right)^2 + \left(\text{Err}\left(\frac{\partial v}{\partial \vartheta}\right)\right)^2} \quad (4.11)$$

and vertical velocity $\text{Err}(\omega)$ for each height z_i is calculated as

$$\text{Err}(\omega(z_i)) = -g\bar{\rho} \int_0^{z_i} \text{Err}(D(z)) \, dz \quad (4.12)$$

with $g = 9.81 \, \text{m s}^{-2}$ and the mean air density $\bar{\rho}$.

4.5.4 APPENDIX D - PROFILES FROM REGRESSION METHOD AND LINE-INTEGRAL METHOD WITH ERA5 FOR ALL GRID BOXES

Figure 4.8-4.13 show a comparison of vertical profiles of temperature and specific humidity, divergence, and pressure velocity of ERA5 to the observations with the line-integral method and the regression method for all flights.

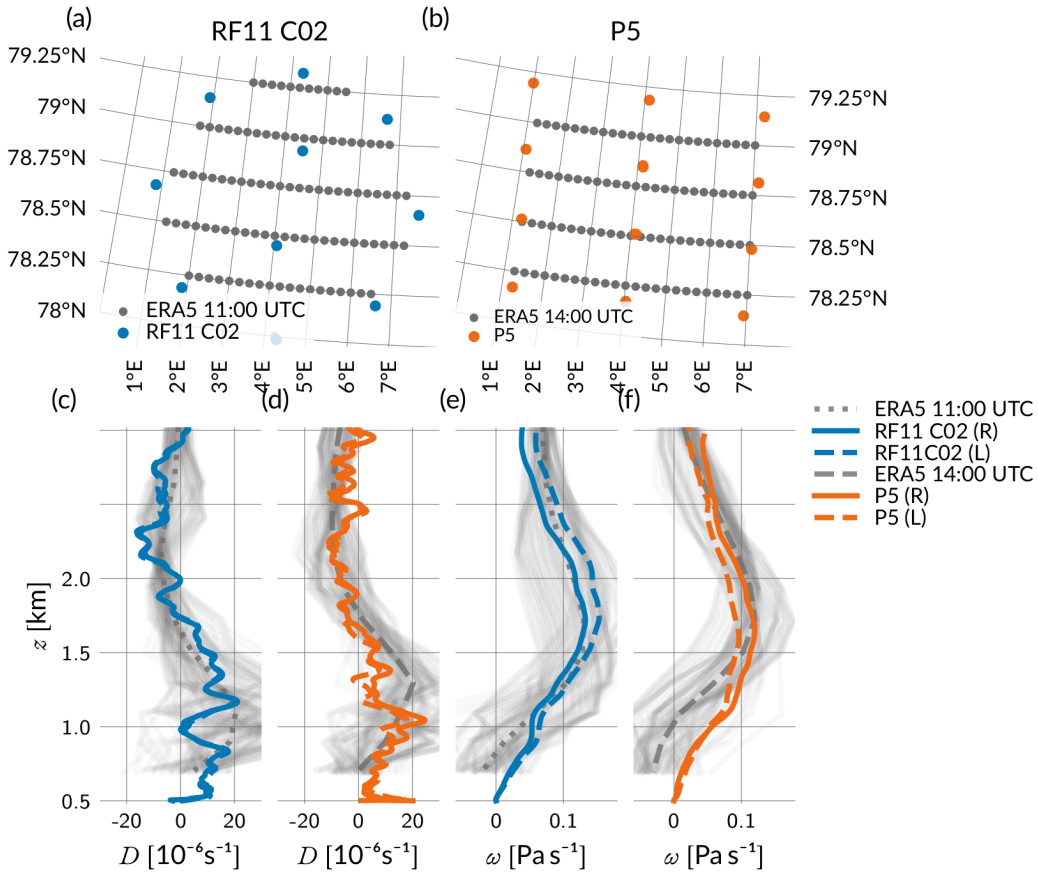


Figure 4.8: Dropsonde locations of RF11 C02 (a) and P5 (b) and corresponding grid points of ERA5 (29th March 2022 11:00 UTC (a) and 30th March 2022 14:00 UTC (b)) inside the sampling area, vertical profiles of divergence D from RF11 C02 (c) and P5 (d) and pressure velocity ω from RF11 C02 (e) and P5 (f) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method using all dropsondes (solid (R,all)) and Line-Integral Method (dashed (L)).

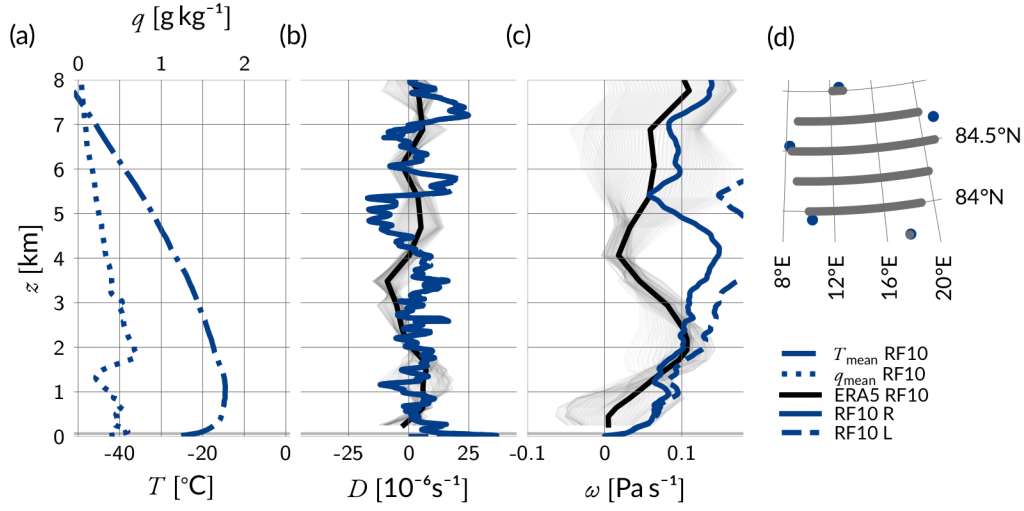


Figure 4.9: Vertical profiles of temperature T and specific humidity q (a), vertical profiles of divergence D (b) and pressure velocity ω (c) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method (blue) and Line-Integral Method (blue, dashed), Dropsonde locations of RF10 and grid points of ERA5 (29th March 2022 14:00 UTC) inside the sampling area (d).

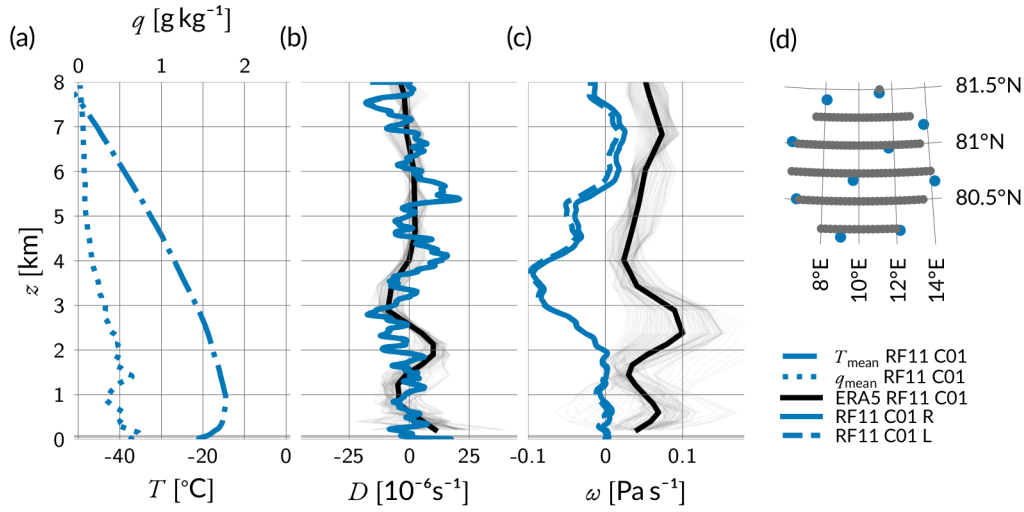


Figure 4.10: Vertical profiles of temperature T and specific humidity q (a), vertical profiles of divergence D (b) and pressure velocity ω (c) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method (blue) and Line-Integral Method (blue, dashed), Dropsonde locations of RF11 C01 and grid points of ERA5 (30th March 2022 10:00 UTC) inside the sampling area (d).

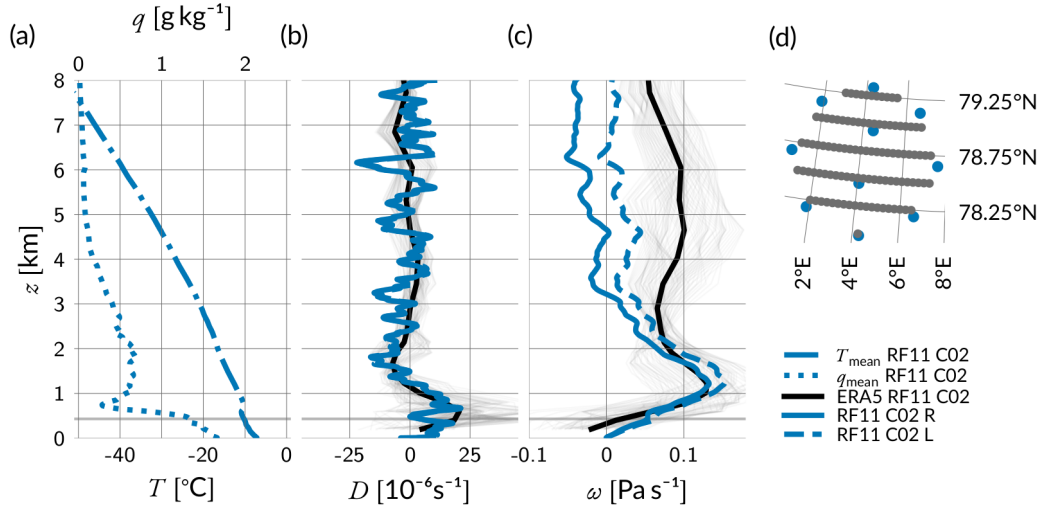


Figure 4.11: Vertical profiles of temperature T and specific humidity q (a), vertical profiles of divergence D (b) and pressure velocity ω (c) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method (blue) and Line-Integral Method (blue, dashed), Dropsonde locations of RF11 C02 and grid points of ERA5 (30th March 2022 11:00 UTC) inside the sampling area (d).

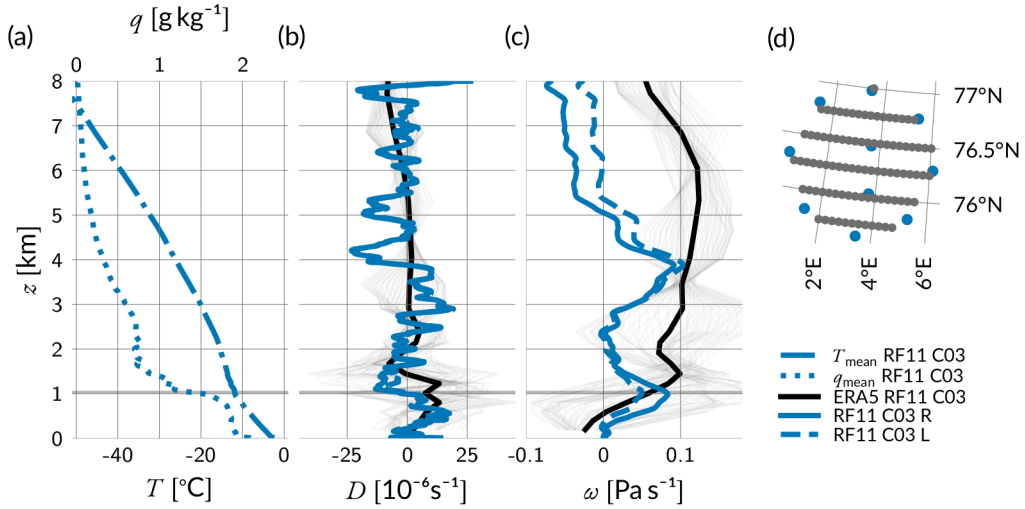


Figure 4.12: Vertical profiles of temperature T and specific humidity q (a), vertical profiles of divergence D (b) and pressure velocity ω (c) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method (blue) and Line-Integral Method (blue, dashed), Dropsonde locations of RF11 C03 and grid points of ERA5 (30th March 2022 12:00 UTC) inside the sampling area (d).

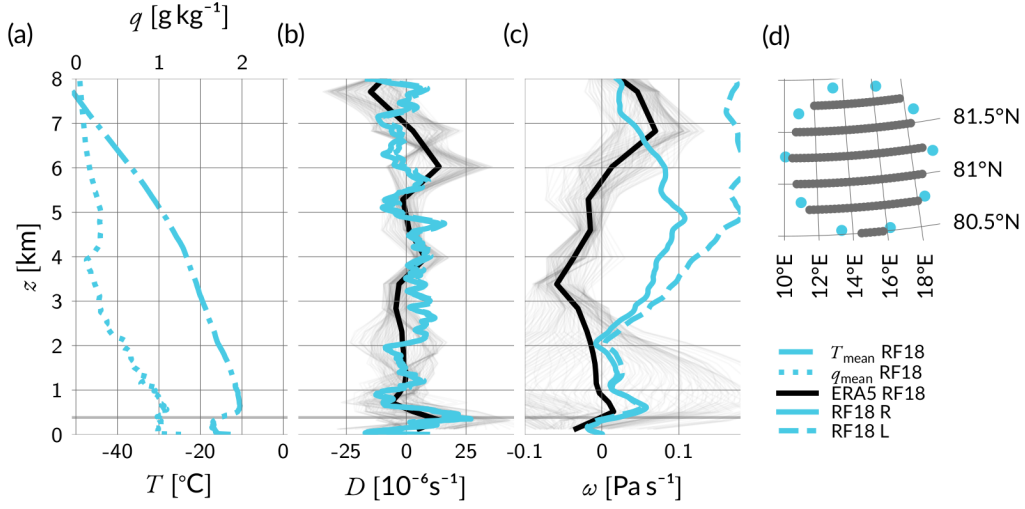


Figure 4.13: Vertical profiles of temperature T and specific humidity q (a), vertical profiles of divergence D (b) and pressure velocity ω (c) at all grid points of ERA5 within the sampling area (light grey), averaged over all ERA5 grid points within the sampling area (black) and from dropsondes obtained with Regression Method (blue) and Line-Integral Method (blue, dashed), Dropsonde locations of RF18 and grid points of ERA5 (12th April 2022 10:00 UTC) inside the sampling area (d).

4.5.5 APPENDIX E - BOUNDARY LAYER HEIGHT CALCULATION

The boundary layer height z_{BL} is calculated as the height, where the Bulk-Richardson Number Ri_b exceeds a critical threshold of $Ri_{b,\text{crit}} = 0.25$ following the implementation in IFS Cy43r1 (ECMWF, 2016). The moist static energy s_v is calculated for each level n as

$$s_{v,n} = c_p T_n (1 + \varepsilon q_n) + g z_n \quad (4.13)$$

with the specific heat capacity of air $c_p = 1003 \text{ J g}^{-1} \text{ K}^{-1}$, the temperature at each level T_n in K, the specific humidity at each level q_n , the gravitational constant $g = 9.81 \text{ m s}^{-2}$, the height of each level z_n in m and $\varepsilon = 0.61$. Together with the squared wind velocity $\vec{v} = (u, v)$ at each level

$$|\vec{v}_n|^2 = u_n^2 + v_n^2 \quad (4.14)$$

the Bulk-Richardson Number Ri_b is calculated.

$$Ri_{b,n} = z_n \frac{2g(s_{v,n} - s_{v,0})}{(s_n + s_0 - g(z_n + z_0))|\vec{v}|^2} \quad (4.15)$$

The boundary layer height is then $z_{\text{BL}} = z (Ri_{b,n} = Ri_{b,\text{crit}})$.

4.5.6 APPENDIX F - MEAN PROFILES OF HORIZONTAL WIND

Figure 4.14 shows vertical profiles of the horizontal wind velocities and hodographs for the colocated flights.

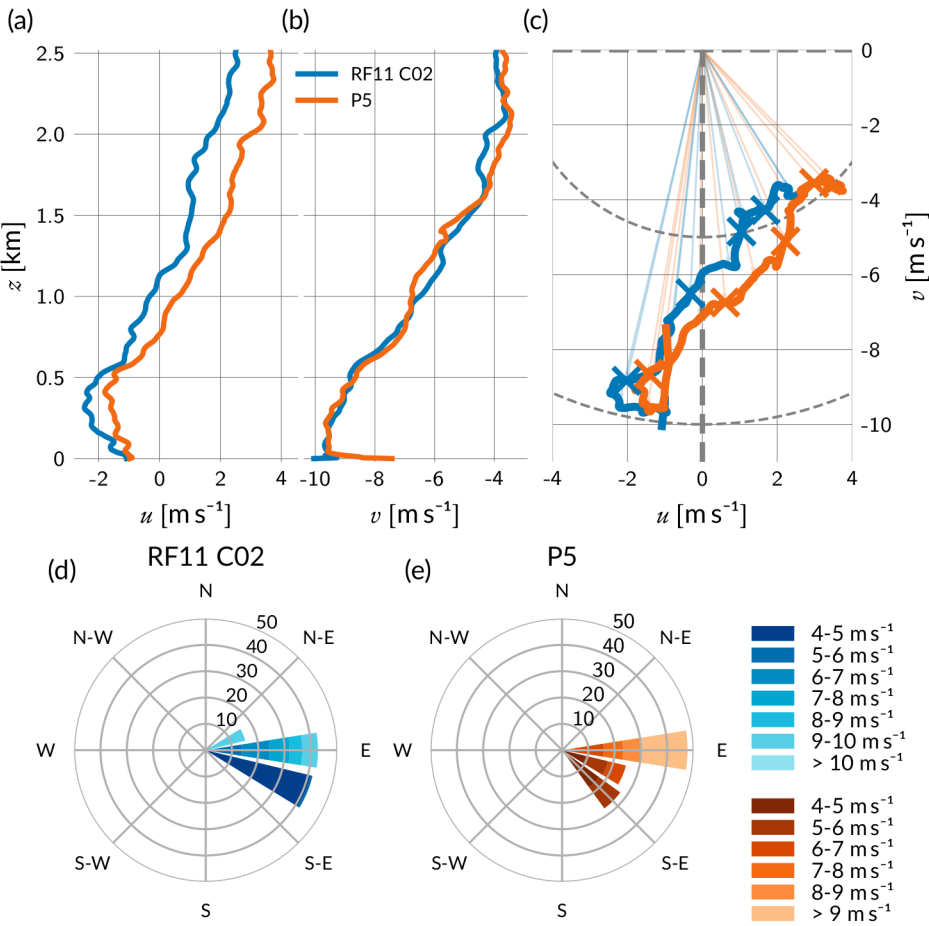


Figure 4.14: Vertical profiles of horizontal wind velocities u (a) and v (b), hodograph (c) for co-located flights RF11 C02 and P5 and windrose plots for RF11 C02 (d) and P5 (e).

4.5.7 APPENDIX G - FIT PARAMETERS OF ERROR STATISTICS

Table 4.1 shows fit parameters of error statistics for RF11 C01, RF11 C02, RF11 C03, P5, and RF18.

Table 4.1: Fit parameters of error statistics for RF11 C01, RF11 C02, RF11 C03, P5 and RF18 (see Figure 4.5), fitted to a power law function for the estimated error of Divergence D : $\text{Err}(D) = a \cdot \frac{1}{N^b} + c$, where N is the number of sondes and a , b , and c are the fit parameters.

Flight	a [10^{-6}s^{-1}]	b	c [10^{-6}s^{-1}]
RF11 C01	74.64 ± 1673.96	1.30 ± 0.29	1.21 ± 8.84
RF11 C02	59.15 ± 680.22	1.29 ± 0.19	0.98 ± 3.72
RF11 C03	69.37 ± 994.72	1.30 ± 0.20	1.15 ± 5.16
P5	139.08 ± 385.05	1.74 ± 0.01	2.35 ± 0.08
RF18	105.75 ± 6536.18	1.32 ± 0.55	1.88 ± 30.32

ACKNOWLEDGEMENTS

We gratefully acknowledge the funding by the Deutsche Forschungsgemeinschaft (DFG, German Research Foundation) - Project 268020496 - TRR 172, within the Transregional Collaborative Research Center “Arctic Amplification: Climate Relevant Atmospheric and Surface Processes, and Feedback Mechanisms (AC)³”. We are further grateful for funding of project grant 316646266 by the Deutsche Forschungsgemeinschaft (DFG, German Research Foundation) within the framework of Priority Programme SPP 1294 to promote research with HALO. We also gratefully acknowledge the funding by the Deutsche Forschungsgemeinschaft (DFG, German Research Foundation) - Project 442649391 - Observing local and Remote Controls on Arctic Air mass evolution (ORCA2). We acknowledge support from the Swedish Research Council (VR, project 2020-04064). We appreciate the financial contributions of Alfred Wegener Institute (AWI, Bremerhaven), German Aerospace Centre (DLR, Oberpfaffenhofen), Max Planck Institute for Meteorology (MPI-M, Hamburg). Pilots and technicians that have flown and operated the airplanes are thanked for all their enormous efforts. We thank Marcus Klingebiel and Davide Ori who planned and executed the research flights and the operators of the dropsonde systems Sebastian Schmidt, Imke Schirmacher and Anja Schwarz for creating the database for this study. We acknowledge the use of imagery from the Worldview Snapshots application (<https://wvs.earthdata.nasa.gov>), part of the Earth Observing System Data and Information System (EOSDIS). ChatGPT has been used to help create first versions of this manuscript.

5

Impacts of Mesoscale Atmospheric Subsidence on Cloud Glaciation and Decoupling in Arctic Marine Cold Air Outbreaks

The first part of this chapter will address the second research question:

CAN A LAGRANGIAN LARGE EDDY SIMULATION, CONSTRAINED WITH OBSERVATION-DERIVED LARGE-SCALE FORCINGS SUCH AS SUBSIDENCE, REALISTICALLY REPRODUCE AN OBSERVED MARINE COLD AIR OUTBREAK?

A Lagrangian large eddy simulation is initialised and constrained with surface observations and observation-derived large-scale forcings, including subsidence, obtained in the previous

chapter. The simulation is subsequently evaluated against independent airborne remote-sensing observations.

Afterwards, the third research question is addressed:

WHAT IS THE IMPACT OF ATMOSPHERIC SUBSIDENCE ON CLOUD
GLACIATION AND BOUNDARY LAYER DECOUPLING DURING
ARCTIC MARINE COLD AIR OUTBREAKS?

Building on the validated modelling framework, a series of sensitivity experiments is conducted to isolate the role of atmospheric subsidence in Arctic MCAOs.

This work is currently under revision with the Journal of Atmospheric Chemistry and Physics of the European Geosciences Union (EGU), where it has been published as a preprint under the Creative Commons Attribution 4.0 International license. The format has been changed to match this thesis.

Paulus, F., Müller, J. J., Kirbus, B., Sodemann, H., van Gelder, L., Walbröl, A., Wendisch M., Neggers R. A. J.: Impacts of mesoscale atmospheric subsidence on cloud glaciation and decoupling in Arctic marine cold air outbreaks, EGU sphere [preprint], <https://doi.org/10.5194/egusphere-2026-554>, 2026.

AUTHOR CONTRIBUTION

FP ran and evaluated the simulations and prepared the manuscript. JM determined the surface temperature from VELOX; BK calculated the Lagrangian trajectories; LvG and AW calculated the integrated water vapour from HAMP and the ice water path from MiRAC, respectively; and HS contributed the CMET observations, and MW and RN revised various intermediate versions of the manuscript.

CODE AND DATA AVAILABILITY

The pre-processed dropsondes data are available at [George et al. \(2024\)](#), and the regression processing followed [Paulus et al. \(2024\)](#); [Paulus and Karalis \(2023\)](#). The integrated water vapour and liquid water path retrieved from HAMP are available at [Walbröl et al. \(2026a\)](#)

and Walbröl et al. (2026b), respectively. The MiRAC observations will be published to the Pangaea repository soon and will be openly available until acceptance of the manuscript, at the latest; until then, the data can be obtained from the corresponding author upon request. Radar reflectivities from MiRAC used for the retrieval of ice water path are available at Mech et al. (2024) and can be accessed via the ac3airborne tool (Mech et al., 2022). VELOX two-dimensional cloud-top and surface brightness temperatures with 1 Hz temporal resolution, derived at flight altitude, are available from Schäfer et al. (2022). The CMET data is also being prepared for publication and will be publicly available on the Pangaea repository until acceptance of this manuscript. The model code, input and configuration files, generated forcing files, and selected model output are provided in Paulus (2026b). The forcing files are compatible with typical single column model (SCM) forcings, as DALES is run with horizontally homogeneous initial and boundary conditions. The DALES version used in this study is 4.3 with an extension for mixed-phase microphysics; the specific version applied here is included in Paulus (2026b), while the official DALES repository is available at Arabas et al. (2021). ERA5 datasets were retrieved from the Copernicus Climate Change Service (C3S) Climate Data Store (CDS) at <https://cds.climate.copernicus.eu/>.

IMPACTS OF MESOSCALE ATMOSPHERIC SUBSIDENCE ON CLOUD GLACIATION AND DECOUPLING IN ARCTIC MARINE COLD AIR OUTBREAKS

ABSTRACT

The impact of mesoscale vertical motion on the thermodynamic, microphysical, and convective transformations of marine cold air outbreaks (MCAOs) is still largely unknown, partly due to scarce high-resolution observations in upstream Arctic regions. Therefore, this study investigates the effects of mesoscale subsidence on the evolution of the atmospheric boundary layer (ABL), cloud phase, and precipitation for a case study of a shallow MCAO observed in the Fram Strait in March 2022 during the HALO-($\mathcal{AC}$)³ campaign. Quasi-Lagrangian large eddy simulations (LESs) are conducted with observational initialisation and larger-scale forcing, based on airborne in-situ and remote-sensing measurements. The LES control simulation accurately reproduces the measured thermodynamic ABL structure and the temporal evolution of the observed air mass moving over the Arctic sea-ice onto the open ocean. Specifically, the measured ABL height, integrated water vapour, and cloud water paths are well represented by the LES. Sensitivity experiments using the LES with prescribed subsidence reveal that weaker subsidence substantially alters the evolution of cloud phase during the MCAO, featuring a deeper ABL and an earlier onset of cloud glaciation. This study shows that ABL internal decoupling plays a key role in this process. Decoupling occurs sooner under weaker mesoscale subsidence, triggering convective graupel formation that subsequently intensively converts liquid water droplets. This strong link between glaciation and decoupling arguably explains the typical evolution of the cloud liquid water path observed in many MCAOs. These results provide a process-based framework for interpreting the role of large-scale vertical motion in Arctic air mass transformations.

5.1 INTRODUCTION

During marine cold air outbreaks, cold and dry Arctic air masses move southward over the cold sea-ice surface towards the warmer open ocean, where they heat up and pick up humidity from below. These MCAOs are an essential phenomenon in the Arctic climate system that frequently occurs over extended marine regions at northern high latitudes (Fletcher et al., 2016a). Additionally, as these air masses cross the sea-ice edge and move over open ocean, the strong near-surface temperature contrast between the warmer open ocean surface and the overlaying colder air generates intense turbulent and convective mixing, producing large sensible heat fluxes (Shapiro et al., 1987a; Renfrew and Moore, 1999; Papritz et al., 2015; Kirbus et al., 2023a). This vertical turbulent exchange rapidly transforms the air mass from below, deepening the atmospheric boundary layer (ABL) through convection. As a result, spatially organised cloud structures evolve in the ABL that initially appear as cloud streets (Atkinson and Wu Zhang, 1996; Lenschow and Agee, 1976; Etling and Brown, 1993; Kirbus et al., 2024; Klingebiel et al., 2025). As the air mass continues to respond to the warm open ocean surface, lower-level warming and moistening, together with persistent vertical surface energy flows, precipitation, and cloud microphysical processes such as riming and secondary ice production, may cause a transition from cloud streets to deeper, broken cloud fields (Tornow et al., 2021; Murray-Watson et al., 2023; Schirmacher et al., 2024). These evolving cloud and ABL characteristics modulate cloud-radiative interactions, determine precipitation formation, increase surface winds, and can generate polar lows (Kolstad, 2017; Terpstra et al., 2021; Mateling et al., 2023). Through this ocean-atmosphere coupling, MCAOs can also influence atmospheric and oceanic circulations (Rasmussen et al., 1992; Isachsen et al., 2013; Papritz and Spengler, 2017; Marcheggiani and Spengler, 2025), which are linked to feedback mechanisms of Arctic amplification. The air mass transports during MCAOs and the corresponding transformation processes may also contribute to heat and moisture exchange between the Arctic and mid-latitudes (Pithan et al., 2018).

Despite extensive prior studies, key aspects of MCAO evolution remain poorly understood, particularly the onset of ABL decoupling in the subcloud layer, which accelerates stratocumulus breakup and promotes transitions between cloud regimes (Bretherton and Wyant, 1997; McCoy et al., 2017; Lloyd et al., 2018; Zhou et al., 2015). Furthermore, precipitation and cloud microphysical developments that are influenced by riming and secondary ice production, reducing vertical mixing and altering ABL energy budgets, are not well repre-

sented in models (Abel et al., 2017; Karalis et al., 2022; Wu et al., 2025). Persistent observational gaps, particularly in remote upstream regions of MCAO pathways, continue to limit our understanding of these processes and their role in organising MCAO transitions. Consequently, major questions remain regarding how dynamical and microphysical processes support the transition to open-cell convection and how these should be represented in numerical models (Tomassini et al., 2017; de Roode et al., 2019).

The role of mesoscale vertical motion in MCAO evolution, and mixed-phase clouds in particular, is complex and not yet fully understood. Previous studies have shown that subsidence in surface-coupled convective layers strengthens the capping temperature inversion by counteracting top entrainment, thus increasing ice and liquid water path below. In return, enhanced cloud-top radiative cooling and convective overturning under stronger subsidence can increase precipitation production in both rain and snow, stabilising stratiform cloud layers by suppressing cloud-top ascent and prolonging cloud persistence (Young et al., 2018). In contrast, Neggers et al. (2019) found that strong subsidence can cause mixed-phase cloud layers over the central Arctic sea-ice to collapse, shifting thermal-infrared radiative cooling to the surface and hindering cloud reformation. These findings highlight that mesoscale subsidence can strongly modulate mixed-phase cloud layers, but it remains unknown how this modulation depends on the ABL structure. The involved interactions span a wide range of spatial scales, from mesoscale to synoptic subsidence. It covers processes such as turbulent entrainment, microphysical processes, and aerosol particle evolution (Tornow et al., 2021, 2023). A prime cause for our lack of insight into the impact of mesoscale subsidence on mixed-phase clouds is the distinct observational data gap on this variable at high latitudes. Measuring large-scale vertical motion or horizontal divergence remains difficult, as small errors in horizontal wind gradients lead to large uncertainties when using mass conservation. A more robust method applies regression to dropsonde data released in mesoscale patterns by fast aircraft, pioneered during two aircraft campaigns in the subtropical trades (Bony and Stevens, 2019; Stevens et al., 2021; George et al., 2021). The application of this method to the coherent flow of an Arctic MCAO, can help to narrow this data gap and enhance our understanding of the impacts of subsidence on the development of mixed-phase clouds in growing boundary layers.

The Fram Strait is a hotspot for MCAOs. It is relatively easy to access, making it a frequent target area for atmospheric observational and modelling studies. Although many re-

cent Fram Strait studies are model-based, several field campaigns have filled key data gaps of MCAO processes. For example, the Cold air Outbreaks in the Marine Boundary Layer Experiment (COMBLE) campaign in 2019 and 2020 observed several MCAOs with two permanent observational sites that provided continuous measurements in the Fram Strait region (Geerts et al., 2022), and the Springtime Atmospheric Boundary Layer Experiment (STABLE) campaign in March 2013 provided airborne observations with a research aircraft in the Fram Strait (Michaelis et al., 2022). The HALO- $(\mathcal{AC})^3$ campaign in 2022 sampled numerous MCAOs with multiple co-located aircraft, probing upstream and downstream regions (Walbröl et al., 2024; Wendisch et al., 2024; Ehrlich et al., 2025; Wendisch et al., 2025). During HALO- $(\mathcal{AC})^3$, the mesoscale dropsonde pattern technique was applied to sample subsidence at multiple points along a MCAO trajectory for two consecutive days, producing the first robust high-latitude subsidence dataset of its kind (Paulus et al., 2024). This unique dataset creates new opportunities for investigating the role of subsidence in MCAO evolution.

The main objective of this study is to use quasi-Lagrangian large eddy simulation (LES), driven by HALO- $(\mathcal{AC})^3$ observational data, to gain insight into the impacts of mesoscale subsidence on MCAOs. For a particular case study, the LES follows an observed MCAO along its two-day southbound trajectory, initialised with in-situ dropsonde profiles from the High Altitude and Long Range Aircraft (HALO) in the far north. The simulated air mass evolves freely, with large-scale forcings, including vertical motion, geostrophic wind, and advection tendencies, derived from circular dropsonde pattern observations and prescribed along the trajectory (Paulus et al., 2024). The simulated ABL and cloud properties are compared against independent downstream aircraft observations.

Given this observationally constrained setup, we ask how well the Lagrangian LES reproduces the independently observed cloud development. After evaluating the model performance, we conduct perturbation experiments on the mesoscale subsidence to assess the sensitivity of the MCAO, focusing on the evolution of the ABL structure, cloud mass, thermodynamic phase, and surface precipitation. This study aims to discuss how microphysical and dynamical properties and processes interact in their response to perturbed larger-scale subsidence forcing within the simulated MCAO system.

5.2 OBSERVATIONS

5.2.1 THE HALO-($\mathcal{AC}$)³ CAMPAIGN

The HALO-($\mathcal{AC}$)³ field campaign took place from March to April 2022 in the Arctic west of Norway and the Fram Strait. The main objective of this campaign was to understand the transformation of air masses between the Arctic and the mid-latitudes and to investigate the influence of these interactions on ABL processes and cloud formation. To this end, a quasi-Lagrangian sampling strategy was applied during the campaign, in which the transformation of individual air parcels of an air mass was tracked along their trajectories using three research aircraft flying at different altitudes (HALO, Polar 5 (P5) and Polar 6 (P6)) (Wendisch et al., 2024). An overview of the synoptic situation during the HALO-($\mathcal{AC}$)³ campaign is provided by Walbröl et al. (2024).

This study focuses on a case study analysing two HALO Research Flights (RF10 and RF11) conducted on the two days of 29th March 2022 and 30th March 2022 respectively (Figure 5.1). On these two days, a weak MCAO with an MCAO index of four was investigated (Papritz and Spengler, 2017; Walbröl et al., 2024) (Section 5.7.1). Both days were characterised by a persistent, steady, and relatively slow northerly wind of about 3 to 10 m s⁻¹ measured at 10 m altitude by dropsonde observations. These MCAO characteristics define a weak MCAO compared to climatological MCAO values from the Fram Strait (Slättberg et al., 2025)). The slowly moving air mass was in the range of HALO during the investigated period of two days. The observed MCAO case occurred during a cold period between two intervals of strong MCAOs from 21–26 st March 2022 and 1–2 st April 2022 (Walbröl et al., 2024).

The synoptic situation in the Fram Strait was characterised by a high-pressure system with its centre located at the east coast of Greenland and a low-pressure area in the Barents Sea. The cloud field was dominated by shallow convective mixed-phase clouds, with clearly defined street features near the sea-ice edge. These streets broke up further downstream of the flow in the area of Co3 (Figure 5.1), where the cloud cover increased significantly. A notable cloud-free area was located on the west coast of Svalbard (Figure 5.2), which can be attributed to the foehn effect during easterly flow over Svalbard, similarly observed during other field campaigns (e.g., Shestakova et al., 2022).

The HALO flight plans on these two days were created by tracking the location of an air mass sampled on the first day at a location in the high north using trajectory estimates based on forecast data. At four locations along the trajectory, the air mass was sampled using a circular mesoscale dropsonde pattern, each time with a diameter of 150 km. This technique is inspired by [Lenschow et al. \(2007\)](#) and has previously been applied in the North Atlantic and subtropical region by [Stevens et al. \(2019, 2021\)](#) to measure mesoscale divergence and subsidence profiles. The centre of the initial circle (IC) (Figure 5.1) on the first day at 14:00 UTC was located at (84.4 °N, 13.6 °E) and featured only five dropsondes, a number limited by airspace restrictions. On the second day, three further circles were flown along the predicted southern trajectory of the air mass. The circles sampled during the two subsequent days each featured eight evenly distributed dropsondes, with two additional dropsondes launched in their inner areas. Only one of the day-two circles, (Co2) at (78.6 °N, 4.0 °E, at 11:30 UTC), was located at the exact predicted location of the air mass as sampled on the previous day at the IC location. As shown in Figure 5.1, this circle was situated over open ocean. Two further circles were flown on day two, one (Co1) further upstream over sea-ice (80.8 °N, 10.7 °E, at 10:30 UTC), and another (Co3) was flown at a downstream location at a similar distance (76.3 °N, 3.3 °E, at 12:30 UTC). The P5 aircraft took additional measurements in the Co2 area, shortly after the departure of the HALO aircraft from this area (at 12:00 UTC), using a rectangular flight pattern with three cross-flow legs. Additional measurements were conducted by a controlled meteorological balloon (CMET) launched from Ny-Ålesund, as part of the ISLAS2022 field experiment, which took place at the same time as the HALO-(AC)³ campaign.

5.2.2 OBSERVATIONAL DATASETS

This study uses a combination of in-situ and remote sensing observations from the HALO-(AC)³ campaign to constrain the LES experiments and evaluate the model performance. A comprehensive overview of the measurements collected during HALO-(AC)³ is described by [Ehrlich et al. \(2025\)](#), and a summary of the datasets used is presented in Table 5.1, including the variable names, associated instruments, and references to more detailed technical and scientific descriptions.

The deployed Vaisala RD-41 dropsondes provide vertical profile measurements of the thermodynamic state, including air temperature at a resolution of 0.01 K, air humidity at a

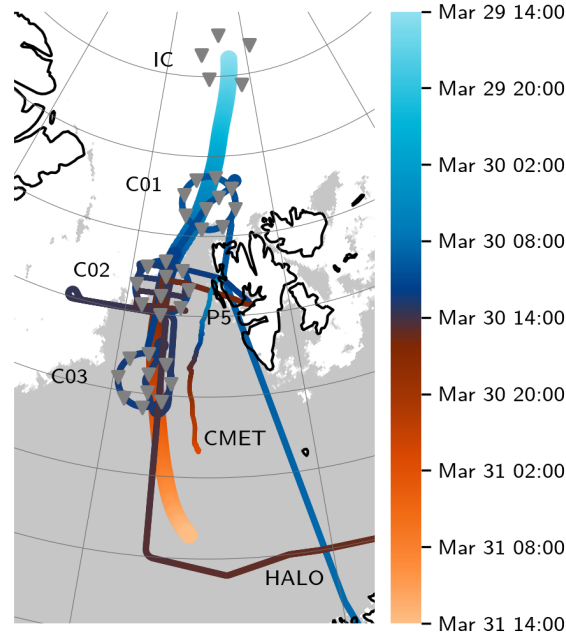


Figure 5.1: Map of the simulation trajectory for the Lagrangian LES (bold line) initiated in the centre of the circular flight leg on 29th March 2022 of the HALO-($\mathcal{AC}$)³ campaign, together with the flight paths of the HALO and P5 aircraft on 30th March 2022 and the trajectory of the CMET. Locations where dropsondes were released in IC on 29th March 2022 and in C01, C02 and C03 of 30th March 2022 from HALO are indicated by ▼. The sea-ice concentration retrieved from satellite observations (Spreen et al., 2008) is indicated in grey and white, and the colour bar denotes the time (UTC) along the trajectory and flight paths.

resolution of 0.1 % and air pressure at a resolution of 0.01 hPa at frequencies between 2 and 4 Hz and descend at speeds between 10 and 20 m s⁻¹ (Vaisala, 2023). The measured profiles are interpolated to a uniform vertical grid of 5 m spacing. The horizontal wind speed profiles were determined from GPS data with an estimated accuracy of 0.1 m s⁻¹ (Hock and Franklin, 1999; Wang et al., 2015). These observations were used to derive vertical profiles of subsidence (Figure 5.3), geostrophic wind, and advective tendencies of temperature, humidity, and wind through a regression algorithm applied to the dropsondes deployed in each circle. A comprehensive description of the calculation technique is provided by Paulus et al. (2024).

To constrain the surface skin temperature T_{skin} , airborne measurements of the thermal infrared imager on board HALO VELOX (Video airborne Longwave Observations within siX channels; Schäfer et al. (2022)) were used, which records two-dimensional fields (640 pixel $\times$ 512 pixel) of the thermal infrared brightness temperature. At typical HALO

flight altitudes, this corresponds to a horizontal resolution of 10 m with a temporal resolution of 1 Hz for a field-of-view of $6.4 \text{ km} \times 5.1 \text{ km}$. From these brightness temperature images, we retrieve T_{skin} and sea-ice concentration along the circular flight legs of HALO with accuracies of 1.2 K and 5 %, respectively (Müller et al., 2025). For partly cloudy open-ocean segments in Co2 and Co3, we apply a cloud mask and classify pixels whose retrieved T_{skin} exceeds $-1.8 \text{ }^{\circ}\text{C}$ above the surrounding (cloudy) background as open ocean. The final surface skin temperature is the horizontal mean of each masked 2D-field at a 1 Hz time resolution.

The observational datasets collected during the HALO-(AC)³ campaign provide all atmospheric state variables required to run Lagrangian LES experiments following the air mass as it moves south. A small number of surface parameters, specifically surface albedo, the aerodynamic roughness length for momentum, and the thermal roughness length for heat, could not be observed, but are needed to prescribe the (latitude-dependent) surface boundary condition in the LES experiments. Therefore, these data were retrieved from the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis product version 5 (ERA5; Hersbach et al. (2020)). By combining the dropsonde and VELOX observations with the surface values from ERA5, a complete set of forcing fields for the Lagrangian LES has been compiled.

For model validation, independent datasets also obtained from the HALO-(AC)³ campaign are used. The Airborne Mobile Aerosol Lidar (AMALi) on the P5 aircraft provides measurements of cloud top height, which are compared to the simulated cloud top heights (Stachlewska et al., 2010). The liquid water and ice water paths are derived from the Microwave Radar/Radiometer for Arctic Clouds (MiRAC) on board P5 (Mech et al., 2019). The ice water path is derived from the 94 GHz-radar reflectivity using the $Ze-IWC$ relationship of Hogan et al. (2006). We additionally include integrated water vapour IWV and liquid water path LWP data retrieved from the HALO Microwave Package (HAMP) over open ocean (not available over sea-ice) (Mech et al., 2014). The additional data allows a direct comparison with the quantities computed by the LES.

Further thermodynamic observations were performed using Lagrangian CMET soundings (Voss et al. (2010); Løklingholm et al. (2026), manuscript to be submitted). The balloon was equipped with an air temperature sensor that operates over a range of $-40 \text{ }^{\circ}\text{C}$ to $105 \text{ }^{\circ}\text{C}$ with an interchangeability of $\pm 0.5 \text{ }^{\circ}\text{C}$ and measured relative humidity with respect to water with

an accuracy of 1.5 %. Due to the cold, dry conditions, and the absence of active ventilation of the sensor package on the CMET, actual measurement errors would be substantially larger than the sensor interchangeability, in particular during daytime when the balloon was drifting horizontally, creating an overall warm and dry bias (Løklingholm et al., 2026). This particular CMET-04 was launched in Ny-Ålesund on 30th March 2022 following a trajectory parallel to the simulated trajectory (Figure 5.1), complementing the dropsonde observations and providing additional in-situ validation data within the lower ABL.

5.3 MODEL CONFIGURATION

5.3.1 THE DUTCH ATMOSPHERIC LARGE EDDY SIMULATION

The simulations were performed using the Dutch Atmospheric Large Eddy Simulation (DALES) (Heus et al., 2010), which is an open-source model available at <https://github.com/dales-team/dales>. DALES has been used extensively in previous scientific studies of ABL turbulence and clouds in various regions of the Earth (van der Dussen et al., 2013; Corbetta et al., 2015; Roode et al., 2016; Laar et al., 2019; Reilly et al., 2020), including the Arctic (Neggers et al., 2019; de Roode et al., 2019; Egerer et al., 2021; Chylik et al., 2023; Schnierstein et al., 2024).

To represent Arctic mixed-phase clouds, the DALES code was equipped with the double-moment microphysics scheme of Seifert and Beheng (2006), which is predictive of both the mass and number concentrations of five hydrometeor types (cloud water, cloud ice, rain, snow, and graupel) (Neggers et al., 2019; Chylik et al., 2023). The number concentration of cloud condensation nuclei (CCN) is prognostic, initialised with a constant vertical profile of 100 cm^{-3} within the range observed during the HALO-(AC)³ campaign in the Fram Strait (Wendisch et al., 2024) and in the high Arctic during MOSAiC (Ansmann et al., 2023). The number concentration of ice nucleating particless (INPs) is prescribed at a value of $1 \cdot 10^{-3} \text{ cm}^{-3}$, following Hartmann et al. (2019). Heterogeneous freezing processes are limited to temperatures below -15°C , and the maximum number of ice particles resulting from deposition nucleation is restricted to 200/L, while homogeneous freezing and secondary ice formation via the Hallett-Mossop process remain unchanged as represented in the Seifert and Beheng (2006) model.

Table 5.1: Variables used in the LES for initialisation (I), evaluation (E), boundary condition (BC) and larger scale forcing (F).

Variable	Unit	Description	Instrument	Platform	Reference	Usage
T	K	Air temperature	Dropsonde, CMET	HALO	Vaisala (2023)	I,E
q	kg kg^{-1}	Dpecific humidity		CMET	Voss et al. (2010)	I,E
p	hPa	Pressure	Dropsonde	HALO	Vaisala (2023)	I,E
u, v	m s^{-1}	Horizontal wind components				I
$-\vec{v}_h \cdot \nabla T$	K s^{-1}	Advection of air temperature				F
$-\vec{v}_h \cdot \nabla q$	$\text{kg kg}^{-1} \text{ s}^{-1}$	Advection of specific humidity				F
$-\vec{v}_h \cdot \nabla u$	m s^{-2}	Advection of meridional/zonal wind	Regression from Dropsondes	HALO	Paulus et al. (2024)	F
$-\vec{v}_h \cdot \nabla v$						
u_g, v_g	m s^{-1}	Geostrophic wind				F
w	m s^{-1}	Subsidence				F
T_{skin}	K	Surface skin temperature	VELOX	HALO	Müller et al. (2025)	BC
siC	%	sea-ice fraction				BC
α		Surface albedo				BC
z_0	m	Roughness length for momentum		ERA5	Hersbach et al. (2020)	BC
$z_{0,h}$	m	Roughness length for heat				BC
z_c^t	m	Cloud top height	AMALi	P5	Stachlewska et al. (2010)	E
IWW	kg m^{-2}	Integrated water vapour	HAMP	HALO	Mech et al. (2014)	E
LWP	g m^{-2}	Liquid water path	HAMP, MiRAC	HALO, P5	Mech et al. (2014, 2019)	E
IWP	g m^{-2}	Ice water path	MiRAC	P5	Mech et al. (2019)	E

This microphysical model setup was thoroughly tested by [Schnierstein et al. \(2024\)](#) against a year of in-situ measurements of Arctic clouds in the central Arctic during the MOSAiC expedition.

Further model settings follow the setup described by [Schnierstein et al. \(2024\)](#). They include a radiative transfer scheme interactive with liquid and ice hydrometeors applying a four-stream radiative transfer solver in combination with Monte Carlo spectral integration ([Pincus and Stevens, 2009](#)). The optical properties of ice particles are based on the physical properties of the ice crystals derived from the microphysics scheme ([McFarquhar and Heymsfield, 1998](#); [Fu and Liou, 1993](#); [Baran, 2005](#); [Schnierstein et al., 2024](#)). Resolved advection is calculated using a fifth-order central difference numerical scheme for momentum and a κ -limiter scheme for the other prognostic scalar variables. Subgrid-scale transport uses a prognostic turbulent kinetic energy SFS-TKE scheme ([Deardorff, 1980](#); [Hundsdoerfer et al., 1995](#)). The simulation uses a sponge layer in the upper 25 % to dampen disruptive gravity waves. The surface-fluxes of heat, moisture, and momentum are calculated based on Monin-Obukhov theory using Arctic stability functions derived by [Grachev et al. \(2007\)](#). Horizontally periodic boundary conditions are applied, while large-scale forcings are prescribed as time-dependent tendency profiles to account for advection, subsidence, and horizontal pressure gradients (geostrophic wind vector $\vec{v}_g$).

5.3.2 TRAJECTORY CALCULATION

To set up a Lagrangian LES experiment, the trajectories (pathways) of the air parcels forming the air mass need to be estimated. As described in Section 5.2.1, trajectories for the low-level air parcels were calculated during flight planning using the Lagrangian Analysis Tool (LAGRANTO) ([Sprenger and Wernli, 2015](#)) and based on IFS forecast winds. For this study, this was repeated using three-dimensional wind fields of the ERA5 reanalysis. Air parcels were identified, and their trajectories were calculated starting every 3 km lat/lon within the IC on 29th March 2022 and matching the circle locations on 30th March 2022 ([Kirbus et al., 2024](#)). The best match of the trajectories was found at 870 hPa in the initiation region. For the simulations, a mean trajectory was calculated by averaging all trajectories initiated within the first circle at 870 hPa. Each trajectory is included for every circle region it intersects, but is excluded from further analysis once it exits a circle region without entering the subsequent one. This procedure is illustrated in Figure 5.14.

5.3.3 OBSERVATIONAL FORCING AND BOUNDARY CONDITIONS

The lower boundary condition consists of a prescribed surface skin temperature T_{skin} based on VELOX measurements (Section 5.2.2). These observations are averaged longitudinally and interpolated along the simulation trajectory and constrain the surface boundary of the LES (Figure 5.2). The lower boundary condition for atmospheric specific humidity was computed as the saturation humidity based on the surface skin temperature and the corresponding sea-ice fraction within the LES model.

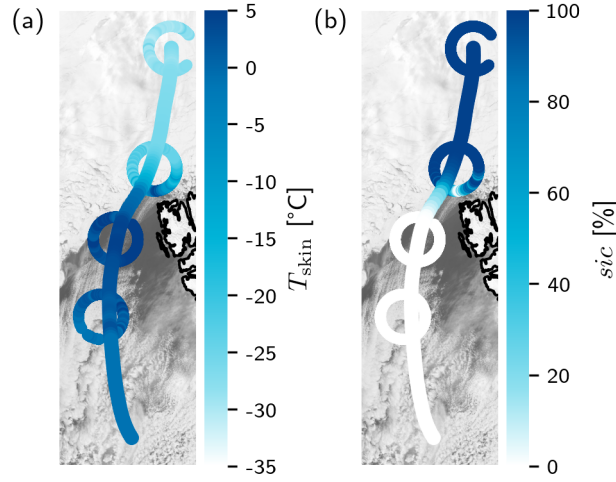


Figure 5.2: VELOX observations of surface skin temperature (T_{skin}) and sea-ice concentration sic on the circular flight legs of the HALO flights on 29th March 2022 and 30th March 2022 during HALO- $(AC)^3$ (circles), together with interpolated T_{skin} and sic along the LES trajectory used as urface boundary conditions (bold line).

The prescribed large-scale forcing profiles are time-dependent and they are obtained by interpolation along the trajectory between the sonde data at the four circle locations. A regression algorithm is used to estimate horizontal gradients within each circle area, following the method of [Bony and Stevens \(2019\)](#). The horizontal wind divergence and the associated subsidence ω are calculated as described by [Paulus et al. \(2024\)](#). At high latitudes, Cartesian latitude-longitude coordinates distort horizontal distances. To address this issue, [Paulus et al. \(2024\)](#) introduced a regression method in spherical coordinates $[r, \phi, \vartheta]$. Assuming linear fluctuations of any variable $\xi \in [u, v, q, T, p]$ within the dropsonde area, ξ at the centre of the pattern can then be approximated at every vertical level by a first-order Taylor expansion:

$$\xi = \xi_0 + \frac{\partial \xi}{\partial \vartheta} \Delta \vartheta + \frac{\partial \xi}{\partial \phi} \Delta \phi . \quad (5.1)$$

Here, ξ_0 denotes the mean of all dropsonde measurements, and ϑ and ϕ are the polar and azimuth angles, respectively. This approach is applied in this study to estimate divergence:

$$D = \vec{\nabla} \cdot \vec{v}_h = \frac{1}{r \cdot \sin \vartheta} \left[\frac{\partial u}{\partial \phi} + \frac{\partial \sin \vartheta v}{\partial \vartheta} \right]. \quad (5.2)$$

This allows for the estimation of the associated subsidence at height z by vertical integration from the surface:

$$\omega = -g \cdot \varrho \int_0^z D \, dz. \quad (5.3)$$

Figure 5.3 shows the resulting subsidence profiles in the four circles, provided in pressure and height units for reference, with key characteristics summarised in Table 5.2. In IC and Co2, subsidence dominates the lowest 3 km, with mean downward motions of -0.74 and -0.60 cm s^{-1} , respectively, while Co3 exhibits weaker subsidence throughout the profile. In contrast, Co1 shows a mean upward motion over the 8 km profile. Uncertainties were estimated following Paulus et al. (2024) and reach up to 0.025 cm s^{-1} (0.002 Pa s^{-1}) below 3 km across all circles, which cannot be resolved in the figure and are therefore not shown. The subsidence magnitudes are consistent with reanalysis-based estimates of Arctic MCAOs (Paulus et al., 2024; Tornow et al., 2023), but are approximately an order of magnitude lower than those reported for the tropical Atlantic, possibly due to the specific weather conditions of the sampled case (Bony and Stevens, 2019; George et al., 2021).

The same regression approach using spherical coordinates can be applied to calculate the horizontal advective tendencies in the circular areas,

$$-\vec{v}_h \cdot \vec{\nabla} \xi = - \left[u \frac{1}{r} \frac{\partial \xi}{\partial \vartheta} + \frac{1}{r \cdot \sin \vartheta} \frac{\partial \xi}{\partial \phi} \right], \quad (5.4)$$

and similarly, the horizontal pressure gradient force, here expressed as the geostrophic wind vector $\vec{v}_g$,

$$\vec{v}_g = \frac{1}{\varrho \cdot f} \cdot \left(\frac{\frac{\partial p}{\partial \phi}}{\frac{1}{\cos \phi} \frac{\partial p}{\partial \vartheta}} \right). \quad (5.5)$$

These profiles, which extend up to HALO's flight altitude at about 8 km, are linearly interpolated between the four circles and kept constant south of the last circle along the simulation trajectory (Figure 5.4). What stands out in the interpolated mesoscale vertical pressure velocity field $\omega(t, z)$ is the distinct maximum at Co2 below 4 km, which indicates a relatively

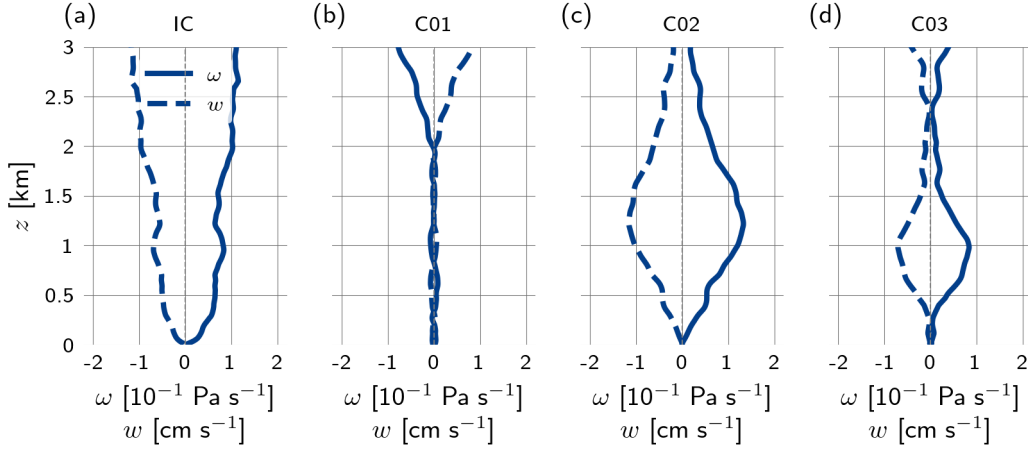


Figure 5.3: Subsidence profiles obtained through regression from a dropsonde pattern in pressure (ω , solid line) and height (w , dashed line) coordinate for all dropsonde patterns on 29th March 2022 and 30th March 2022 of the Airborne Research Campaign in the Arctic in March and April 2022 (HALO-($\mathcal{AC}$)³) campaign. Associated uncertainty estimates cannot be resolved in the figure and are discussed in the main text.

strong signal in the prescribed subsidence that will play a crucial role in this study. Concerning the advective tendencies, the temperature advection $-\vec{v}_h \cdot \vec{\nabla} T$ is dominated by cooling at low altitudes, below 1 km. The specific humidity exhibits low-level drying before the sea-ice edge, transitioning to moistening south of C02. The horizontal wind components show low signals, except for a slight positive tendency of the meridional wind in the deepening ABL south of C01. These likely reflect Ekman-layer effects of Reynolds stresses, inducing a deviation of the wind from the geostrophic wind direction. The geostrophic wind ($|\vec{v}_g|$) expresses an eastward zonal pressure gradient force throughout the lower levels, consistent with the general synoptic situation during the two days. It should be noted that at a trajectory height of 870 hPa corresponding to heights between 1.34 km and 1.38 km, advective tendencies are, by definition, zero, as the air mass speed is subtracted from the wind in the Lagrangian forcing method.

Above 8 km altitude, which is well above the simulated ABL height (max 2 km), the sonde data are extended by ERA5 fields. The main objective of this procedure is to allow radiative flux density (irradiance) calculations above the model domain in which turbulence is resolved, which has a ceiling far below HALO's flight altitude (described in Section 5.3.4). As a result, it is concluded that the use of ERA5 fields above this height has no discernible impact on the model results.

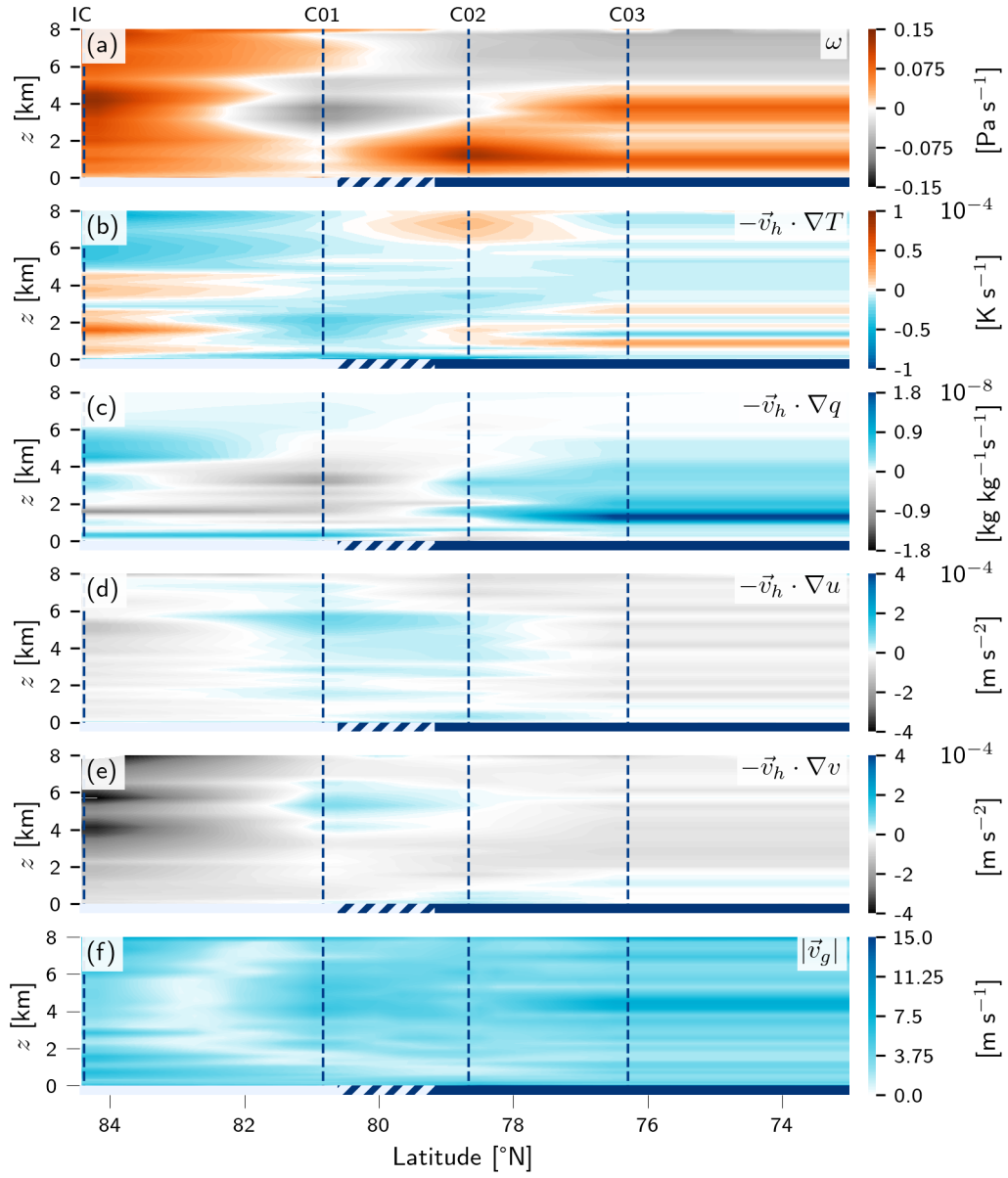


Figure 5.4: Simulation forcing of subsidence ω (Ctrl) (a), horizontal temperature advection $-\vec{v}_h \cdot \nabla T$ (b), horizontal humidity advection $-\vec{v}_h \cdot \nabla q$ (c), horizontal advection of the zonal and meridional wind components $-\vec{v}_h \cdot \nabla u$ (d), $-\vec{v}_h \cdot \nabla v$ (e), and the geostrophic wind magnitude $|\vec{v}_g|$ (f). All fields are interpolated between dropsonde observation locations IC, C01, C02 and C03 (dashed blue lines). The surface cover is indicated at the bottom of each subfigure as sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

5.3.4 EXPERIMENTAL SETUP

Apart from the control setup, four additional LES experiments are conducted using DALES to investigate the sensitivity of ABL evolution to the mesoscale subsidence forcing. The experimental setup includes (i) the control simulation (Ctrl) that employs the observed subsidence (Section 5.3.3), (ii) a simulation without subsidence (NoSub) to isolate the effects of entrainment on ABL development, and (iii) further experiments in which observed subsidence was scaled by 10 % (Sub10), 50 % (Sub50) and 200 % (Sub200). These scaling factors correspond to a range of subsidence rates covering weaker values (-0.12 cm s^{-1}) comparable to those used in LES studies of mixed-phase Arctic clouds (Solomon et al., 2018; Young et al., 2018), up to stronger rates of approximately -2.3 cm s^{-1} , and consistent with observations during the EUREC⁴A campaign in the tropics (Poujol and Bony, 2024). In all sensitivity experiments, the temporal and vertical structure of the observed subsidence was retained, while only its magnitude was modified.

The model domain spans $12.8 \text{ km} \times 12.8 \text{ km}$ with 50 m horizontal resolution and 191 vertical levels, starting at 10 m with 20 m spacing up to 2.1 km and increasing resolution aloft to a ceiling height of 6.8 km with maximum grid spacing of 177 m. Periodic horizontal boundaries are applied, and large-scale forcings are horizontally homogeneous but variable in time and height along the trajectory. The prescribed subsidence acts on the evolving model profile. Continuous Newtonian nudging towards the observed profiles of $\{T, q, u, v\}$ is applied above the ABL, determined by the strongest inversion in liquid water potential temperature, with a buffer layer of 100 m where the nudging time scale increases linearly up to 3 h. Nudging increases in the upper quarter of the domain to suppress gravity waves, while below the ABL, the flow evolved freely. Each experiment is 45 h long, reaching the MIZ approximately 15 h after initialisation.

5.4 RESULTS I: EVALUATION OF THE CONTROL SIMULATION

In this section, we evaluate the performance of the control (Ctrl) LES in reproducing key characteristics of the MCAO observed on 29th March 2022 and 30th March 2022 during the HALO-(AC)³ campaign. We analyse the evolution of cloud liquid and ice water contents and paths, the vertical structure and evolution of the ABL height, air temperature and humidity profiles along the simulation trajectory, and the associated turbulent heat energy

fluxes. As outlined in Section 5.2, observations from dropsondes, lidars, and radiometers are used to assess the model’s ability to capture the spatial and temporal variability of the atmosphere, and the representation of turbulent and thermodynamic processes.

5.4.1 ATMOSPHERIC BOUNDARY LAYER AND CLOUD EVOLUTION

First, we examine the evolution of the ABL and clouds, focusing on the temporal development of cloud liquid and ice water contents as well as the deepening of the ABL along the simulation trajectory of the MCAO, expressed as a function of latitude in Figure 5.5. The initial atmospheric column at circle IC includes a shallow ABL of 65 m, which remains low over the sea-ice in the simulation. The ABL begins to deepen once it reaches the MIZ where warmer surface conditions trigger enhanced vertical latent and sensible heat fluxes. Convective updrafts form a thin mixed-phase cloud between Co1 and Co2, with ice reaching to the surface and a shallow liquid layer existing near the cloud top during the initial phase of the growing ABL. As the air mass moves over the open ocean, the ABL continues to deepen, reaching values of liquid water content q_l up to 0.35 g kg^{-1} . At the end of the 45 h simulation, the ABL has grown to approximately 1.5 km.

To evaluate the Ctrl simulation with respect to the ABL structure, several independent measurements of ABL height are compared with the simulation output. The ABL height was determined from dropsonde profiles using the Bulk Richardson number Ri_b , which quantifies the balance between buoyancy and shear (Stull, 1988). A critical threshold $Ri_b^{\text{crit}} = 0.25$ was used to define the height of the ABL z_i , consistent with the diagnostics of the IFS Cy43r1 model and the ERA5 reanalysis (ECMWF, 2016; Seidel et al., 2012). Lidar measurements from the P5 aircraft (AMALi) provide cloud top height (z_c^t), which is comparable to the ABL height in MCAO cases due to the strong coupling of ABL growth to cloud development. Finally, the simulation diagnoses the height of the ABL from the minimum flux of total liquid water potential temperature (Stevens et al., 2001), $\overline{w'\theta'_l}$, within the lowest 3 km of the domain. AMALi measured a mean cloud top height of 0.63 ± 0.11 km at location Co2, yielding an RMSE of 0.12 km compared to the simulation ABL height. The ABL height derived from the dropsondes was 0.07 km at Co1, 0.43 km at Co2, and 1.04 km at Co3, resulting in an overall root mean square error (RMSE) of 0.14 km compared to the simulation. The conclusion from this evaluation against the three datasets is that the model reproduces the observed ABL height within 150 m throughout the simulation. The temporarily reduced

deepening rate between Co2 and Co3 in the simulation is supported by the measured data, which is an interesting feature that could be related to the low-level maximum in subsidence during that period (Figure 5.4a).

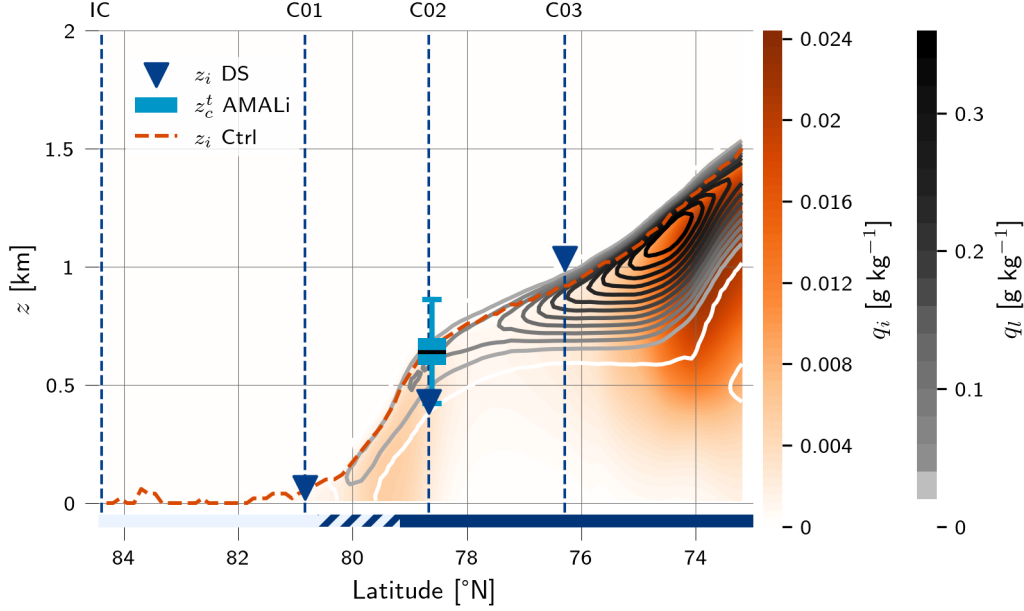


Figure 5.5: Latitudinal evolution of cloud liquid (grey contour lines) and ice (red filled contours) along the MCAO trajectory. Results from the 45 h semi-Lagrangian DALES simulation Ctrl are shown, including model ABL height (red dashed), compared to dropsonde ABL heights ($\blacktriangledown$) and cloud-top heights from AMALi lidar aboard the P5 aircraft (light blue boxplot, outliers not shown for better visibility). Dashed vertical lines indicate dropsonde circle locations.

5.4.2 TEMPERATURE AND HUMIDITY PROFILES

The vertical structure of virtual potential temperature θ_v and specific humidity q in the Ctrl experiment is evaluated against in-situ observations from HALO dropsondes, closest to the simulation trajectory; the variability of all soundings in each circle is indicated as blue shading in Figure 5.6. This comparison reveals possible systematic biases in the thermodynamic state. Figure 5.6 focuses on the lowest 3 km of the atmosphere to highlight the ABL properties. Above 3 km altitude, the simulation matches the observations of θ_v with an agreement of more than 99 % (Equation (5.14)) at all three locations, following a logarithmic profile. Profiles of q above 3 km are reproduced with an agreement of more than 67 % at Co1, Co2 and Co3, with the larger differences resulting from observed humidity layers not captured in the smoother simulated humidity structure.

In the region of Co1 (Figure 5.6 a and d), located above sea-ice near its edge, an inversion in θ_v is detected near the surface at 30 m in the simulation, and at 20 m in the dropsonde profiles, with a thermodynamic wind profile above it. The simulated near surface values of θ_v of -25.6°C agree within the standard deviation with the dropsonde observations with a measured value of -24.2°C . The general agreement of the temperature profiles below 3 km is 99.7 %. In contrast, the measured humidity profiles exhibit a larger spread among the sondes; however, the closest profile agrees well with the slightly smoother simulated profile, indicating good overall consistency. In contrast, the humidity profiles differ in their variability, as the dropsondes resolve several humid layers that are not captured by the modelled profile. The near-surface value of q at Co1 was 0.5 g kg^{-1} in the simulation and 0.54 g kg^{-1} in the observations, and the overall profiles agree by 81 % below 3 km. The good agreement at Co1 reflects the continuous nudging that is active above the ABL top, which is still very low at this stage.

As the air mass moved over the MIZ and the open ocean to the location of Co2 (Figure 5.6 b and e), surface warming caused an increase in the near surface air temperature $\theta_{v,0}$ in the simulation of $+14.5 \text{ K}$ to -11.1°C , while observed warming was even stronger with a measured increase in $\theta_{v,0}$ of $+15.6 \text{ K}$ to -8.6°C . The inversion height was approximately 590 m in both profiles, but the simulated inversion was less steep. The profile below the inversion shows a cold deviation of 3.3 K, but the shape of the curve matches the measured profile, corresponding to an agreement of 98 % below the thermal inversion. The increase in evaporation over the open ocean led to an increase in humidity with surface moisture q_0 of 1.5 g kg^{-1} in the simulation and 1.6 g kg^{-1} in the observations. Both show a pronounced drop in humidity near 590 m, with the observed jump being stronger and sharper than the simulation, induced by the prescribed advection of q_t (shown in Figure 5.4c). Within the mixed layer, where temperature and humidity are uniform with height, the model underestimates the humidity by an average of 0.2 g kg^{-1} with an agreement of 88 % below the temperature inversion. Due to observed moisture layers above the temperature inversion, which did not appear in the simulation, the agreement in the overall profile of q is 82 % below 3 km.

Further south, over the open ocean at Co3 (Figure 5.6 c and f), the ABL has deepened further and θ_v has continued to increase, with a simulated surface value of $\theta_{v,0}$ of -7.5°C and an observed value of -4.6°C representing an underestimation of near-surface warming by 2.9 K. The modelled inversion base is at 830 m, compared to 1000 m in the observations, with the

simulated inversion being less steep. Below the temperature inversion, the model underestimates θ_v by approximately 2 K although the agreement of the entire shown profile is still 99.6 %. Therefore, there is a consistent cold bias of -2 K to -3 K across all circles, a tendency also found by (Wendisch et al., 2025) in the limited-area configuration of the ICON model for all observed MCAO cases of HALO-(AC)³. The humidity in the boundary layer at Co3 has also increased further, reaching 1.7 g kg^{-1} in the simulation and 2 g kg^{-1} in the observations, with the model underestimating the surface layer of strong humidity directly above the ocean. Also, in humidity, the simulated jump is weaker. Below the temperature inversion, the model underestimates q by 0.2 g kg^{-1} , which corresponds to an agreement of 90 % below, while above the inversion, the absence of observed moisture layers reduces the agreement to 33 %. A consistent moisture deviation of -0.15 g kg^{-1} to -0.22 g kg^{-1} was found in all three profiles.

Additional measurements taken with the CMET sonde drifting south along a similar trajectory further support these results. At the same latitude as Co1, the CMET flew at an altitude of 340 m at a distance of 225 km to the simulated trajectory. The CMET reached the latitude of interest about 2.5 h later than the simulated air parcel. Here, $\theta_v = -8.8^\circ\text{C}$ and $q = 0.52 \text{ g kg}^{-1}$ were measured, which largely corresponds to the simulation in terms of humidity (difference of only 0.02 g kg^{-1}), but with θ_v being 5.7 K warmer, which may be due to the proximity to Svalbard and the associated local warming effects. As the trajectory of CMET was closer to Svalbard and travelled through the cloud-free region associated with lee-effects, likely causing a warming of the lower layers. The CMET reached the latitude of Co2 7.5 h later at a distance of 132 km from the simulated trajectory. The CMET measured at 902 m above sea level a temperature of $\theta_v = -1.8^\circ\text{C}$ and $q = 0.25 \text{ g kg}^{-1}$, which is 3.7 K warmer and 0.2 g kg^{-1} drier than the simulation. At Co3, the CMET was 2 h later and 149 km away from the simulation at an altitude of 1780 m. A virtual potential temperature θ_v of 2.7°C was measured, 1.8 K warmer than in the model and $q = 0.78 \text{ g kg}^{-1}$, only 0.06 g kg^{-1} more humid. As the CMET moved south and the influence of Svalbard diminished, the measured air mass increasingly resembled that observed by the dropsondes, with the humidity in all three circles corresponding very closely to the simulated q values measured by the dropsondes.

The overall conclusion from these results is that the LES model reproduces the observed evolution in the vertical thermodynamic structure after Co1 to a reasonable degree. This is not

trivial, as the nudging takes place above the temperature inversion only, and below it, the simulated ABL is free to evolve. A slight cold and dry bias is evident within the ABL, and the inversions at its top height appear less pronounced than observed. These differences may arise from (i) uncertainties in the surface boundary condition T_{skin} , and (ii) shortcomings in the bulk flux parametrisation at the surface. Nevertheless, the overall correspondence between simulation and observation indicates that the model adequately reproduces the thermodynamic structure of the observed MCAO, providing a robust basis for further analysis.

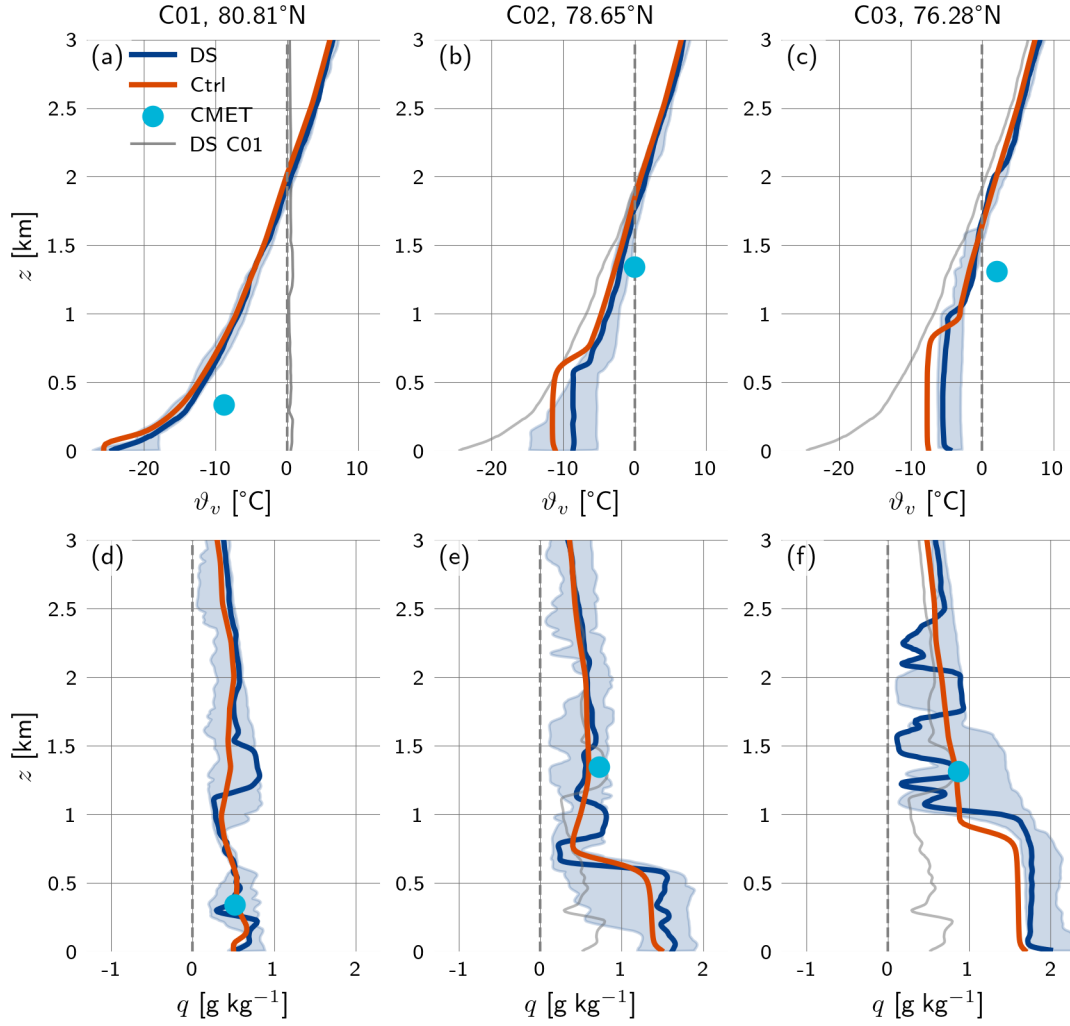


Figure 5.6: Vertical profiles of virtual potential temperature (θ_v) and specific humidity (q). Dropsonde profile (blue) of the dropsonde closest to the simulation trajectory and spread between values measured by 10 sondes per circle (shading); simulation profiles are shown in red. CMET measurements are indicated as blue dots.

5.4.3 INTEGRATED WATER PATHS

The model performance in representing various column-integrated quantities is assessed next. In ?? a, the simulated integrated water vapour (*IWV*) is evaluated against observations from two sources. The first source are the data from the HAMP radiometer on board HALO which measured along the flight trajectory, and the second is the *IWV* determined from the dropsonde humidity profiles, averaged over the ten sondes in each circular pattern. The simulation accurately replicated the observed *IWV* at Co1 of 1.9 kg m^{-2} with a small difference of 0.1 kg m^{-2} . At Co2, the dropsondes measured 2.7 kg m^{-2} *IWV*, reproduced with $+0.04 \text{ kg m}^{-2}$ in the simulation. At Co3, the observed *IWV* was 4.4 kg m^{-2} , which the simulation was slightly overestimated by $+0.3 \text{ kg m}^{-2}$. This temporal evolution of *IWV* closely follows the HAMP observations, with an RMSE of 0.2 kg m^{-2} compared to the simulation. These results suggest that the net input of humidity into the deepening ABL is well captured by the model.

Figure 5.7 b compares the simulated liquid water path *LWP* against two measurements. The *LWP* retrieved from HAMP onboard HALO exhibits a much higher short-term variability than in the simulation, resulting in an RMSE of 30 g m^{-2} . Despite this, the overall temporal evolution aligns with the simulated *LWP*, with liquid water appearing as soon as the MIZ is reached. The MiRAC instrument on P5 showed even higher variability due to its high spatial resolution and low flight altitude above the clouds. When averaged, MiRAC observations yield a mean *LWP* of 0.01 g m^{-2} , which appears much lower than the simulated *LWP* of 12 g m^{-2} , but the simulation value lies within the spread of the *LWP* observed by MiRAC. These results suggest that, while the observed *LWP* fluctuates more strongly due to higher resolution, capturing individual cloud streets, than the smoother, domain-average model output, the simulation captures the general trend and magnitude of the evolution of liquid water throughout the event.

The simulated ice water path of all frozen hydrometeors *IWP* shown in Figure 5.7 c is only compared to MiRAC measurements aboard P5. The *IWP* retrievals by HALO were excluded due to large uncertainties for this case. The average observed *IWP* from MiRAC was 0.6 g m^{-2} , but similar to the *LWP* observations, the variability due to the high resolution of individual clouds is large and values range up to 6.2 g m^{-2} . This high variability may also result from the substantial uncertainties associated with *IWP* retrievals. The *Ze-IWC*

relationship of (Hogan et al., 2006) exhibits uncertainties around a factor of 2, caused by the strong temperature dependence of ice particle size distributions and densities. At warmer temperatures, the ice water content IWC tends to be overestimated, whereas at colder temperatures it is generally underestimated. This temperature sensitivity is particularly relevant for 94 GHz radar, where Mie-scattering further increases the retrieval uncertainty. Compared to the simulated IWP in the location of the P5 measurements of 2.58 g m^{-2} , an RMSE of 73.34 g m^{-2} is found. It should also be noted that the P5 observations do not coincide with the air mass trajectory but were collected at a later time, limiting the comparability of the data (Figure 5.1). A notable feature of the latitudinal evolution of the simulated IWP is the local maximum reached in the simulation just before Co2; the realism cannot be verified due to the absence of reliable IWP measurements along the whole trajectory.

5.4.4 SURFACE HEAT FLUXES

The surface heat fluxes are key drivers of ABL evolution in MCAOs. To evaluate the simulated sensible (H) and latent (E) heat fluxes, we estimate these data from dropsonde observations following the thermodynamic method suggested by Hartmann et al. (1997). This method assumes steady-state conditions, negligible radiative divergence and cross-flow advection, and minimal entrainment above the mixed layer. Then, the sensible and latent heat energy fluxes can be calculated as

$$H_{\text{DS}} = \varrho \cdot c_p \cdot u_{\text{ML}} \int_{z_i}^0 \frac{\Delta\theta}{-\Delta y} dz - H_C, \quad (5.6)$$

$$E_{\text{DS}} = \varrho \cdot L_v \cdot u_{\text{ML}} \int_{z_i}^0 \frac{\Delta q}{-\Delta y} dz + H_C, \quad (5.7)$$

where ϱ represents the air density, c_p the specific heat capacity, L_v the latent heat of evaporation, u_{ML} the mean horizontal wind speed within the mixed layer and z_i the height of the ABL. Δy is the horizontal distance between two profiles. The contribution of condensational heating within the cloud layer is represented by the following formulas:

$$H_C = \varrho \cdot L_v \cdot u_{\text{ML}} \int_{z_i}^{z_c^b} \frac{\Delta l}{-\Delta y} dz, \quad (5.8)$$

where z_b represents the height of the cloud base, defined as the lowest altitude at which the relative humidity reaches saturation with respect to the ice or liquid. Meanwhile, Δl refers to the difference in q_l between the profiles, which is approximated as the upper limit of q_l within the cloud layer. This is estimated as the difference of q_t within the cloud and the average $\overline{q_{t,BC}}$ below the cloud layer, following (Hartmann et al., 1997). In the model, the surface heat fluxes can be sampled continuously as

$$H_{\text{Ctrl}} = \rho \cdot c_p \cdot \overline{w' \cdot \theta'}, \quad (5.9)$$

$$E_{\text{Ctrl}} = \rho \cdot L_v \cdot \overline{w' \cdot q'}, \quad (5.10)$$

with $\overline{w'\theta'}$ and $\overline{w'q'}$ representing the turbulent kinematic fluxes of temperature and moisture at the surface.

The simulation produced H_{Ctrl} values up to 830 W m^{-2} and E_{Ctrl} values up to 102 W m^{-2} over the open ocean. In particular, the sensible heat flux is substantially higher than the climatological mean sensible flux of about 200 W m^{-2} during MCAOs, while the latent heat flux is comparable to the climatological mean latent heat flux of 100 W m^{-2} reported by Papritz and Spengler (2017) for MCAOs in the region west of Svalbard. The difference in potential temperature between the near-surface atmosphere and the ocean surface causes a direct response in the sensible heat flux, while the latent heat flux is limited by the saturation specific humidity at the sea surface temperature (Papritz et al., 2015; Papritz and Spengler, 2017). In the observed region, H was generally more efficient than E in extracting energy from the ocean surface, but the relative importance of E increased as the air is mass advected away from the sea-ice edge over warmer waters. South of 75°N , the mean values of H_{Ctrl} and E_{Ctrl} were 69 W m^{-2} and 17 W m^{-2} , respectively, consistent with the findings of Brümmer (1997).

Dropsonde-derived H_{DS} values considerably underestimate simulated surface sensible heat fluxes by 564 W m^{-2} between Co1 and Co2 and 306 W m^{-2} between Co2 and Co3. We speculate that this offset is mostly due to numerous assumptions in the method of deriving them from observed profiles. To improve comparability between model and measurements and gain insight, the same diagnostic technique is now applied to the simulated profiles (Figure 5.6), which yields much better agreement. The associated RMSE is 95 W m^{-2} between H_{DS} and $H_{\text{Ctrl, Prof}}$ due to a large underestimation of $H_{\text{Ctrl, Prof}}$ between Co2 and Co3 of only 21 W m^{-2} , while $\text{DS} = 170 \text{ W m}^{-2}$. Note that the dropsonde-derived DS in the marginal

sea-ice zone (CoI-Co2) is very small (2.6 W m^{-2}) compared to the simulated fluxes that exceeded 100 W m^{-2} , but increased to 16 W m^{-2} between Co2 and Co3 , closer to the simulated values. The RMSE of 8.9 W m^{-2} between E_{DS} and $E_{\text{Ctrl, Prof}}$ indicates that although the observational profile-based method struggles to fully capture the true flux magnitudes in the early MCAO phase, it reproduces their relative variability well when applied consistently.

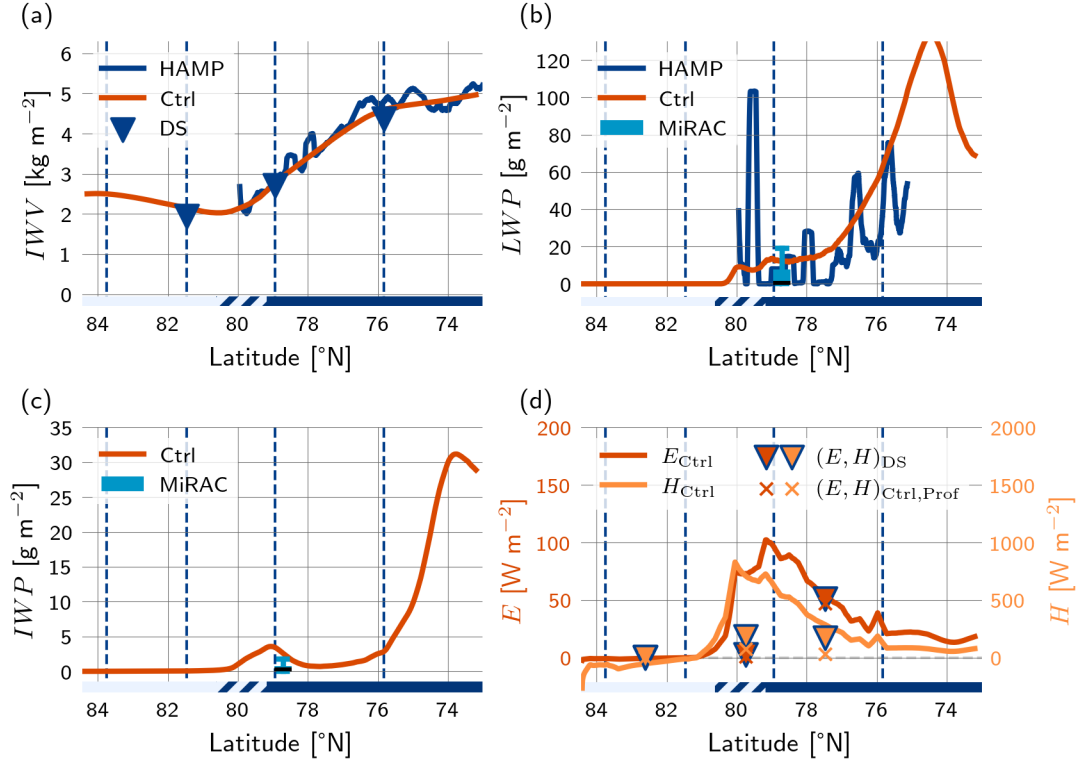


Figure 5.7: Latitudinal evolution of various bulk properties of the simulated MCAO in the Ctrl experiment. Integrated water vapour IWV simulation (red line) compared to longitudinally averaged observations from HALO (blue line) and estimates based on dropsonde data ($\blacktriangledown$; a). Simulated liquid water path LWP (b) and ice water path IWP (red lines) compared to HALO (dark blue line) and P5 (light blue boxplot, outliers not shown for better visibility) observations (c). Simulated surface latent (E , red line) and sensible (H , yellow line) heat fluxes compared to estimates based on dropsonde data following the method of [Hartmann et al. \(1997\)](#) ($\blacktriangledown$; d). The same method applied to model profiles is also shown, for reference ($\times$). In all figures, the vertical dashed lines mark the location of the dropsonde circles. The coloured bar at the bottom indicates sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

5.5 RESULTS II: SUBSIDENCE IMPACTS

The results discussed so far indicate that the Ctrl simulation reproduces the temporal evolution of the ABL well, including various water paths and surface energy fluxes. Therefore, the Ctrl experiment represents a robust and realistic foundation for subsequent Lagrangian LES sensitivity experiments, as applied in this section to investigate the impact of mesoscale subsidence on the ABL and cloud evolution. In total, five experiments with differing mesoscale subsidence are performed and analysed. As described in Section 5.3.4, four additional experiments were conducted with the Ctrl subsidence field multiplied by a factor $\{0, 0.1, 0.5, 2.0\}$, respectively.

5.5.1 SENSITIVITY ANALYSIS

Figure 5.8 shows the evolution of q_l and q_i along the MCAO trajectory. In the no-subsidence (NoSub) setup, the cloudy ABL deepens the fastest, approaching a parabolic deepening rate ($\sim t^{1/2}$) after the air mass moves over the open ocean (Stevens, 2007). In this setup, the cloud mass is the largest, both in liquid and frozen form. The higher the prescribed subsidence forcing, the more the ABL deepening driven by turbulent/convective entrainment is suppressed by subsidence. This in particular applies during the time period between Co2 and Co3, where the observed low-level subsidence has its maximum (Figure 5.4a). This leads to the development of thinner clouds and less q_l and q_i . In the Sub200 case, only a very thin, shallow mixed-phase cloud is observed.

To gain more insight into the spatial structure of the cloud field, Figure 5.8 presents horizontal cross-sections of q_l and q_i through the centre of the simulated domain for all five simulations. These locations are sampled at regular intervals along the trajectory, beginning at Co1 near the sea-ice edge and extending southward to 73.25 °N at the end of the simulation period. The grey shading highlights regions in which the cloud layer is decoupled from the surface, following the criterion of Jones et al. (2011). This diagnostic is defined by the thresholds $\Delta q_t = \overline{q_{t,\text{bot}}} - \overline{q_{t,\text{top}}} > 0.5 \text{ g kg}^{-1}$, based on the layer-mean total-water mixing ratio in the upper and lower 25 % of the boundary layer, and $\Delta \theta_l = \overline{\theta_{l,\text{top}}} - \overline{\theta_{l,\text{bot}}} > 0.5 \text{ K}$, calculated analogously for the layer-mean liquid-water potential temperature. These metrics have been associated with the onset of boundary layer decoupling and the subsequent transition from roll convection to open-cell structures in MCAOs (Karalis et al., 2022; Abel et al., 2017).

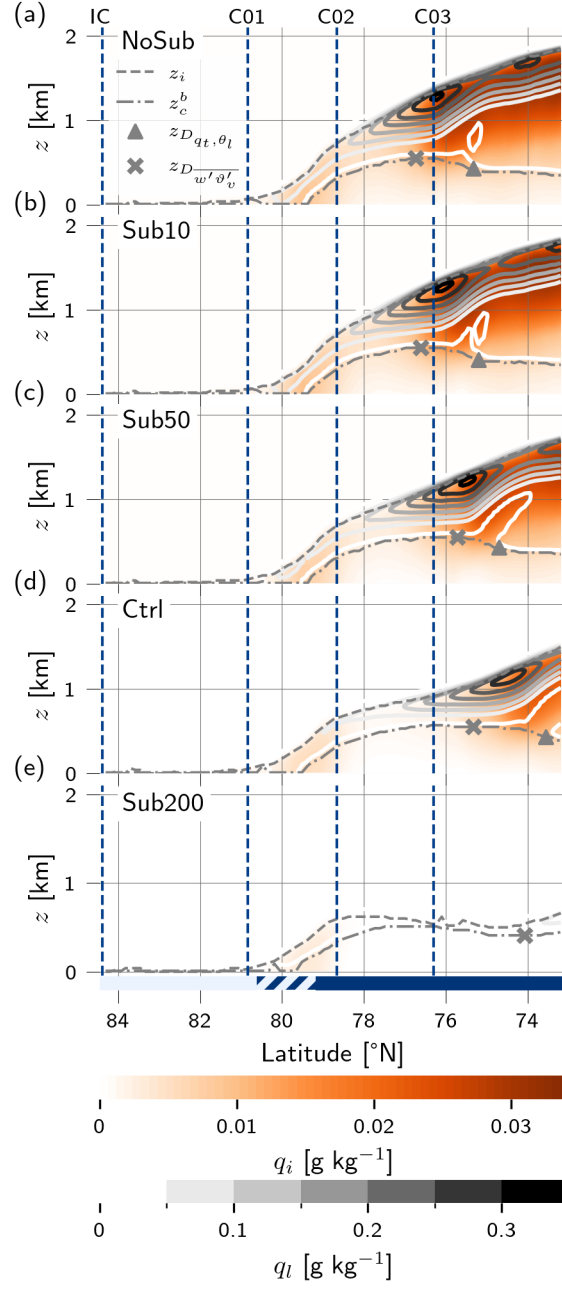


Figure 5.8: Latitudinal evolution of cloud liquid (grey) and ice (red) along the MCAO trajectory for all sensitivity experiments. Dashed horizontal lines indicate dropsonde circle locations. Boundary layer height z_i and cloud base z_c^b are indicated by grey dashed and dot-dashed lines, respectively. The markers $\times$ and $\blacktriangledown$ indicate decoupling points and heights with respect to $\overline{w'\theta_v'}$ and Δq_t and $\Delta \theta_l$ respectively (Section 5.5.2 for definition). The coloured bar at the bottom indicates sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

What stands out during the decoupled phase is the presence of liquid-phase shallow cumulus clouds, rising into the capping mixed-phase cloud layer.

Across the simulations, the onset of decoupling exhibits a systematic dependence on the prescribed subsidence: in cases with reduced subsidence, the decoupling occurs closer to the sea-ice edge and is accompanied by overall larger maxima in both liquid and ice water content. While all simulations show the initial formation of a thin ice cloud at Co_1 , substantial differences arise beginning at Co_1 , most notably in cloud-top height and the vertical distribution of liquid and ice. The Sub200 simulation produces only a shallow ice cloud capped by a thin liquid layer, with little variation in cloud-top altitude throughout the trajectory. In contrast, the simulations with weaker subsidence develop considerably deeper clouds, characterised by more complex internal structure and enhanced condensate mass.

The impact of subsidence on the developing ABL is characterised by various bulk quantities, as shown in Figure 5.10. The integrated water vapour IWV (Figure 5.10 a) is similar in all simulations, but the boundary layer (Figure 5.10 d) in the NoSub case reaches substantially greater depths (2 km), following a parabolic deepening rate due to unimpeded entrainment at the ABL top. All simulations exhibit a similar strong initial boundary layer growth within the MIZ and immediately downstream of the sea-ice edge (up to Co_2), driven primarily by intense surface heat fluxes in this region (Figure 5.7). However, after this initial phase, the Sub200 simulation boundary layer actually becomes shallower, with the strong subsidence keeping z_i below 1 km. In contrast, the Ctrl, Sub50, and Sub10 simulations all maintain continued ABL deepening, though z_i remains below the parabolic limit visible in NoSub.

The liquid water path LWP (Figure 5.10 c) exhibits an initial peak near the sea-ice in all simulations, coinciding with the region of strong boundary layer growth at approximately $80^\circ N$ on average. The RMSE in the location of the peak between the simulations is $0.01^\circ N$, and the strength of the peak decreases with increasing subsidence by 18 g m^{-2} , with an overall RMSE of 4.9 g m^{-2} . A second, more pronounced LWP maximum occurs later at $76.3^\circ N$ in NoSub, with a maximum value of 173 g m^{-2} . For Sub50 and Ctrl, this secondary peak shifts to lower latitudes and reduced magnitudes, occurring at $76^\circ N$ (167 g m^{-2}) and $74.4^\circ N$ (133 g m^{-2}), respectively. In Sub200, the liquid water path does not recover after the initial maximum and shows only a gradual increase later in the simulation, beyond $74^\circ N$.

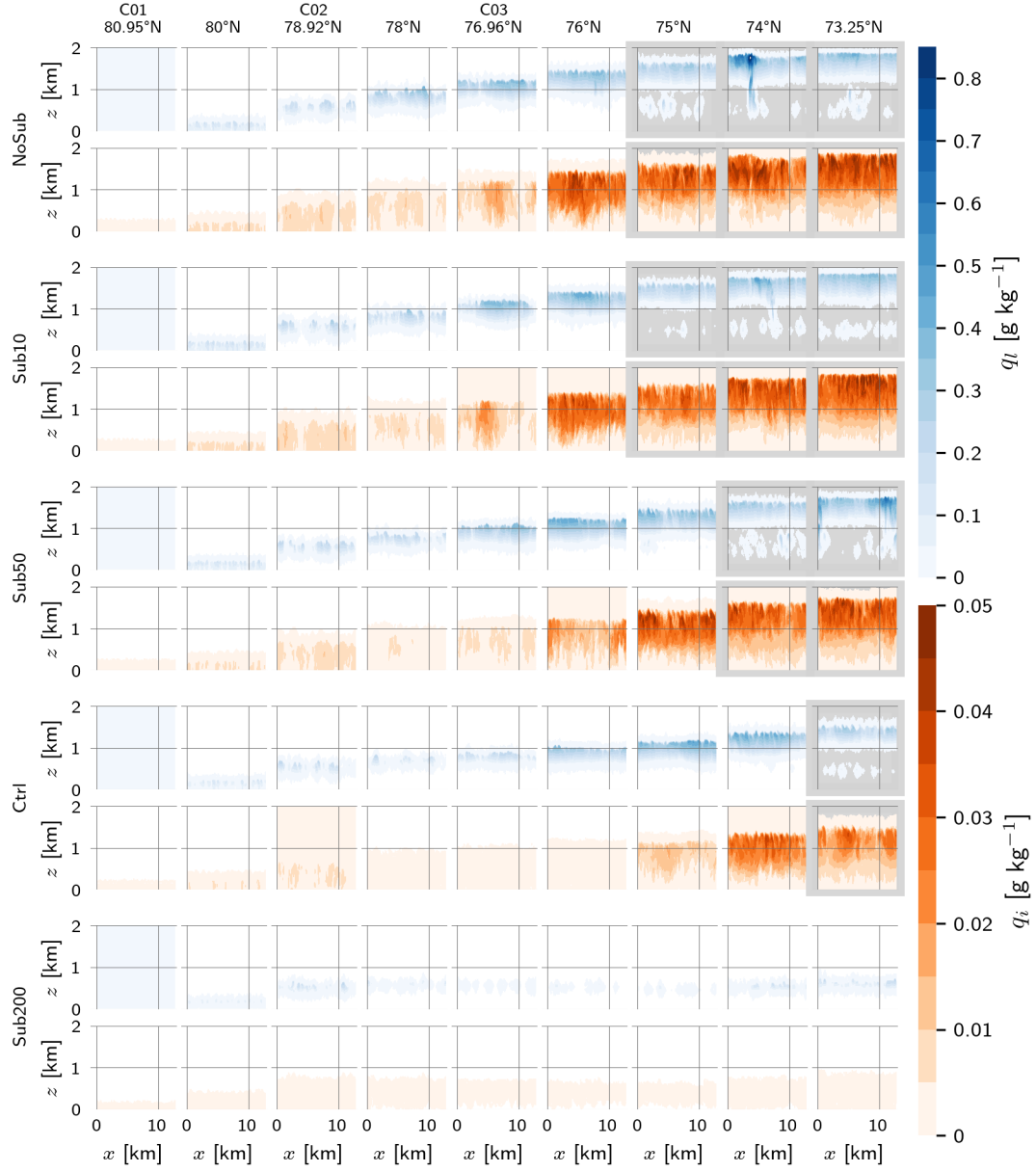


Figure 5.9: Horizontal cross-sections of cloud liquid water (q_l) and cloud ice (q_i) at the domain centre ($y = 6.4$ km), at different locations along the trajectory and for different subsidence experiments. Cross-sections of decoupled clouds ($\Delta q_t > 0.5 \text{ g kg}^{-1}$ and $\Delta \theta_l > 0.5 \text{ K}$) are highlighted in grey.

The ice water path IWP (Figure 5.10 e), including all frozen hydrometeors, displays a similar spatial evolution, with an initial peak forming in the MIZ around 79.4°N , consistent in all simulations. Subsequently, IWP increases over the open ocean following the sea-ice edge. Compared to LWP the ice growth begins slightly earlier (also visible in Figure 5.9), which is consistent with previous studies of Arctic MCAOs (Murray-Watson and Gryspeerdt, 2024; Inoue et al., 2021). The peak IWP in NoSub occurs at 75.7°N with 48 g m^{-2} , whereas for Sub50 and Ctrl it shifts to 75.7°N and 73.8°N respectively with reduced magnitudes of 41 g m^{-2} and 31 g m^{-2} respectively. The Sub200 simulation exhibits no such shift or comparable peak values, indicating a systematic suppression of ice growth under stronger subsidence.

The ratio of ice to liquid water, $\frac{IWP}{LWP}$, calculated only for regions with a significant liquid content ($LWP > 1\text{ g m}^{-2}$), highlights these differences more clearly (Figure 5.10 b). The initial shallow cloud developing above the MIZ shows a similar structure among NoSub, Sub10, Sub50, and Ctrl, while Sub200 displays a slightly larger ice fraction in the initial cloud. However, because of the overall thinness of both liquid and ice layers, this difference cannot be regarded statistically significant. In later stages, LWP in Sub200 falls below 1 g m^{-2} , producing a discontinuity in the $\frac{IWP}{LWP}$ ratio, while in the remaining cases a clear shift toward later growth of IWP is evident.

Finally, the impact of subsidence on the total surface precipitation rate (P_{tot}) (Figure 5.10 f) is broadly similar to the IWP . The NoSub simulation reaches precipitation rates up to 60 mm h^{-1} toward the end of the simulation, compared with 25 mm h^{-1} at the peak near 74°N for Ctrl. In Sub200, precipitation remains confined to the initial shallow cloud phase, with maximum rates of only 2.1 mm h^{-1} .

5.5.2 GLACIATION AND DECOUPLING

The results so far indicate that the behaviour of the cloud ice phase is particularly affected by the strength of subsidence. To gain further insight, the relative contributions by the three frozen hydrometeor classes in the LES microphysical scheme are investigated. The integrated column water paths of cloud ice ($CIWP$), snow ($SIWP$), and graupel ($GIWP$) are shown in Figure 5.11. The behaviour of $CIWP$ (Figure 5.11 a) closely follows that of the total ice water path (IWP), reflecting its dominant contribution to the total ice mass. The maximum values of $CIWP$ reach 38 g m^{-2} at 73.4°N for NoSub and 32 g m^{-2} at 73.7°N for Ctrl, with

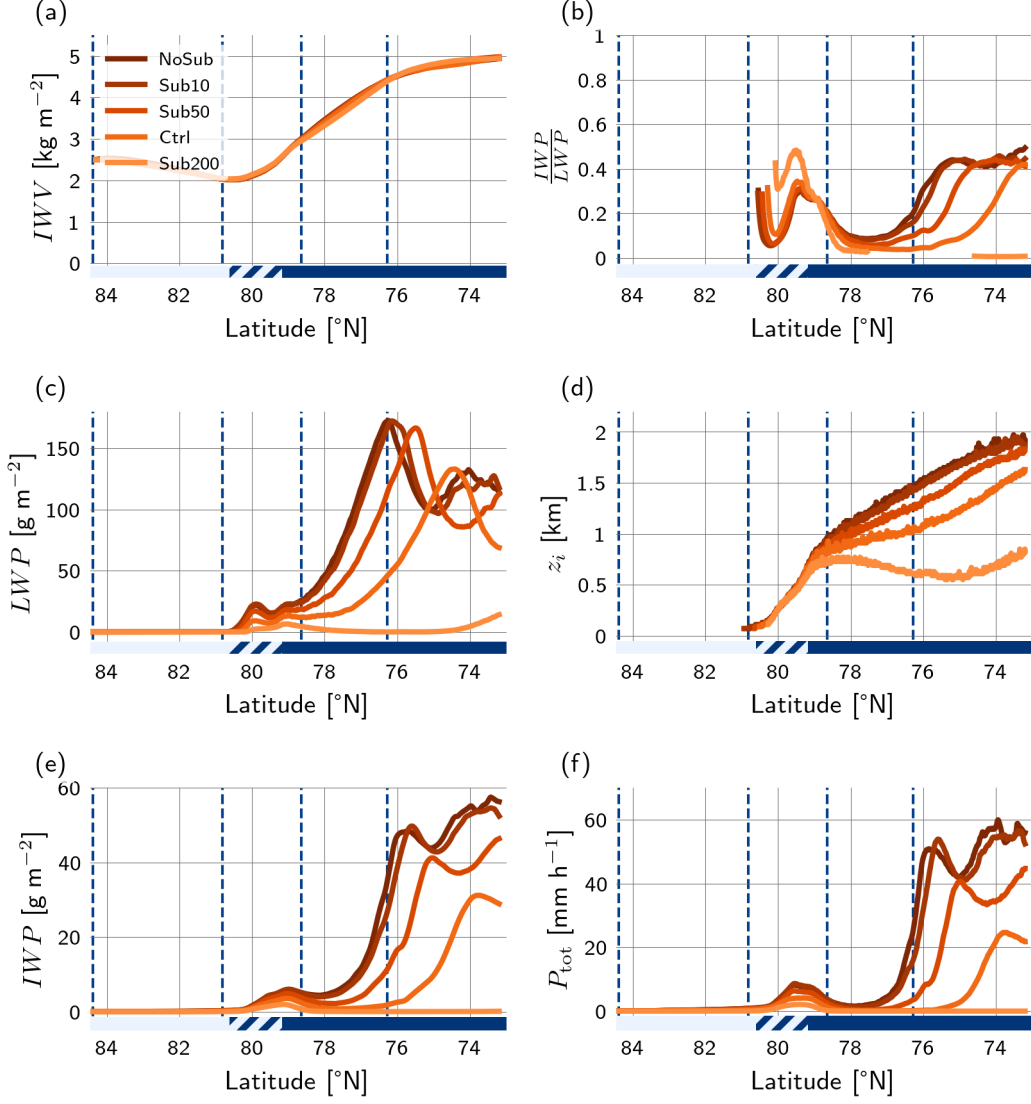


Figure 5.10: Integrated water vapour IWP (a), liquid water path LWP (c), ice water path IWP (e), including all frozen hydrometeors, together with its ratio $\frac{IWP}{LWP}$ (b), the boundary layer height z_i (d) and the total precipitation rate P_{tot} (f) for Lagrangian LES with different subsidence forcings. The locations of the dropsonde circles are shown as vertical dashed lines, and the surface cover is indicated as sea-ice ($s_{ic} > 0.9$, light blue), marginal sea-ice zone ($0 < s_{ic} \leq 0.9$, hatched), and open ocean ($s_{ic} = 0$, dark blue).

a systematic decrease in amplitude under stronger subsidence. The *SIWP* (Figure 5.11 c) distribution exhibits a similar latitudinal progression, with peak values occurring later and at reduced magnitudes as subsidence increases. However, unlike *CIWP*, *SIWP* does not sharply decline after the peak but plateaus afterwards, reaching 9.6 g m^{-2} at 73.3°N for No-Sub and 5.5 g m^{-2} at 73.4°N for Ctrl. The graupel path *GIWP* (Figure 5.11 b) exhibits the most distinct and localised peak. In NoSub the peak occurs at 76°N with a magnitude of 8.4 g m^{-2} , generally shifting southward with increasing subsidence to 74°N for Ctrl, weakening in amplitude to 2.7 g m^{-2} . The Sub200 experiment does not develop graupel. In this southward progression, the width of the graupel peak remains similar.

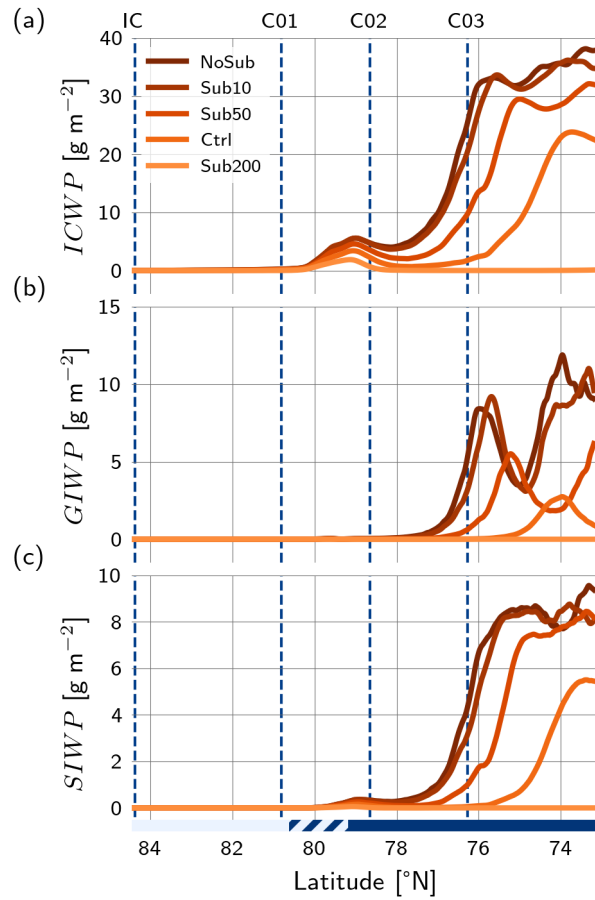


Figure 5.11: Integrated cloud ice water path *CIWP* (a), integrated snow water path *SIWP* (c) and integrated graupel water path *GIWP* (b). The locations of the dropsonde circles are shown as vertical dashed lines, and the surface cover is indicated as sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

Graupel is typically formed in convective updrafts, as it requires substantial vertical velocities. Accordingly, the southward shift and progressive weakening of the peak in $GIWP$ with increasing subsidence probably reflects a similar shift in convective dynamics. This motivates a deeper investigation of the behaviour of boundary layer internal decoupling. Previous MCAOs studies have established that such decoupling plays a central role in the transition from organised roll convection to open-cellular convection. Enhanced precipitation, often driven by secondary ice production and riming, has been shown to promote evaporation and sublimation below the cloud, thereby cooling and moistening the sub-cloud layer. This pre-conditioning in water vapour at lower levels accelerates the breakup of stratocumulus decks and favours the onset of cellular convection. As a result, the surface supply of moisture and heat to the cloud layer is reduced, weakening roll structures and promoting the development of cellular patterns (Abel et al., 2017; Tornow et al., 2021; Karalis et al., 2022; Wu et al., 2025). A slightly different form of decoupling that does not involve precipitation has been proposed earlier by Bretherton and Wyant (1997), showing that during deepening marine subtropical stratocumulus-topped ABLs a layer of negative buoyancy flux $\overline{w'\theta_v'}$ can form below cloud base, as a result of top-down and bottom-up driven mixing becoming too far separated in the vertical. This negative buoyancy flux layer then disrupts the ABL-deep circulation, inducing decoupling.

Both mechanisms of decoupling and the link to graupel formation are investigated next. The buoyancy-based decoupling mechanism is assessed by defining a decoupling point $D_{\overline{w'\theta_v'}}$ as the time point at which $\overline{w'\theta_v'}$ becomes negative below cloud base for the first time, and remains so until the end of the simulation (indicated as ▼ in Figure 5.8). The humidity-based decoupling mechanism involving precipitation evaporation is assessed by defining a second decoupling point, this time inferred from Δq_t and $\Delta \theta_l$ denoted D_{q_t, θ_l} , as discussed above (Section 5.5.1). Comparing these two definitions gives insight into how exactly decoupling works in MCAOs and how it affects the shifts in ice-phase hydrometeors.

Figure 5.12 compares the locations of the largest peaks in liquid water path ($LWP_{\max}$), graupel water path ($GIWP_{\max}$), and surface precipitation rate ($P_{\text{tot}, \max}$) with the two decoupling points D_{q_t, θ_l} and $D_{\overline{w'\theta_v'}}$, for all simulations except Sub200 (in which no graupel appears). A key result that stands out is that buoyancy-based decoupling (vertical axis) structurally happens first, in all simulations (also indicated in Figure 5.8). This is followed by the peak in LWP , then integrated graupel, then surface precipitation, and finally the precipitation-

based decoupling point. This sequence of events typically covers 1 degree of latitude in this marine cold air outbreak (MCAO) case, and shifts southward with increasing prescribed mesoscale subsidence. The order of appearance of the various maxima is consistent with the idea that i) internal buoyancy-decoupling initiates events and causes stronger surface-driven convection, ii) driving stronger precipitation formation (mainly in graupel form) that efficiently depletes cloud liquid water mass (as discussed by [Abel et al., 2017](#)), iii) forming intense graupel for a while, iv) associated with enhanced surface precipitation which then v) further stabilises the lower boundary layer through evaporation below cloud base, suppressing vertical mixing and further enhancing decoupling ([Wu et al., 2025](#)). The last step is also suggested by the slight lowering of liquid cloud base height visible in Figure 5.8. This sequence of events ultimately yields a fully decoupled structure with respect to Δq_t and $\Delta \theta_l$ at D_{q_t, θ_l} .

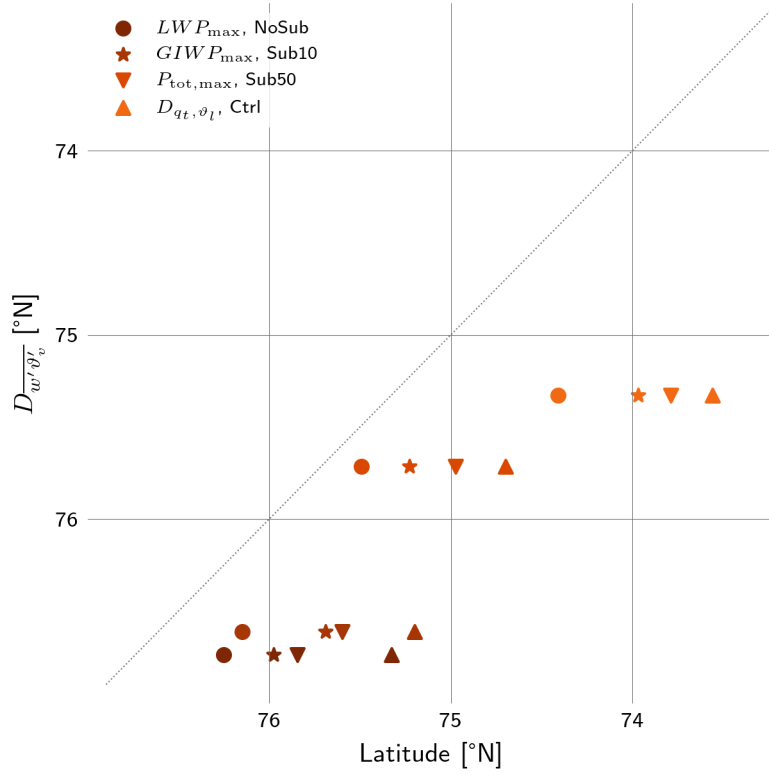


Figure 5.12: Location of the largest peak in liquid water path $LWP_{\max}$, graupel water path $GIWP_{\max}$ and precipitation $P_{\text{tot}, \max}$ and the location of decoupling with respect to the q_t and θ_l profiles D_{q_t, θ_l} (following [Jones et al. \(2011\)](#)) in relation to the location of decoupling with respect to the buoyancy flux $D_{w' \theta_v'}$ (following [Bretherton and Wyant \(1997\)](#)). The colours represent the different simulation experiments as in Figure 5.11. The grey dashed line indicates the identity.

5.6 CONCLUSIONS

This study investigates the sensitivity of Arctic mixed-phase cloud evolution in developing marine cold air outbreaks with respect to mesoscale subsidence, using a Lagrangian large eddy simulation framework constrained by HALO-($\mathcal{AC}$)³ observations. The control simulation was evaluated against an extensive and independent set of airborne in-situ and remote-sensing and meteorological balloon measurements. In the control run, the model reproduced the vertical thermodynamic structure of the observed marine cold air outbreak reasonably well, capturing the evolution of air temperature, humidity, and ABL height along the Lagrangian trajectory. Compared to dropsonde, radiometer, and lidar data, the simulation accurately represented the integrated water vapour and liquid water path, with a slight cold and dry bias at lower levels. The general agreement demonstrates that the applied forcing and initialisation derived from the HALO-($\mathcal{AC}$)³ measurements provide a robust foundation for reproducing the dynamics and properties of the Arctic atmospheric boundary layer, underscoring the importance of constraining simulations with detailed observations.

Sensitivity experiments with the tested model revealed that subsidence exerts a strong control on the depth of the atmospheric boundary layer, the height of the cloud top, and the partitioning between the liquid and ice phases in clouds. Weak or absent subsidence enables deeper growth of the atmospheric boundary layer, allowing it to approach the theoretical parabolic deepening rate implied by standard bulk mixed-layer modelling. Reduced subsidence also enhances convective overturning, promoting stronger mixed-phase cloud formation and an earlier onset of (also stronger) graupel formation. In contrast, strong subsidence (Sub200) suppresses vertical development and convective activity, leading to thinner and shorter-lived clouds. The earlier and more pronounced graupel peak in the weak-subsidence cases, and an apparent time-correlation with the formation of a decoupled vertical structure in the cloud field (Figure 5.9) indicates that graupel formation is tightly linked to the onset of significant surface-driven convection featuring moist updrafts.

A deeper analysis of the decoupling process provides an additional perspective on these dynamical-microphysical interactions. Two expressions of decoupling are investigated: (i) one based on the buoyancy flux profile, and (ii) the other reflecting the internal vertical thermodynamic structure of the atmospheric boundary layer. A systematic dependence of the onset of decoupling on the prescribed subsidence is identified, indicating an earlier convec-

tive transition in cases with reduced subsidence. Buoyancy-based decoupling always happens first, closely followed by (i) a distinct maximum in the cloud liquid water path, (ii) a peak in the graupel path, (iii) a peak in surface precipitation, and (iv) complete thermodynamic decoupling. This robust order of events and its southward shift with increasing subsidence further highlight the role of subsidence in the transformation of air masses during marine cold air outbreaks. Its impact on the atmospheric boundary layer depth plays a crucial role, promoting buoyancy decoupling below cloud base, which subsequently induces precipitation-driven stabilisation at low levels by accelerating the depletion of the cloud liquid water through graupel formation. This sequence of events eventually drives a transition toward a fully decoupled structure and to open cellular convection later on. These results underscore that subsidence not only constrains the atmospheric boundary layer and cloud depth but also controls the mixed-phase partitioning and the timing of falling hydrometeor formation during marine cold air outbreaks.

These results highlight the role of subsidence in shaping the structure and lifetime of Arctic mixed-phase clouds, although they are based on a single case study. Their applicability across different Arctic conditions, particularly during stronger marine cold air outbreaks, has not yet been assessed. Accordingly, future work should extend this analysis to additional cases and explore how the sensitivity to subsidence compares to variations in other factors, including aerosol and ice nucleating particles (INP) concentrations. Such investigations will help to disentangle the relative importance of dynamical versus microphysical controls across scales on mixed-phase cloud evolution in the convective atmospheric boundary layer at high latitudes.

5.7 APPENDICES

5.7.1 APPENDIX A - COLD AIR OUTBREAK AND STABILITY INDICES

Following [Papritz and Spengler \(2017\)](#) and [Dahlke et al. \(2022\)](#), the MCAO Index M is defined as the difference between the sea surface potential temperature θ_{SKT} and the potential temperature at 850 hPa $\vartheta_{850 \text{ hPa}}$:

$$M = \theta_{\text{SKT}} - \vartheta_{850 \text{ hPa}} . \quad (5.11)$$

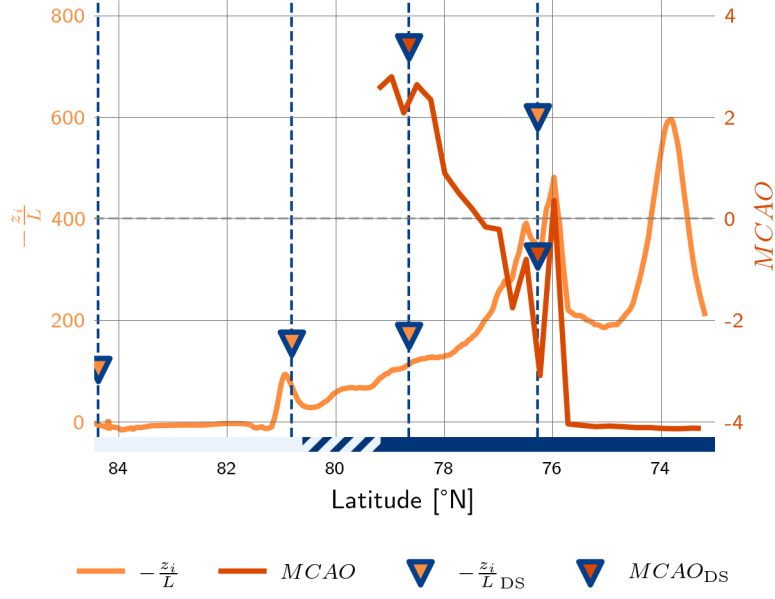


Figure 5.13: MCAO Index from Ctrl model output (red line) and from surface observations using VELOX and dropsonde profiles (red triangles). Stability index $-\frac{z_i}{\mathcal{L}}$ from Ctrl model output (yellow line) and from observations (yellow triangles) along the simulation trajectory.

The Stability Index is expressed as the ratio $\frac{z_i}{\mathcal{L}}$, where z_i is the boundary layer height, diagnosed from the model as the minimum local gradient of virtual potential temperature, and calculated from the averaged dropsonde profiles as the level that reaches a critical value of the bulk Richardson number $Ri_{b,\text{crit}} > 0.25$ (ECMWF, 2016; Troen and Mahrt, 1986). The Obukhov length $\mathcal{L}$ characterises the role of buoyancy in turbulent flows:

$$\mathcal{L} = -\frac{u_*^3 \overline{\theta_v}}{\kappa g (\overline{w'\theta_v'})_s}, \quad (5.12)$$

with frictional velocity u_* , mean virtual potential temperature $\overline{\theta_v}$, surface virtual potential temperature flux $(\overline{w'\theta_v'})_s$ and Kármán constant κ (Stull, 1988). $\mathcal{L}$ is included in the model output and can be estimated from observations using a similar method by applying Monin-Obukhov similarity theory and iteratively solving the stability functions for $\mathcal{L}$ using Ri_b as an initial guess (Heus et al., 2010).

5.7.2 APPENDIX B - TRAJECTORY CALCULATION

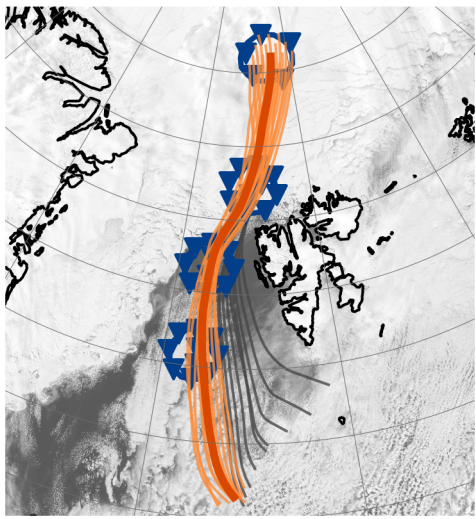


Figure 5.14: Lagrangian trajectories initiated inside IC at 870 hPa that were recaptured in C01, C02 and C03 (orange). Trajectories that were not resampled in the subsequent circle were discarded (grey) from the calculation of the mean trajectory used for the simulation (red). Dropsonde locations are marked with ▼. Satellite image at the time of RF11 29th March 2022 from Terra MODIS (NASA (2022), Worldview Snapshots) in the background.

5.7.3 APPENDIX C - STATISTICS OF OBSERVED MESOSCALE SUBSIDENCE

Table 5.2: Summary of subsidence statistics derived from dropsonde observations for each circle. Mean values are computed over the full 8 km profile, while ranges and uncertainty estimates refer to the lowest 3 km.

Circle	Mean 8 km		Range 3 km		Mean error 3 km		Error range 3 km	
Unit	Pa s ⁻¹	cm s ⁻¹	Pa s ⁻¹	cm s ⁻¹	Pa s ⁻¹	cm s ⁻¹	Pa s ⁻¹	cm s ⁻¹
Factor	10 ⁻²		10 ⁻²		10 ⁻³	10 ⁻³	10 ⁻³	10 ⁻³
IC	8	-0.74	0-11	-1.17-0	2	18	0-2	0-25
C01	-1	0.12	-8-1	-0.07-0.82	1	8	0-1	0-12
C02	7	-0.60	0-13	-1.16-0	1	6	0-1	0.001-9
C03	3	-0.26	0-8	-0.71-0	1	5	0-1	0-8

5.7.4 APPENDIX D - MEAN ABSOLUTE PERCENTAGE ERROR AND AGREEMENT

To calculate the agreement of the simulation with the observational data, this study uses the mean absolute percentage error (MAPE)

$$\text{MAPE} = \frac{100\%}{n} \sum_{i=1}^n \left| \frac{S_i - O_i}{O_i} \right|, \quad (5.13)$$

with the simulation output S_i and the observational data O_i . It follows:

$$\text{Agreement} = 100\% - \text{MAPE}. \quad (5.14)$$

ACKNOWLEDGEMENTS

We gratefully acknowledge the funding by the Deutsche Forschungsgemeinschaft (DFG; German Research Foundation) - Project 268020496 - TRR 172, within the Transregional Collaborative Research Center “Arctic Amplification: Climate Relevant Atmospheric and Surface Processes, and Feedback Mechanisms ($\mathcal{AC}$)³.” We are further grateful for the funding of Project Grant 316646266 by the DFG (German Research Foundation) within the framework of Priority Programme SPP 1294 to promote research with HALO. We also gratefully acknowledge the funding by the DFG (German Research Foundation) - Project 442649391 - Observing Local and Remote Controls on Arctic Air Mass Evolution (ORCA2). We gratefully acknowledge the Gauss Centre for Supercomputing e.V. (www.gauss-centre.eu) for providing computing time on the GCS Supercomputer JUWELS at the Jülich Supercomputing Centre (JSC) under projects RCONGM and HR-AFC. We appreciate the financial contributions of the Alfred Wegener Institute (AWI, Bremerhaven), the German Aerospace Centre (DLR, Oberpfaffenhofen), and the Max Planck Institute for Meteorology (MPI-M, Hamburg). HS acknowledges funding from the European Research Council within the H-2020 programme (Isotopic Links to Atmospheric Water’s Sources (ISLAS), grant nr. 773245). We acknowledge the use of imagery from the Worldview Snapshots application (<https://wvs.earthdata.nasa.gov>), part of the Earth Observing System Data and Information System (EOSDIS). ChatGPT-4o has been used to help create the first versions of this manuscript. Thank you to Niklas Schnierstein and Jan Chylik for developing the DALES-CGN model with representation of Arctic mixed-phase clouds. At last, we want to

acknowledge and thank everyone involved in the HALO-($\mathcal{AC}$)³ and ISLAS2022 campaigns, especially Mario Mech, who kindly provided the liquid water path data from MiRAC for this study.

6

Impacts of Cloud Condensation Nuclei on the Evolution of Clouds in Marine Cold Air Outbreaks

DATA AVAILABILITY

The model code, input and configuration files, generated forcing files, and selected model output are provided in [Paulus \(2026a\)](#). The forcing files are compatible with typical single column model forcings, as DALES is run with horizontally homogeneous initial and boundary conditions. The DALES version used in this study is 4.3 with an extension for mixed-phase microphysics; the specific version applied here is included in [Paulus \(2026a\)](#), while the official DALES repository is available at [Arabas et al. \(2021\)](#).

6.1 INTRODUCTION

Following the analysis of mesoscale subsidence effects on MCAO cloud development in Chapter 5, the following chapter will address the fourth research question:

HOW DOES THE INFLUENCE OF SUBSIDENCE ON CLOUD
EVOLUTION COMPARE TO THE EFFECTS OF THE
CONCENTRATION OF CLOUD CONDENSATION NUCLEI?

Therefore, this chapter investigates how aerosol-controlled processes modify the evolution of MCAOs and whether these sensitivities interact with the previously identified dependence on large-scale subsidence forcing. The previous results demonstrate a link between liquid and ice content, graupel precipitation and boundary layer decoupling, depending on the subsidence forcing. As outlined in Section 1.3.3, these processes are strongly driven by microphysical interactions among cloud hydrometeors, raising the question of whether perturbations in CCN modify these pathways and interact with subsidence forcing in MCAOs.

In the Arctic, cloud evolution may be particularly sensitive to aerosol perturbations because background CCN concentrations are often very low and small changes can strongly alter cloud liquid water, phase partitioning, and cloud persistence (e.g., [Mauritsen et al., 2011](#); [Leaitch et al., 2016](#); [Stevens et al., 2018](#); [Eirund et al., 2019](#)). Observational studies over the central Arctic and the Fram Strait report that CCN concentrations are spatially variable, and influenced by transport, marine emissions, and entrainment from the free troposphere ([Bigg et al., 2001](#); [Koike et al., 2019](#); [Sterzinger and Igel, 2024](#)). In MCAOs, [Williams et al. \(2024\)](#) observed a large spread of CCN concentrations around 100 cm^{-3} . These conditions motivate perturbation experiments spanning pristine to polluted CCN states to investigate the sensitivity of CCN concentrations on the evolution of an MCAO, following the setup of [Chylik et al. \(2023\)](#).

The interaction between aerosol sensitivities and subsidence remains largely unresolved. Recent MCAO studies emphasise the importance of mixed-phase processes, aerosol depletion, and ice-related preconditioning for cloud breakup and downstream transitions. For example, [Tornow et al. \(2021\)](#) showed that riming-related processes can precondition MCAO clouds by reducing cloud liquid water, depleting CCN, and cooling and moistening the subcloud layer prior to substantial rain formation, while [Wu et al. \(2025\)](#) highlighted the role of ice pro-

cesses in accelerating cloud breakup through precipitation evaporation and boundary layer stabilisation. [Young et al. \(2018\)](#) demonstrated that variations in droplet number concentration, which relates to CCN concentrations, exert a relatively weak influence on bulk cloud properties such as liquid and ice water path, but modulate precipitation processes, with lower droplet concentrations enhancing rain formation and showed that these CCN-driven microphysical responses were amplified under stronger large-scale subsidence, acting to regulate the overall sensitivity of the cloud system to aerosol perturbations.

Against this background, the present chapter investigates the sensitivity of different cloud hydrometeors and the ABL development to prescribed CCN perturbations during an MCAO, to identify how aerosol-induced microphysical changes project onto the previously analysed subsidence control on cloud development and ABL decoupling.

6.2 MODEL CONFIGURATION

Following the setup described in Section 5.3, Lagrangian LES were conducted with the DALES model ([Heus et al., 2010](#)) and sensitivity experiments were performed to investigate the influence of CCN concentrations on the ABL evolution and its interaction with subsidence. The experiments were based on the observationally forced control simulation (Ctrl), a simulation without large-scale subsidence (NoSub), and a simulation with reduced subsidence at 50 % of the observed value (Sub50), which are described further in Section 5.3.4.

In the default simulations (Ctrl, NoSub, Sub50), the CCN concentration was initialised with a constant vertical profile of 100 cm^{-3} (see Section 5.3). To study the aerosol sensitivity, additional simulations were conducted with CCN concentrations of 10 cm^{-3} and 1000 cm^{-3} , following the setup of [Chylik et al. \(2023\)](#). To further investigate the CCN sensitivity, two additional simulations were performed for the Ctrl setup with intermediate CCN concentrations of 50 cm^{-3} and 500 cm^{-3} .

Consistent with the simulations discussed in Chapter 5, the model domain covers 12.8 km by 12.8 km, with a horizontal resolution of 50 m. In the vertical, the simulation resolves 191 levels starting at 10 m altitude, with a resolution of 20 m, gradually increasing above 2.1 km.

The large-scale forcings and surface boundary conditions were observationally derived, following the setup of the Ctrl simulation described in Section 5.3.3. Each simulation is run for 45 hours, with the simulated air mass reaching the MIZ approximately 15 hours after the initialisation.

6.3 RESULTS

6.3.1 CCN EFFECTS ACROSS SUBSIDENCE REGIMES

The impact of varying CCN concentrations on the boundary layer structure for different subsidence rates is presented in Figure 6.1, including the sensitivity of liquid water path LWP , ice water path IWP , total precipitation P_{tot} , and boundary layer height z_i to variations in CCN concentration across the different subsidence regimes (Ctrl, NoSub and Sub50). In all simulations, the ABL height z_i shows only minor sensitivity to CCN concentrations. For each subsidence configuration, z_i remains close to the respective default simulations, suggesting that large-scale dynamical control dominates over microphysical perturbations (Figure 6.1 j,k,l). These results are in agreement with [Young et al. \(2018\)](#) who found little change in inversion height when varying droplet number concentrations (used as a proxy for CCN) under different subsidence conditions in Arctic mixed-phase marine stratocumulus during an MCAO case. In contrast, [Chylik et al. \(2023\)](#) reported that higher CCN concentrations can deepen the ABL and strengthen the inversion in mixed-phase convective clouds over the open ocean in the Fram Strait.

The strongest sensitivity to CCN is found in the LWP (Figure 6.1 a,b,c). In all subsidence configurations, LWP increases for both lower and higher CCN concentrations. In the NoSub case (Figure 6.1 b), reducing CCN (CCN_{10}) leads to a pronounced increase of 80 g m^{-2} to 100 g m^{-2} after Co_3 , while increasing CCN (CCN_{1000}) results in a more moderate enhancement of 30 g m^{-2} to 50 g m^{-2} . Similar but weaker responses are found in Sub50 and Ctrl (Figure 6.1 a and c, respectively). After the trajectory initially passes the sea-ice edge, prior to Co_3 , the LWP evolves similarly for all CCN simulations and increases until reaching the peak observed in the default case. However, the magnitude of this peak is enhanced in the perturbed simulations. After the peak, the temporal evolution of LWP follows a similar shape for all cases but remains offset relative to the default simulation.

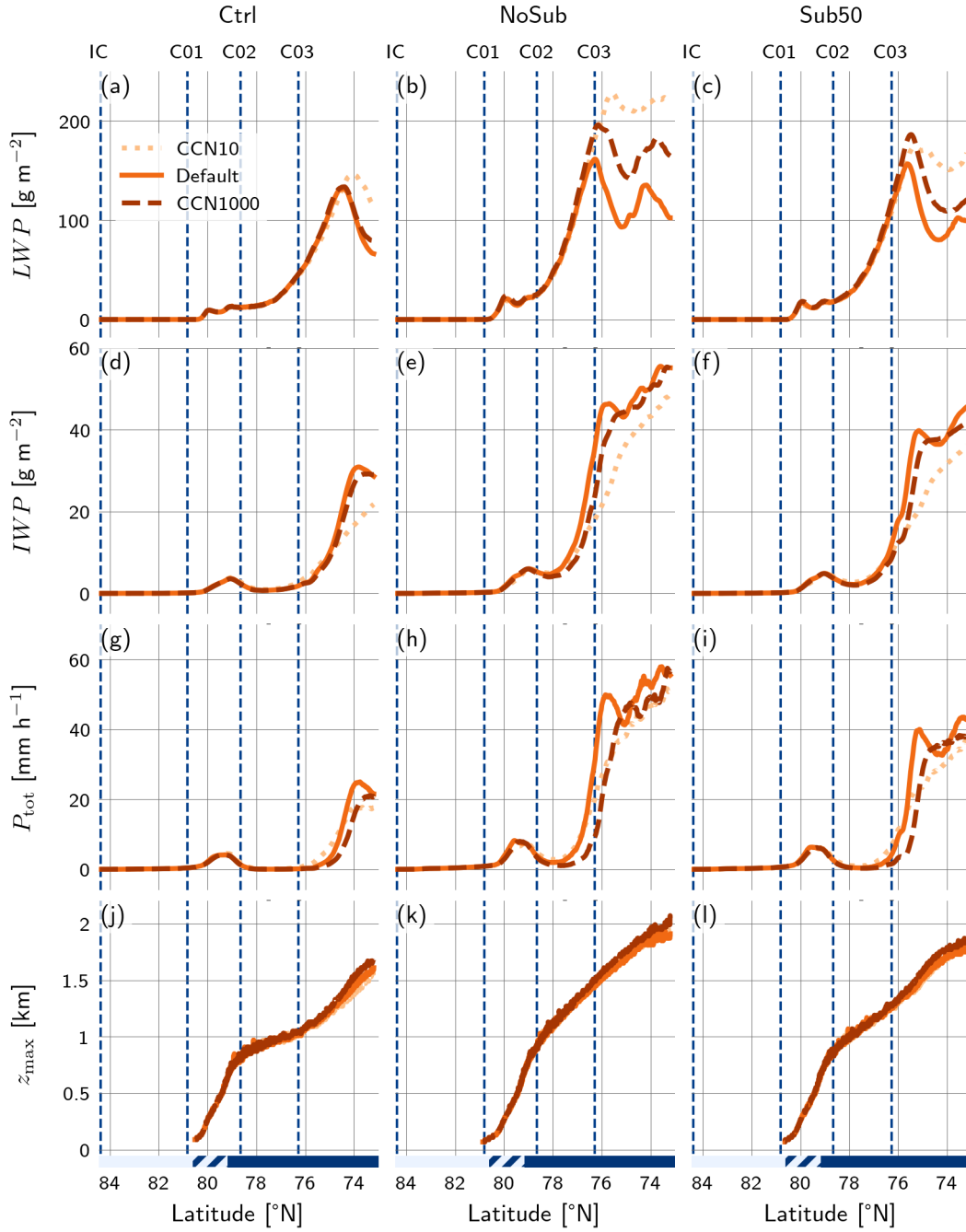


Figure 6.1: Liquid water path (LWP ; panels a, b, c), ice water path (IWP ; panels d, e, f), total precipitation rate (P_{tot} ; panels g, h, i), and boundary layer height (z_i ; panels j, k, l) for Lagrangian LES with different subsidence forcings: Ctrl (left column), NoSub (center column), and Sub50 (right column). Simulations are performed with initial CCN concentrations of 100 cm^{-3} (Default), 10 cm^{-3} (CCN10), and 1000 cm^{-3} (CCN1000). The locations of the dropsondes are indicated by vertical dashed lines. Surface conditions along the trajectory are indicated as sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

This non-linear response reflects competing microphysical pathways. According to the Twomey-Effect (see Section 1.3.4; Twomey, 1974), lower CCN leads to fewer but larger droplets, which can enhance collision-coalescence and promote earlier rain formation (Stevens and Feingold, 2009; Young et al., 2018; Barrett and Hoose, 2023). This accelerates liquid-phase precipitation and partially removes condensate before glaciation. At the same time, larger droplets are more likely to undergo heterogeneous freezing and riming processes (Lance et al., 2011; Chylik et al., 2023).

In contrast, higher CCN produce more numerous, but smaller droplets. This suppresses autoconversion and accretion rates in the Seifert and Beheng (2006) scheme, weakens riming efficiency, and delays precipitation formation. As a result, liquid water is retained longer in the cloud. The ice-water path shows a qualitatively different behaviour (Figure 6.1 d,e,f). Over the MIZ, *IWP* evolves similarly across all simulations. After Co_3 , however, clear differences emerge. Lower CCN leads to a slower increase in *IWP*, with values 5 g m^{-2} to 10 g m^{-2} lower than in the default simulations. Higher CCN delays the onset of *IWP* increase, but this condition eventually reaches comparable values.

The evolution of total precipitation closely follows the development of *IWP* (Figure 6.1 g,h,i). In the default simulations, precipitation increases rapidly after Co_3 and exhibits a pronounced peak. In contrast, low-CCN simulations show an earlier but more gradual onset, resulting in a broader and flatter precipitation curve. This pattern reflects an earlier shift of mass into rain and graupel, reducing the later buildup of suspended ice. High-CCN simulations show delayed precipitation onset, consistent with suppressed rain formation and weaker riming. These simulations reach a plateau without a pronounced peak.

Overall, the response is non-linear, indicating a transition between at least two distinct regimes: (i) a low-CCN regime with enhanced warm-rain and riming, and (ii) a high-CCN regime marked by reduced precipitation efficiency and delayed ice development. The transition between these regimes occurs gradually as CCN increases, with changes in amplitude rather than structure across different subsidence configurations. Consequently, the following analysis focuses on the control simulation (Ctrl) to explore these regime transitions in greater detail within a controlled environment.

Figure 6.2 refines this picture by including intermediate CCN concentrations (CCN₅₀ and CCN₅₀₀). These simulations try to bridge the gap between the CCN₁₀ and CCN₁₀₀₀

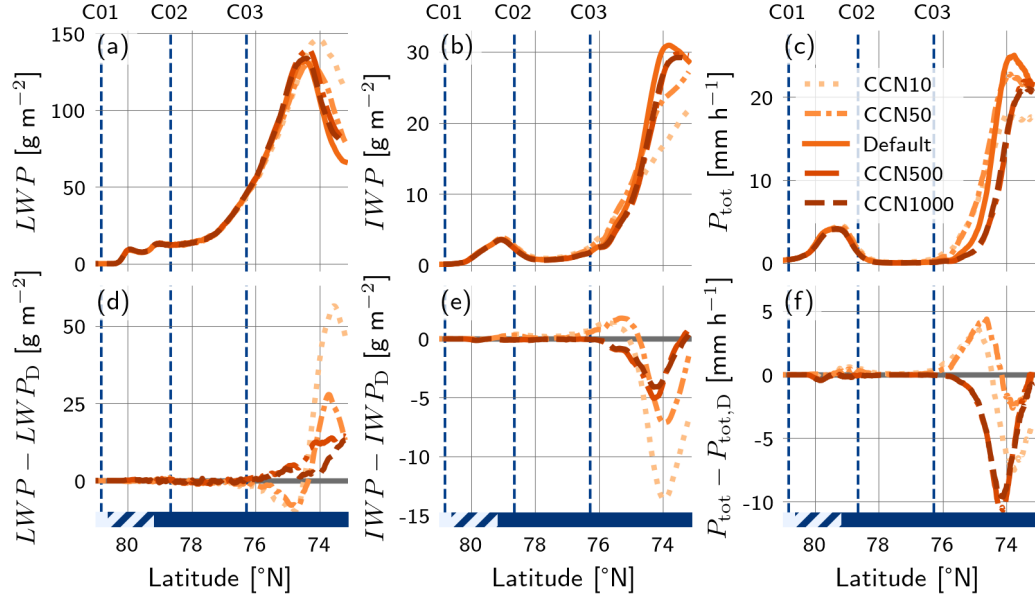


Figure 6.2: Liquid water path (LWP ; a), ice water path (IWP ; b) and total precipitation rate (P_{tot} ; c) for Lagrangian LES with initial CCN concentrations of 100 cm^{-3} (Default), 10 cm^{-3} (CCN10), 50 cm^{-3} (CCN50), 500 cm^{-3} (CCN500) and 1000 cm^{-3} (CCN1000) in the Ctrl subsidence setup. The lower panels indicate the differences in LWP (d), IWP (e) and P_{tot} (f) between the sensitivity studies and the Default simulation. The evolution is shown starting at the marginal sea-ice zone (location of C01), locations of the dropsondes are indicated by vertical dashed lines. Surface conditions along the trajectory are indicated as sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

simulations to better understand responses to increasing and decreasing CCN. The intermediate cases demonstrate that the strong enhancement in LWP (Figure 6.2 a,d) is primarily associated with the low-CCN regime consistently increasing with lower CCN, whereas the high-CCN regime seems to approach saturation. Once droplets are sufficiently small at CCN500, further increases in CCN concentration to CCN1000 do not significantly alter the liquid water path. Similar results are found in the IWP response (Figure 6.2 b and e). Lower CCN results in progressively delayed and reduced peaks, while higher CCN produces only slightly less IWP than the default, with no differences between the CCN500 and CCN1000 simulations.

The precipitation response reflects the combined evolution of liquid and ice phases (Figure 6.2 c,f). In the low-CCN simulations (CCN10 and CCN50), precipitation onset occurs earlier, with rates exceeding the default by approximately 5 mm h^{-1} between 76°N and 74°N . However, peak precipitation is reduced relative to the default simulation, by about

7 mm h^{-1} for CCN_{10} and 2.5 mm h^{-1} for CCN_{50} . Consequently, the total accumulated precipitation remains comparable to the default case but is redistributed over time, with a more gradual onset. In contrast, the high-CCN simulations (CCN_{500} and CCN_{1000}) exhibit later and lower precipitation throughout, with negligible difference between the two.

6.3.2 ICE-PHASE PARTITIONING AND HYDROMETEOR RESPONSE

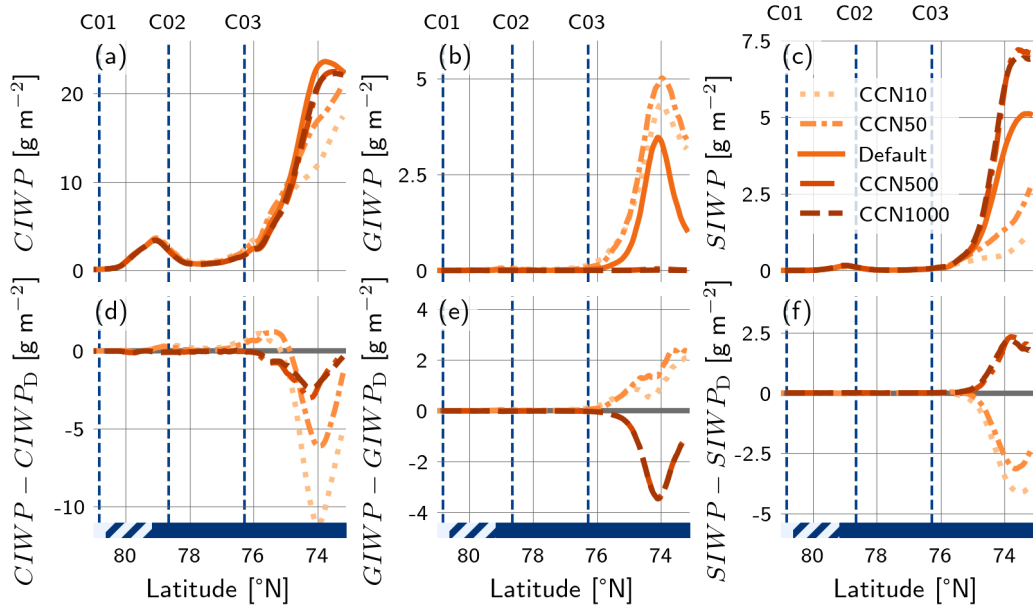


Figure 6.3: Cloud ice water path ($CIWP$; a), graupel water path ($GIWP$; b) and snow water path ($SIWP$; c) for Lagrangian LES with initial CCN concentrations of 100 cm^{-3} (Default), 10 cm^{-3} (CCN_{10}), 50 cm^{-3} (CCN_{50}), 500 cm^{-3} (CCN_{500}) and 1000 cm^{-3} (CCN_{1000}) in the Ctrl subsidence setup. The lower panels indicate the differences in $CIWP$ (d), $GIWP$ (e) and $SIWP$ (f) between the Default configuration and the sensitivity studies. The evolution is shown starting at the marginal sea-ice zone (location of C01), locations of the dropsondes are indicated by vertical dashed lines. Surface conditions along the trajectory are indicated as sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

The response of individual ice hydrometeor species is shown in Figure 6.3. The behaviour of cloud ice $CIWP$ (Figure 6.3 a,d) largely mirrors that of the total ice water path, indicating that cloud ice is the dominant contributor to the IWP signal in these simulations; a similar behaviour was observed in Section 5.5.2 for the subsidence sensitivity. For lower CCN concentrations (CCN_{10} and CCN_{50}), cloud ice formation begins slightly earlier, resulting in a modest increase in $CIWP$ of approximately 2 g m^{-2} between 76°N and 75°N . This

indicates a faster glaciation of larger droplets. However, the subsequent growth of *CIWP* is slower compared to the default simulation. As a result, the peak is delayed and not reached within the simulation period. At the time corresponding to the peak in the default simulation, *CIWP* is reduced by approximately 10 g m^{-2} for CCN_{10} and 6 g m^{-2} for CCN_{50} . For higher CCN concentrations (CCN_{500} and CCN_{1000}), the peak *CIWP* is slightly reduced by approximately 3 g m^{-2} , with no significant differences between the two simulations. This is again consistent with the behaviour observed in the total *IWP* and indicates less efficient glaciation for smaller droplets.

The graupel water path *GIWP* exhibits a different sensitivity to CCN (Figure 6.2 b,e). In the high-CCN simulations (CCN_{500} and CCN_{1000}), graupel formation is entirely suppressed throughout the simulation period. In contrast, for lower CCN concentrations (CCN_{10} and CCN_{50}), graupel production begins slightly earlier, with similar onset timing for both cases, and reaches higher peak values. The magnitude of *GIWP* increases with decreasing CCN, resulting in an enhancement of approximately 2 g m^{-2} towards the end of the simulation for CCN_{10} and CCN_{50} relative to the default case. Graupel growth occurs through riming of supercooled droplets, which appears to be more efficient for larger droplets; similar to the results of [Chylik et al. \(2023\)](#).

The snow water path *SIWP* shows an opposing response (Figure 6.3 c,f). For lower CCN concentrations, snow formation is delayed and less pronounced. This results in a reduction of *SIWP* by approximately 3 g m^{-2} for CCN_{50} and about 4 g m^{-2} for CCN_{10} at the end of the simulation compared to the default case. In contrast, higher CCN concentrations (CCN_{500} and CCN_{1000}) lead to an increase in *SIWP*. Both simulations show nearly identical behaviour, with snow initiating at approximately the same time as in the default simulation but resulting in an enhancement of about 2.5 g m^{-2} in *SIWP*. The microphysics parametrisation by [Seifert and Beheng \(2006\)](#) efficiently converted snow to graupel under strong riming conditions, while snow can develop when riming growth does not exceed depositional growth of the partially rimed snowflakes, with smaller droplet sizes and reduced collisions between ice particles and cloud droplets.

Figure 6.4 illustrates the evolution of the profiles of the five hydrometeor species along the trajectory for all CCN simulations. As expected from the *LWP*, cloud liquid water q_l is slightly enhanced in the high-CCN cases, although this increase is subtle in the vertical profiles.

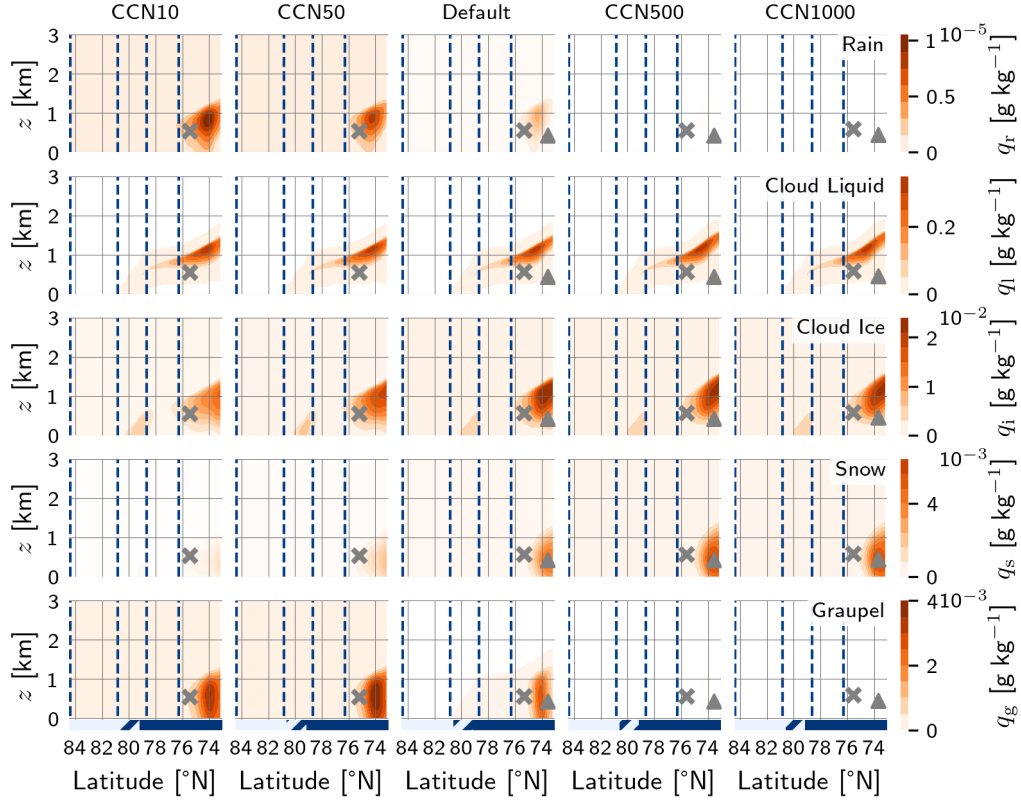


Figure 6.4: Latitudinal evolution five hydrometeor species (rain q_r , cloud liquid q_l , cloud ice q_i , snow q_s and graupel q_g) along the MCAO trajectory for all sensitivity experiments. Dashed horizontal lines indicate dropsonde circle locations. The markers $\times$ and $\blacktriangledown$ indicate decoupling points and heights with respect to $w'\theta_v'$ and Δq_i and $\Delta \theta_l$ respectively (see Section 5.5.2). The coloured bar at the bottom indicates sea-ice ($sic > 0.9$, light blue), marginal sea-ice zone ($0 < sic \leq 0.9$, hatched), and open ocean ($sic = 0$, dark blue).

The additional liquid water is primarily located near the cloud top, reflecting the absence of efficient rain and graupel formation and thus reduced removal of liquid water from the upper cloud layer. Cloud ice shows a maximum around 74°N at about 1 km in all simulations, with reduced magnitude for both low and high CCN, which indicates an optimal droplet number and size for the production of cloud ice in the default CCN case.

Low-CCN simulations exhibit strong rain production with a maximum at about 1 km in the CCN₁₀, CCN₅₀ and a weak maximum in the default CCN case, which seems to evaporate below the cloud, not reaching the surface. This indicates enhanced self-collection processes leading to faster droplet growth and conversion of cloud droplets to liquid precipitation. Enhanced graupel formation and precipitation indicate efficient riming processes.

Snow production shows a clear dependence on CCN, increasing with higher aerosol concentrations. In the high-CCN simulations (CCN₅₀₀ and CCN₁₀₀₀), snow formation becomes more pronounced towards the end of the simulation, consistent with the increase in snow water path. In contrast, the low-CCN simulations produce less snow overall. Graupel production displays the opposite behaviour: it is strongest in the CCN₅₀ case and also pronounced in CCN₁₀, while it is weaker in the default simulation and entirely absent in the high-CCN simulations.

6.3.3 BOUNDARY LAYER DECOUPLING

The evolution of boundary layer decoupling (introduced in Section 5.5.2) shows a notable dependence on CCN. The decoupling point diagnosed from the buoyancy flux remains located in a similar region for all simulations, between approximately 76 °N and 75 °N. However, thermodynamic decoupling, as defined by vertical gradients in moisture and liquid water potential temperature, is delayed in the low-CCN simulations (CCN₁₀ and CCN₅₀), such that it is not detected until the end of the simulation (see Figure 6.4). In contrast, for the default and high-CCN simulations (CCN₅₀₀ and CCN₁₀₀₀), decoupling occurs slightly earlier with increasing CCN concentration.

In the vertical evolution of hydrometeors shown in Figure 6.4, low-CCN simulations exhibit enhanced rain and graupel production, associated with efficient riming and a more distributed precipitation signal. Despite the early onset of precipitation, the gradual removal of condensate seems to delay the stabilisation of the subcloud layer, postponing thermodynamic decoupling. In contrast, high-CCN simulations are characterised by reduced rain and graupel formation, enhanced snow production, and increased liquid water retention near cloud top, potentially strengthening the inversion and promoting earlier thermodynamic separation through snow precipitation.

In Chapter 5, thermodynamic decoupling follows a characteristic sequence in which buoyancy-driven decoupling is followed by maxima in liquid water, graupel, and precipitation, ultimately leading to a fully decoupled structure. The present results indicate that variations in CCN modify both the timing and the phase composition of this sequence. While precipitation remains a key driver of the transition toward decoupling, thermodynamic separation can occur in the absence of graupel, with snow-dominated precipitation providing a similar pathway.

6.4 SUMMARY AND OUTLOOK

This chapter investigated the sensitivities of cloud properties to variations in cloud condensation nuclei concentrations during an MCAO and how these microphysical effects interact with large-scale subsidence. Therefore, sensitivity experiments with CCN concentrations representing pristine conditions (10 cm^{-3}) and continental conditions (1000 cm^{-3}) were performed in addition to the default concentration of 100 cm^{-3} used for the subsidence sensitivity study in Chapter 5. These three simulations were conducted for three subsidence-forcing strengths: the control simulation with observed subsidence (Ctrl), a case without subsidence (NoSub), and a reduced subsidence case with 50 % of the observed forcing (Sub50).

A non-linear sensitivity of the liquid water path, ice water path, and total precipitation was found consistently across all subsidence regimes. The *LWP* is enhanced for both increased and decreased CCN concentrations, with the magnitude of the sensitivity increasing under weaker subsidence. This amplifies the previously identified increase in *LWP* under reduced subsidence conditions (Chapter 5). No significant sensitivity of the boundary layer height to CCN was found in any subsidence regime, indicating that ABL height is primarily controlled by large-scale forcing rather than aerosol-cloud interactions, consistent with [Young et al. \(2018\)](#). The response to increasing CCN appears to saturate at concentrations below 500 cm^{-3} , as evidenced by the nearly indistinguishable results from the CCN500 and CCN1000 simulations across all investigated variables. In contrast, the sensitivity increases toward lower CCN concentrations, with a stronger response between CCN50 and CCN10, consistent with the Twomey effect introduced in Section 1.3.4.

The diagnosed decoupling point based on buoyancy flux remains largely unchanged across all CCN concentrations, indicating that the turbulence-driven boundary layer structure is not strongly affected by microphysical perturbations. In contrast, thermodynamic decoupling shows a clear dependence on CCN. It is delayed in low-CCN simulations (CCN10, CCN50), while in default and high-CCN cases (CCN500, CCN1000) it is shifted upstream. This behaviour can be interpreted in the context of the decoupling sequence identified in Chapter 5, where graupel formation efficiently deplets cloud liquid water and contributes to the stabilisation of the lower boundary layer through precipitation. In the present study, low CCN conditions favour efficient riming and enhanced graupel production, which would be expected to accelerate precipitation-driven stabilisation. However, this process appears more

spatially distributed and gradual, leading to a delayed onset of thermodynamic decoupling. In contrast, high CCN suppresses riming and reduces graupel formation, shifting the system toward snow-dominated precipitation and enhanced liquid water retention aloft. This alters the vertical structure of latent heating and precipitation removal, enabling earlier thermodynamic decoupling despite weaker graupel production.

To further understand the transition between microphysical regimes, additional analysis of process rates, microphysical pathways, and radiative effects is required. The observed non-linear, partially buffered responses of *LWP* and ABL structures differ from those reported in previous studies, underscoring the importance of regime dependence, indicated by [Stevens and Feingold \(2009\)](#), who emphasise that aerosol-cloud-precipitation interactions can be buffered and strongly controlled by environmental conditions.

Future work might also investigate the role of ice nucleating particles in combination with CCN, as both aerosol types jointly control phase partitioning and precipitation formation in mixed-phase clouds. Previous studies demonstrate that INP perturbations can strongly modify *IWP* and precipitation pathways, possibly opposing CCN effects (e.g., [Eirund et al., 2019](#); [Solomon et al., 2009](#); [Wu et al., 2025](#); [Tornow et al., 2021](#)). Understanding the combined influence of CCN and INPs will therefore be essential to constrain microphysical pathways in MCAOs.

7

Conclusions and Outlook

7.1 SUMMARY AND CONCLUSIONS

This thesis investigates the role of atmospheric subsidence in the evolution of marine cold air outbreaks. To understand this process, four research questions were formulated and addressed in three studies utilising dropsonde observations from a case study conducted during the HALO-($\mathcal{AC}$)³ campaign.

The first research question

HOW CAN ATMOSPHERIC SUBSIDENCE BE QUANTIFIED IN THE
ARCTIC USING OBSERVATIONS?

was addressed in Chapter 4 by utilising dropsonde observations from multiple flights and aircraft during the HALO-($\mathcal{AC}$)³ campaign to estimate mesoscale divergence. The sondes were deployed in mesoscale patterns, providing vertically resolved profiles of horizontal wind over an area of about 150 km in diameter. This enables the calculation of

horizontal wind divergence and, through mass continuity, the derivation of mesoscale vertical velocity.

Two methods were employed to calculate divergence: the regression method and the line-integral method. Both approaches had previously been successfully applied in the subtropics (e.g. [Bony and Stevens, 2019](#)) and show remarkable agreement in their divergence profiles in this thesis. The application of these methods to independent observations from two different aircraft over the same target area showed strong agreement, with minor differences likely attributable to small flow changes during the 45 min interval between samplings. The observed divergence profiles generally fluctuate around zero with small magnitudes, typically within $-25 \cdot 10^{-6} \text{ s}^{-1}$ to $25 \cdot 10^{-6} \text{ s}^{-1}$, and exhibit vertically confined layers, which are often less than 1 km deep. These characteristics reflect the comparatively stable Arctic atmosphere and result in weak but dynamically significant subsidence signals.

An error analysis reveals a nonlinear decrease in divergence uncertainty with an increasing number of dropsondes, following a power-law relationship. The greatest reduction in uncertainty occurs when increasing the number of sondes between five and approximately ten, after which improvements become more gradual. Consequently, an optimal range of 10 to 12 dropsondes per pattern is identified to balance statistical robustness and cost efficiency while accounting for potential sonde failures. Overall, these results demonstrate that mesoscale divergence and associated subsidence in the Arctic can be robustly quantified using dropsonde observations arranged in mesoscale sampling patterns, providing reliable vertical profiles for subsequent process studies and modelling applications.

Based on these results, the second research question

CAN A LAGRANGIAN LARGE EDDY SIMULATION, CONSTRAINED
WITH OBSERVATION-DERIVED LARGE-SCALE FORCINGS SUCH AS
SUBSIDENCE, REALISTICALLY REPRODUCE AN OBSERVED
MARINE COLD AIR OUTBREAK?

is approached in the first part of chapter 5 using a Lagrangian large eddy simulation following an observed air mass along its trajectory during a marine cold air outbreak sampled during HALO-(AC)³. The simulation is initialised and continuously forced with observation-derived large-scale forcings, including subsidence, horizontal advection, and

geostrophic wind, obtained from dropsonde measurements via the regression method. Surface boundary conditions are constrained by skin temperature and sea-ice fraction data from airborne remote sensing observations, ensuring a realistic representation of air-sea interactions. This constitutes the first almost entirely observation-driven Lagrangian large eddy simulation in the Arctic, resulting in a tightly constrained framework that is largely independent of uncertainties associated with reanalysis products.

This simulation is evaluated against independent observations and exhibits strong agreement. The evolution of the atmospheric boundary layer height is reproduced within approximately 150 m of the dropsonde- and lidar-derived estimates. It captures both the initial shallow boundary layer over sea-ice and its progressive deepening over the open ocean. The thermodynamic structure is well represented; temperature profiles closely match observations, and humidity profiles show good agreement below the inversion, although some smoothing of observed layering is evident. A small cold and dry bias is present in the atmospheric boundary layer, likely due to uncertainties in the boundary conditions.

The large eddy simulation also reproduces key integrated quantities along the air mass trajectory. Simulated integrated water vapour closely matches both dropsonde and radiometer observations, with minor deviations of approximately 0.1 kg m^{-2} to 0.3 kg m^{-2} . The evolution of the liquid water path is realistically captured, including the onset of cloud formation near the marginal sea-ice zone and its downstream increase. The ice-water path exhibits greater uncertainty due to observational limitations, but simulated values remain within the observed range. The sensible heat fluxes exceed latent heat fluxes, consistent with the strong temperature contrast during cold air outbreaks. Although direct comparisons with observations reveal discrepancies due to methodological differences, applying the same diagnostic method to model output and observations yields good agreement in the relative variability of both heat fluxes. This indicates that the large eddy simulation realistically captures the underlying surface-atmosphere exchange processes.

Overall, these results demonstrate that a Lagrangian large eddy simulation, constrained by detailed observation-based forcing and boundary conditions, can realistically reproduce key dynamical, thermodynamic, and cloud-related features of a marine cold air outbreak. This supports its suitability for process-level investigations and assessment of sensitivities to large-scale forcings such as subsidence, as posed in the third research question:

WHAT IS THE IMPACT OF ATMOSPHERIC SUBSIDENCE ON CLOUD
GLACIATION AND BOUNDARY LAYER DECOUPLING DURING
ARCTIC MARINE COLD AIR OUTBREAKS?

Building on the evaluated control setup, this question is addressed in the second part of Chapter 5 through a series of sensitivity experiments, in which subsidence forcing is systematically modified. In addition to the control simulation, experiments with reduced, enhanced, and entirely removed subsidence are conducted while all other conditions remain constant. This setup enables an isolated investigation of the role of mesoscale vertical motion in cloud and boundary layer structure. The results indicate that subsidence strongly controls the development of the atmospheric boundary layer and cloud fields. Under weak or absent subsidence, the boundary layer deepens more rapidly, exhibiting near-parabolic growth due to enhanced turbulent entrainment. This results in deeper, more vigorous mixed-phase clouds with higher liquid and ice water content.

Conversely, strong subsidence suppresses boundary layer growth, limits mixing, and produces shallower clouds with reduced condensate. Subsidence also significantly affects the cloud glaciation, with reduced subsidence leading to earlier and stronger ice-phase processes, particularly via enhanced graupel formation. The production of graupel is closely linked to convective updrafts, which intensify under weak subsidence. As subsidence increases, these updrafts weaken, delaying and reducing the graupel formation. In the strongest subsidence scenario, graupel formation is nearly entirely suppressed.

A close relationship between cloud glaciation and boundary layer decoupling under varying subsidence conditions was found. Two decoupling types were identified: buoyancy-driven decoupling, indicated by a negative buoyancy flux below the cloud base, followed by thermodynamic decoupling, characterised by vertical gradients in moisture and liquid water potential temperature. Simulations consistently show buoyancy-driven decoupling precedes a sequence of processes shown as a schematic in Figure 7.1. In the deepening boundary layer of a marine cold air outbreak, buoyancy-driven decoupling was found to be followed by a peak in liquid water path, enhanced graupel formation, increased precipitation, and ultimately full thermodynamic decoupling. The timing and location of these processes strongly depend on the subsidence strength. Weaker subsidence results in earlier decoupling and more intense convection and glaciation, whereas stronger subsidence delays and weakens

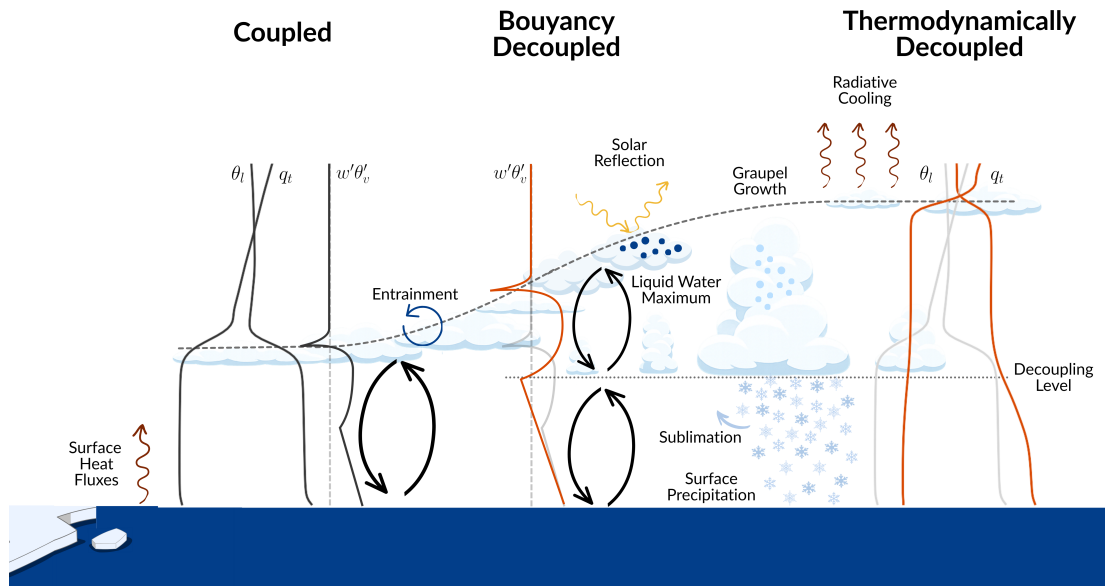


Figure 7.1: Schematic of atmospheric boundary layer development during marine cold air outbreak with coupled (black and grey) and decoupled (orange) profiles of liquid water potential temperature θ_l , total humidity q_t and buoyancy flux $w'\theta'_v$. The top of the boundary layer is indicated as a dashed line.

these transitions. Overall, subsidence influences both boundary layer depth and the internal structure and evolution of the cloud system.

These findings demonstrate that atmospheric subsidence plays a central role in regulating cloud glaciation and boundary layer decoupling. By controlling boundary layer depth, convective activity, and timing of key processes, subsidence governs the transition from coupled stratiform clouds to decoupled convective regimes, thereby shaping the overall evolution of the air mass.

Building on the subsidence sensitivity study, the fourth research question is posed:

HOW DOES THE INFLUENCE OF SUBSIDENCE ON CLOUD
EVOLUTION COMPARE TO THE EFFECTS OF THE
CONCENTRATION OF CLOUD CONDENSATION NUCLEI?

This question is addressed in Chapter 6 by analysing simulations in which both subsidence and cloud condensation nuclei concentrations are systematically varied. In addition to subsidence sensitivity experiments, simulations were conducted with CCN concentrations rang-

ing from pristine (10 cm^{-3}) to polluted (1000 cm^{-3}), while all other forcings remain consistent. This approach enables a direct comparison of subsidence-driven, dynamical and aerosol-driven, microphysical influences on the cloud evolution.

The results indicate that the CCN concentration in this case primarily affects cloud microphysics and precipitation processes, whereas subsidence dominates the atmospheric boundary layer growth and overall evolution of the marine cold air outbreak. Variations in CCN induce a nonlinear response in liquid water path, ice water path, and precipitation. Both elevated and reduced CCN concentrations increase the liquid water path relative to the default case, particularly under weak subsidence. Conversely, the ice water path and total precipitation decrease for both low and high CCN, indicating a buffered, non-monotonic response. The microphysical pathways likely forming these responses differ between high and low CCN regimes. Low CCN concentrations produce fewer, larger droplets, enhancing collision-coalescence and promoting earlier rain formation, which supports efficient riming and increased graupel production. High CCN concentrations generate many smaller droplets, thus suppressing collision-coalescence and delaying precipitation. This increases liquid water retention and shifts the ice growth toward deposition, favouring snow over graupel. Consequently, graupel formation is strongest at low CCN and absent at high CCN. Despite these microphysical effects, CCN was found to exert a minimal influence on the boundary layer height, indicating the dominance of large-scale forcing. Buoyancy-driven decoupling is largely unaffected; however, thermodynamic decoupling was found to be delayed under low CCN conditions and occurred earlier under high CCN, reflecting differences in precipitation phase and amount.

Overall, this thesis presents Lagrangian large eddy simulations of an Arctic marine cold air outbreak, driven by observational boundary conditions and large-scale forcings, using dropsonde-derived subsidence. This setup provides a realistic and tightly constrained basis for controlled sensitivity studies, which show that despite secondary influences of cloud condensation nuclei on the phase-partitioning and precipitation, subsidence governs boundary layer deepening, cloud development, and the timing of buoyancy-driven and thermodynamic boundary layer decoupling.

7.2 OUTLOOK

In this context, upcoming field campaigns such as the COntRasting polar NIght and DAY (CONIDA) campaigns (<https://halo-research.de/science/future-missions/conidaday/> and <https://halo-research.de/science/future-missions/conidanight/>) offer a valuable opportunity to build upon this work. Implementing targeted dropsonde strategies to capture large-scale divergence and vertical motion combined with the long-range research aircraft HALO in these campaigns enables repeated sampling of subsidence along the trajectories of air masses over long distances and multiple days, which was shown to be essential for a Lagrangian model setup. Applying the framework developed in this thesis to cases observed during these campaigns could enhance the process-level understanding of the evolution of air mass transformations and the interactions between mesoscale subsidence, turbulence, and cloud microphysics across different synoptic and surface conditions.

To generalise the findings beyond a single case, it is essential to analyse multiple marine cold air outbreaks of varying strength and structure. However, the currently available observational constraints for other cases often lack coordinated dropsonde strategies specifically designed to quantify mesoscale divergence and subsidence, and might not provide continuous sampling along Lagrangian air mass trajectories. Nevertheless, several cold air outbreaks were observed during HALO-($\mathcal{AC}$)³ and in other recent campaigns such as Cold-Air outbreak Experiment in the Sub-ARctic (CAESAR; Zuidema et al. (2026)), where dropsondes were deployed in flight patterns that could, in principle, be reanalysed to estimate large-scale divergence, even if not optimised for this purpose. In combination with limited-area models such as ICOsahedral Nonhydrostatic model (ICON), which has been shown to reproduce key characteristics of multiple air mass transformation cases during HALO-($\mathcal{AC}$)³ at 2.5 km resolution during HALO-($\mathcal{AC}$)³ (Wendisch et al., 2025), these observations may still provide a basis for reconstructing large-scale forcing. Applying the methods developed in this thesis to such cases would enable a systematic, regime-based analysis of marine cold air outbreaks, distinguishing between different outbreak strengths, surface conditions, and temporal evolutions. In particular, while the present results highlight the dominant role of subsidence, it is plausible that in stronger cold air outbreaks with more intense surface-driven turbulence, its relative influence may be altered. This could be addressed either through sensitivity studies based on the present case with systematically varied air-sea temperature contrasts or by combining additional observed cases with a Lagrangian approach and subsidence sensitivity

experiments. Despite the potential of such extensions, further targeted observations, specifically designed to capture large-scale divergence and to follow air mass trajectories, remain essential to robustly constrain subsidence in Lagrangian LES and ensure realistic representations of Arctic air mass transformation processes in model simulations.

Extending this framework beyond marine cold air outbreaks, warm air intrusions represent a complementary pathway of Arctic air mass transformation, transporting substantial heat and moisture into the Arctic (Lackner et al., 2023; Lauer et al., 2023). As discussed above, the application of a Lagrangian large eddy simulation approach critically depends on sufficiently comprehensive observational constraints, which are currently not available for existing warm air intrusions cases. Nevertheless, these datasets may still provide a useful basis for exploratory studies by combining dropsonde-derived estimates, where feasible, with limited-area models and reanalysis products to approximate the large-scale forcing. Lagrangian large eddy simulations of such cases would allow first assessments of how the impacts of subsidence discovered for marine cold air outbreak translate to warm air intrusions, which are characterised by more complex vertical cloud structures and strong cloud-radiation interactions (Kirbus et al., 2023b). While such approaches would necessarily rely on larger-scale models to constrain the forcings, they may help identify key sensitivities and guide future observational strategies.

Cold air outbreaks in the Southern Hemisphere differ systematically from their Arctic counterparts: they tend to be smaller in spatial extent and intensity but occur more frequently as individual events. They are strongly influenced by katabatic drainage flows and the complex topography of the Antarctic continent, which exerts a strong control on low-level circulation. The Antarctic continent acts as a large reservoir of radiatively cooled air, with major export pathways shaped by features such as the Ross Ice Shelf corridor and coastal orography. Consequently, cold air outbreaks over the Southern Ocean are governed by interactions between katabatic outflows, extratropical cyclones, and the distribution of sea-ice (Papritz et al., 2015; Fletcher et al., 2016a; Smith and Sheridan, 2020). Despite their importance, these processes remain poorly constrained due to the scarcity of observations in the Southern Ocean. Nevertheless, performing controlled subsidence sensitivity studies with Lagrangian large eddy simulations to Antarctic cases, in a more idealised form, could help to disentangle the relative importance of large-scale dynamics and surface-driven processes across hemispheres.

A robust representation of cloud processes in large eddy simulation across different regimes

may require further improvements in the treatment of cloud microphysics, which remains a major source of uncertainty in high-resolution models. Current bulk microphysics schemes rely on simplified descriptions of hydrometeor populations based on assumed particle size distributions and empirical closures, and exhibit substantial uncertainties in key processes, particularly for poorly constrained ice-phase interactions such as riming, aggregation, breakup, and secondary ice production (Morrison et al., 2020). In contrast, particle-based microphysics models such as presented by Brdar and Seifert (2018), explicitly simulate individual ice particles and their interactions, without relying on prescribed distribution shapes or category transitions. This allows for a more physically consistent representation of microphysical pathways and their coupling to dynamics, particularly in mixed-phase cloud regimes where process interactions are highly nonlinear, although computationally expensive. However, it would be valuable to revisit the case presented in this thesis using different microphysical parameterisations. Applying the same Lagrangian framework under identical large-scale forcing but systematically varying the microphysics parametrisation would allow a direct assessment of how robust the identified role of subsidence in cloud glaciation remain across different representations of cloud microphysics.

Nevertheless, the sensitivity experiments conducted in this thesis demonstrate that subsidence exerts systematic control over boundary layer depth and cloud properties. Extending these experiments to ensembles spanning a range of large-scale forcing conditions representative of different dynamical regimes could enable the derivation of quantitative relationships between variables resolved in general circulation models, such as subsidence and stability, and subgrid-scale processes such as entrainment, cloud thickness, and the evolution of the boundary layer depth. A better understanding of these relationships will help to improve the representation of subgrid-scale processes in general circulation models, especially regarding the coupling between large-scale vertical motion and cloud dynamics.

Acknowledgments

I would like to sincerely thank my supervisor, Roel Neggers, for his continuous guidance, support and enthusiasm during my work on this PhD project. His genuine interest in the research and his ability to share his excitement in discussions have been a great source of motivation for me.

I am also very grateful to Vera Schemann and Manfred Wendisch for their support and valuable input as members of my thesis advisory committee. A special thank you to Vera for encouraging me to pursue this work and for always being next door with reassurance and support when I needed it.

I especially want to thank Felix Pithan for examining this thesis.

I would also like to thank Imke, Henning, Niklas, Lena, Xinyuan, Vera and Jan for taking the time to review parts of this work and for their helpful feedback. I feel very lucky to have been part of such a supportive research group. Thank you to Yannick, Xinyuan, Patrick, Ashwath, Niklas, Salima, Jan, Irene and Roel for creating such a collaborative and pleasant work environment. A special thanks to Niklas and Jan for sharing their expertise in LES modelling in the Arctic and the DALES model, and for patiently answering all my questions.

I am very grateful to my office mates Lena, Preethi, and Tracy for making the office such a welcoming place to work. I really appreciate having people around to laugh about coding mistakes, tackle confusing problems together, and celebrate the little moments of success when things finally work out.

My sincere thanks go to Geet for his continuous support, especially at the beginning of my PhD. I also thank Michail and Gunilla for their encouragement and for the many interesting

discussions about measuring subsidence in the Arctic. I'm also grateful to Joshua, Benjamin, Mario, Lars, Andreas, and Harald for all the discussions about gathering data and combining it with simulations, and for helping me gather everything to set up the simulations on which this thesis is based.

I am very thankful to the entire HALO-(AC)³ team and the HALO pilots for making this research possible through their planning and execution of the flights. In particular, I would like to thank the flight PIs Marcus, Davide, and Roel, as well as the dropsonde operators Sebastian, Imke, Anja, and Geet. This work would not have been possible without them.

I would also like to thank the IGMK institute for making me feel welcome, especially during the challenging COVID period. In particular, I would like to thank Dagmar for her efforts in keeping everything organised and running smoothly and always making sure all requirements were met.

Thank you also to the GSGS and IRTG graduate schools, especially Karin and Christa, for their support. I'm grateful to all my fellow (AC)³ PhD students for making this journey such a positive experience. Special thanks to Christa and Marlen for coordinating the project and for always being approachable and supportive.

Finally, I want to thank my friends, my family, and my partner for their unwavering support throughout this journey. Thank you for your patience, encouragement, endless cups of tea, and for always listening to my long (and probably overly detailed) thoughts about the Arctic.

FUNDING ACKNOWLEDGEMENTS

I gratefully acknowledge the funding by the Deutsche Forschungsgemeinschaft (DFG; German Research Foundation) - Project 268020496 - TRR 172, within the Transregional Collaborative Research Center "Arctic Amplification: Climate Relevant Atmospheric and Surface Processes, and Feedback Mechanisms (AC)³." I am further grateful for the funding of Project Grant 316646266 by the DFG (German Research Foundation) within the framework of Priority Programme SPP 1294 to promote research with HALO. I gratefully acknowledge funding from the DFG (German Research Foundation) - Project 442649391 - Observing Local and Remote Controls on Arctic Air Mass Evolution (ORCA2), for making this project possible.

AI STATEMENT

Parts of this thesis were refined with the assistance of the AI-based language models Grammarly and ChatGPT 5.3 to improve clarity and readability. These tools were used for language editing and did not contribute to the intellectual content, analysis, or conclusions and all outputs were critically reviewed and revised by the author.

ENVIRONMENTAL AND ETHICAL CONSIDERATIONS

This study relies on airborne observations, a methodology that enables detailed investigation of Arctic atmospheric processes, particularly in the high Arctic, where conventional measurement infrastructure remains limited. However, it is important to recognise that such approaches are associated with a non-negligible environmental footprint. Aircraft operations contribute to atmospheric emissions, and the deployment of in situ measurement systems, such as dropsondes, introduces additional material waste into these sensitive environments. These impacts are presently unavoidable within current observational strategies in these regions and thus require ongoing, thoughtful evaluation, including environmental impact assessments for scientific expeditions, such as the report by [Macfarlane et al. \(2025\)](#) for the MOSAiC expedition.

This work is also situated within the broader context of Arctic research, much of which is led by institutions based outside the region. This dynamic raises important questions about local participation, representation, and the value of Indigenous knowledge systems. Indigenous approaches, based on long-term local observations, provide critical and complementary insights into environmental changes. Thus striving toward more inclusive and participatory research practices, that recognise the complementarity of different knowledge systems rather than positioning them within a single framework, is important ([Tengö et al., 2014](#)). While addressing these structural challenges extends beyond the scope of this thesis, their relevance needs to be acknowledged and continued efforts toward more inclusive and collaborative research practices are recognised (e.g., [reIMAGINE Arctic Research, 2017](#); [Saami Council, 2025](#)).

List of Mathematical Symbols

$-\frac{z_i}{\mathcal{L}}$	Stability parameter
A	Measurement area of Line-Integral Method
$CIWP$	Integrated water path of cloud ice
D	Horizontal wind divergence
E	Latent heat flux
F_s	Surface flux
$G(\vec{r}, \vec{x})$	Filter function for large eddy simulations
$GIWP$	Integrated water path of graupel
H	Sensible heat flux
IWP	Ice water path
IWV	Integrated water vapour
K_h	Eddy diffusivity coefficient
K_m	Eddy viscosity coefficient
LWP	Liquid water path
L_v	Latent heat of evaporation
M	MCAO Index
P_{tot}	Total surface precipitation rate
Q	Net radiation
$R_{u_i, \xi}$	Residual scalar flux $R_{u_i, \xi} \equiv \widetilde{u_i \xi} - \widetilde{u_i} \widetilde{\xi}$ for an arbitrary variable ξ
Ri_b	Bulk Richardson number
$SIWP$	Integrated water path of snow
T_{skin}	Surface skin temperature
T_v	Virtual air temperature $T_v = \frac{T}{1 - \frac{\epsilon}{p}(1 - \epsilon)}$

T	Air temperature
α	Surface albedo
δ_{ij}	Kronecker-Delta function $\delta_{ij} = \begin{cases} 1 & \text{if } i = j, \\ 0 & \text{if } i \neq j \end{cases}$ for $i, j \in 1, 2, 3$
ϵ_{ijk}	Levi-Civita Symbol $\epsilon_{ijk} = \begin{cases} +1 & \text{if } i, j, k \in \{(1, 2, 3), (2, 3, 1), (3, 1, 2)\} \\ -1 & \text{if } i, j, k \in \{(1, 3, 2), (3, 2, 1), (2, 1, 3)\} \\ 0 & \text{if } i = j, j = k \text{ or } k = i \end{cases}$
κ	Kármán constant $\kappa \approx 0.4$
$\mathcal{L}$	Obukhov length
$\mathcal{R}$	Gas constant for dry air $\mathcal{R} = 287 \text{ JK}^{-1}\text{kg}^{-1}$
ν	Kinematic viscosity
ω	Vertical wind velocity in pressure coordinates
$\overline{w'\theta'_l}$	Flux of total liquid water potential temperature
ϕ	Polar coordinate, azimuth angle
π	Modified pressure $\pi = \frac{\tilde{p}}{\bar{\rho}} - \frac{2}{3}e$
τ_{ij}^r	Anisotropic residual stress tensor $\tau_{ij}^r \equiv \tau_{ij}^R - \frac{2}{3}\delta_{ij} e$
τ	Stress tensor
θ_l	Liquid water potential temperature
θ_v	Virtual potential temperature
θ	Potential temperature of air - temperature an air parcel would attain if adiabatically brought to the level of reference pressure p_0 , $\theta = T \left(\frac{p_0}{p} \right)^{\frac{\mathcal{R}}{c_p}}$
ϱ	Air density
ϑ	Polar coordinate, polar angle
$\vec{\Omega}$	Angular velocity vector of earth's rotation
$\vec{u}$	Three dimensional wind vector $\vec{u} = (u, v, w)$
$\vec{v}_g$	Horizontal geostrophic wind
$\vec{v}_h$	Horizontal wind vector (u, v)
ζ	Stability parameter $\zeta = \frac{z}{\mathcal{L}}$
c_p	Specific heat capacity
e	Residual kinetic energy after applying an LES filter $e = \frac{1}{2}\tau_{ii}^R$
g	Gravitational acceleration, $g = 9.81 \text{ ms}^{-1}$

p	Air pressure
q_i	Specific humidity with respect to ice
q_l	Specific humidity with respect to liquid water
q	Specific humidity
r	Distance from Earth's centre
sic	Sea-ice fraction
t	Time
u_*	Friction velocity
u_g	Zonal component of the geostrophic wind vector
u	Zonal component of the horizontal wind vector
v_g	Meridional component of the geostrophic wind vector
v	Meridional component of the horizontal wind vector
w	Vertical wind velocity
z_0	Roughness length for momentum
z_i	Height of the atmospheric boundary layer
$z_{0,h}$	Roughness length for heat
z_c^b	Cloud base height
z_c^t	Cloud top height
z	Height above sea level

List of Abbreviations

$(\mathcal{AC})^3$	Arctic Amplification: Climate Relevant Atmospheric and Surface Processes and Feedback Mechanisms
AA	Arctic Amplification
ABL	Atmospheric boundary layer
ACTIVATE	Aerosol Cloud Meteorology Interactions over the Western Atlantic Experiment
AMALi	Airborne Mobile Aerosol Lidar
AMS	American Meteorological Society
AVAPS	Airborne Vertical Atmospheric Profiling System
AWI	Alfred-Wegener-Institut
Co1	Circle Leg 1 of RFI1 of HALO- $(\mathcal{AC})^3$
Co2	Circle Leg 2 of RFI1 of HALO- $(\mathcal{AC})^3$
Co3	Circle Leg 3 of RFI1 of HALO- $(\mathcal{AC})^3$
CAESAR	Cold-Air outbreak Experiment in the Sub-ARctic
CCN	Cloud condensation nuclei
CCN ₁₀	Simulation with CCN concentration of 10 cm ⁻³
CCN ₁₀₀₀	Simulation with CCN concentration of 1000 cm ⁻³
CCN ₅₀	Simulation with CCN concentration of 50 cm ⁻³
CCN ₅₀₀	Simulation with CCN concentration of 500 cm ⁻³
CMET	Controlled meteorological balloon
CMIP6	Coupled Model Intercomparison Project Phase 6
COMBLE	Cold air Outbreaks in the Marine Boundary Layer Experiment

CONIDA	COntasting polar NIght and DAy
Ctrl	Control simulation with observational subsidence
DALES	Dutch Atmospheric Large Eddy Simulation
DLR	German Aerospace Center
DNS	Direct numerical simulation
ECMWF	European Centre for Medium-Range Weather Forecasts
ERA5	ECMWF Reanalysis v5
EUREC ⁴ A	Elucidating the role of clouds-circulation coupling in climate
GCM	General circulation model
GPS	Global Positioning System
HALO	High Altitude and Long Range Aircraft
HALO-(AC) ³	Airborne Research Campaign in the Arctic in March and April 2022
HAMP	HALO Microwave Package
IC	Initial Circle of RfIo of HALO-(AC) ³
ICON	ICOsahedral Nonhydrostatic model
IFS Cy43r1	ECMWF Cycle 43r1 model
INP	Ice nucleating particles
ISLAS ₂₀₂₂	Isotopic Links to Atmospheric Water's Sources
LAGRANTO	Lagrangian Analysis Tool
LES	Large eddy simulation
MCAO	Marine cold air outbreak
MiRAC	Microwave Radar/Radiometer for Arctic Clouds
MIZ	Marginal sea-ice zone
MOSAiC	Multidisciplinary drifting Observatory for the Study of Arctic Climate
MOST	Monin–Obukhov similarity theory
MPC	Mixed-phase cloud

NARVAL ₂	Next Generation Remote Sensing for Validation Studies
NoSub	Simulation with no subsidence
P ₅	Polar 5
P ₆	Polar 6
RF	Research Flight
RF ₁₀	Research Flight 10 of HALO-(AC) ³ , 29th March 2022
RF ₁₁	Research Flight 11 of HALO-(AC) ³ , 30th March 2022
RF ₁₈	Research Flight 18 of HALO-(AC) ³ , 12th April 2022
RMSE	Root mean square error
SCM	Single column model
SFS-TKE	Subfilter-scale turbulent kinetic energy
STABLE	Springtime Atmospheric Boundary Layer Experiment
Sub ₁₀	Simulation with 10 % of observed subsidence
Sub ₂₀₀	Simulation with 200 % of observed subsidence
Sub ₅₀	Simulation with 50 % of observed subsidence
TRR	Trans-regional Collaborative Research Center
VELOX	Video airborne Longwave Observations within siX channels
WAI	Warm air intrusion
WBF	Wegener-Bergeron-Findeisen process

References

- S. J. Abel, I. A. Boutle, K. Waite, S. Fox, P. R. A. Brown, R. Cotton, G. Lloyd, T. W. Choularton, and K. N. Bower. The Role of Precipitation in Controlling the Transition from Stratocumulus to Cumulus Clouds in a Northern Hemisphere Cold-Air Outbreak. *Journal of the Atmospheric Sciences*, 74(7):2293–2314, July 2017. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-16-0362.1. URL <https://journals.ametsoc.org/view/journals/atsc/74/7/jas-d-16-0362.1.xml>.
- E. M. Agee. Mesoscale cellular convection over the oceans. *Dynamics of Atmospheres and Oceans*, 10(4):317–341, Jan. 1987. ISSN 0377-0265. doi: 10.1016/0377-0265(87)90023-6. URL <https://www.sciencedirect.com/science/article/pii/0377026587900236>.
- AMAP. Adaptation Actions for a Changing Arctic: Perspectives from the Bering-Chukchi-Beaufort Region. Technical report, Arctic Monitoring and Assessment Programme (AMAP), Oslo, Norway, 2017.
- A. Ansmann, K. Ohneiser, R. Engelmann, M. Radenz, H. Griesche, J. Hofer, D. Althausen, J. M. Creamean, M. C. Boyer, D. A. Knopf, S. Dahlke, M. Maturilli, H. Gebauer, J. Bühl, C. Jimenez, P. Seifert, and U. Wandinger. Annual cycle of aerosol properties over the central Arctic during MOSAiC 2019–2020 – light-extinction, CCN, and INP levels from the boundary layer to the tropopause. *Atmospheric Chemistry and Physics*, 23(19):12821–12849, Oct. 2023. ISSN 1680-7316. doi: 10.5194/acp-23-12821-2023. URL <https://acp.copernicus.org/articles/23/12821/2023/>.
- S. Arabas, S. Axelsen, J. Attema, C. Beets, S. J. Boeing, M. d. Bruine, J. Chylik, H. Cuijpers, J. v. d. Dussen, C. v. Heerwaarden, T. Heus, F. Jansson, H. Jonker, A. Moene, H. Ouwersloot, G. v. d. Oord, S. d. Roode, R. Neggers, X. Pedruzo, P. Siebesma, M. Sikma, B. v. Stratum, J. Vila, and M. v. Zanten. dalesteam/dales: DALES 4.3, Mar. 2021. URL <https://zenodo.org/records/4604726>.
- A. Arakawa and V. R. Lamb. Computational design of the basic dynamical processes of the UCLA general circulation model. Technical report, Jan. 1977. URL <https://ntrs.nasa.gov/citations/19780026757>. NTRS Author Affiliations: California Univ., California, University NTRS Document ID: 19780026757 NTRS Research Center: Legacy CDMS (CDMS).

- A. Arouf, H. Chepfer, J. E. Kay, T. S. L'Ecuyer, and J. Lac. Surface Cloud Warming Increases as Late Fall Arctic Sea Ice Cover Decreases. *Geophysical Research Letters*, 51(3):e2023GL105805, 2024. ISSN 1944-8007. doi: 10.1029/2023GL105805. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2023GL105805>.
- B. W. Atkinson and J. Wu Zhang. Mesoscale shallow convection in the atmosphere. *Reviews of Geophysics*, 34(4):403–431, 1996. ISSN 1944-9208. doi: 10.1029/96RG02623. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/96RG02623>.
- B. Ayarzagüena and J. A. Screen. Future Arctic sea ice loss reduces severity of cold air outbreaks in midlatitudes. *Geophysical Research Letters*, 43(6):2801–2809, 2016. ISSN 1944-8007. doi: 10.1002/2016GL068092. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016GL068092>.
- A. J. Baran. The dependence of cirrus infrared radiative properties on ice crystal geometry and shape of the size-distribution function. 2005. doi: 10.1256/qj.04.91. URL <https://doi.org/10.1256/qj.04.91>.
- A. I. Barrett and C. Hoose. Microphysical Pathways Active Within Thunderstorms and Their Sensitivity to CCN Concentration and Wind Shear. *Journal of Geophysical Research: Atmospheres*, 128(5):e2022JD036965, 2023. ISSN 2169-8996. doi: 10.1029/2022JD036965. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD036965>.
- R. Bennartz, M. D. Shupe, D. D. Turner, V. P. Walden, K. Steffen, C. J. Cox, M. S. Kulie, N. B. Miller, and C. Pettersen. July 2012 Greenland melt extent enhanced by low-level liquid clouds. *Nature*, 496(7443):83–86, Apr. 2013. ISSN 1476-4687. doi: 10.1038/nature12002. URL <https://www.nature.com/articles/nature12002>.
- E. K. Bigg, C. Leck, and E. D. Nilsson. Sudden changes in arctic atmospheric aerosol concentrations during summer and autumn. *Tellus B*, 48(2):254–271, 1996. ISSN 1600-0889. doi: 10.1034/j.1600-0889.1996.t01-1-00009.x. URL <https://onlinelibrary.wiley.com/doi/abs/10.1034/j.1600-0889.1996.t01-1-00009.x>.
- E. K. Bigg, C. Leck, and E. D. Nilsson. Sudden changes in aerosol and gas concentrations in the central Arctic marine boundary layer: Causes and consequences. *Journal of Geophysical Research: Atmospheres*, 106(D23):32167–32185, 2001. ISSN 2156-2202. doi: 10.1029/2000JD900753. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2000JD900753>.
- H. Binder, M. Boettcher, C. M. Grams, H. Joos, S. Pfahl, and H. Wernli. Exceptional Air Mass Transport and Dynamical Drivers of an Extreme Wintertime Arctic Warm Event. *Geophysical Research Letters*, 44(23):12,028–12,036, 2017. ISSN 1944-8007. doi: 10.1002/2017GL075841. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017GL075841>.

- R. Bintanja, R. G. Graversen, and W. Hazeleger. Arctic winter warming amplified by the thermal inversion and consequent low infrared cooling to space. *Nature Geoscience*, 4(11): 758–761, Nov. 2011. ISSN 1752-0908. doi: 10.1038/ngeo1285. URL <https://www.nature.com/articles/ngeo1285>.
- P. Black, L. Harrison, M. Beaubien, R. Bluth, R. Woods, A. Penny, R. W. Smith, and J. D. Doyle. High-Definition Sounding System (HDSS) for Atmospheric Profiling. *Journal of Atmospheric and Oceanic Technology*, 24(4):777–796, 2017. doi: 10.1175/JTECH-D-14-00210.1. URL <https://doi.org/10.1175/JTECH-D-14-00210.1>.
- R. Blackport and J. C. Fyfe. Climate models fail to capture strengthening wintertime North Atlantic jet and impacts on Europe. *Science Advances*, 8(45):eabn3112, Nov. 2022. doi: 10.1126/sciadv.abn3112. URL <https://www.science.org/doi/full/10.1126/sciadv.abn3112>.
- R. C. Boeke and P. C. Taylor. Seasonal energy exchange in sea ice retreat regions contributes to differences in projected Arctic warming. *Nature Communications*, 9(1):5017, Nov. 2018. ISSN 2041-1723. doi: 10.1038/s41467-018-07061-9. URL <https://www.nature.com/articles/s41467-018-07061-9>.
- S. Bony and B. Stevens. Measuring Area-Averaged Vertical Motions with Dropsondes. *Journal of the Atmospheric Sciences*, 2019. doi: 10.1175/JAS-D-18-0141.1. URL <https://journals.ametsoc.org/view/journals/atsc/76/3/jas-d-18-0141.1.xml>.
- S. Bony, R. Colman, V. M. Kattsov, R. P. Allan, C. S. Bretherton, J.-L. Dufresne, A. Hall, S. Hallegatte, M. M. Holland, W. Ingram, D. A. Randall, B. J. Soden, G. Tselioudis, and M. J. Webb. How Well Do We Understand and Evaluate Climate Change Feedback Processes? *Journal of Climate*, 19(15):3445–3482, Aug. 2006. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI3819.1. URL <https://journals.ametsoc.org/view/journals/clim/19/15/jcli3819.1.xml>.
- J. Boussinesq. *Théorie de l'écoulement tourbillonnant et tumultueux des liquides dans les lits rectilignes à grande section / par M. J. Boussinesq...* 1897. URL <https://gallica.bnf.fr/ark:/12148/bpt6k9738517t>.
- S. Brdar and A. Seifert. McSnow: A Monte-Carlo Particle Model for Riming and Aggregation of Ice Particles in a Multidimensional Microphysical Phase Space. *Journal of Advances in Modeling Earth Systems*, 10(1):187–206, 2018. ISSN 1942-2466. doi: 10.1002/2017MS001167. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017MS001167>.
- C. S. Bretherton and M. C. Wyant. Moisture Transport, Lower-Tropospheric Stability, and Decoupling of Cloud-Topped Boundary Layers. *Journal of the Atmospheric Sciences*, 54(1): 148–167, Jan. 1997. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(1997)054

<0148:MTLTSA>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/54/1/1520-0469_1997_054_0148_mtltsa_2.0.co_2.xml.

P.-E. Brilouet, P. Durand, G. Canut, and N. Fourri . Organized Turbulence in a Cold-Air Outbreak: Evaluating a Large-Eddy Simulation with Respect to Airborne Measurements. *Boundary-Layer Meteorology*, 175(1):57–91, Apr. 2020. ISSN 1573-1472. doi: 10.1007/s10546-019-00499-4. URL <https://doi.org/10.1007/s10546-019-00499-4>.

I. Bronshtein, K. Semendiyayev, G. Musiol, and H. M hlig. *Handbook of Mathematics*. Springer, Berlin, Heidelberg, 5 edition, 2007.

L. Bruhwiler, F.-J. W. Parmentier, P. Crill, M. Leonard, and P. I. Palmer. The Arctic Carbon Cycle and Its Response to Changing Climate. *Current Climate Change Reports*, 7(1):14–34, Mar. 2021. ISSN 2198-6061. doi: 10.1007/s40641-020-00169-5. URL <https://doi.org/10.1007/s40641-020-00169-5>.

B. Br mmer. Boundary-layer modification in wintertime cold-air outbreaks from the Arctic sea ice. *Boundary-Layer Meteorology*, 80(1):109–125, July 1996. ISSN 1573-1472. doi: 10.1007/BF00119014. URL <https://doi.org/10.1007/BF00119014>.

B. Br mmer. Boundary Layer Mass, Water, and Heat Budgets in Wintertime Cold-Air Outbreaks from the Arctic Sea Ice. *Monthly Weather Review*, 125:1824–1837, 1997. doi: 10.1175/1520-0493(1997)125<1824:BLMWAH>2.0.CO;2. URL [https://doi.org/10.1175/1520-0493\(1997\)125<1824:BLMWAH>2.0.CO;2](https://doi.org/10.1175/1520-0493(1997)125<1824:BLMWAH>2.0.CO;2).

B. Br mmer. Roll and Cell Convection in Wintertime Arctic Cold-Air Outbreaks. *Journal of the Atmospheric Sciences*, 56(15):2613–2636, 1999. doi: 10.1175/1520-0469(1999)056%3C2613:RACCIW%3E2.0.CO;2. URL [https://doi.org/10.1175/1520-0469\(1999\)056%3C2613:RACCIW%3E2.0.CO;2](https://doi.org/10.1175/1520-0469(1999)056%3C2613:RACCIW%3E2.0.CO;2).

B. Br mmer and S. Pohlmann. Wintertime roll and cell convection over Greenland and Barents Sea regions: A climatology. *Journal of Geophysical Research: Atmospheres*, 105(D12):15559–15566, 2000. ISSN 2156-2202. doi: 10.1029/1999JD900841. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/1999JD900841>.

M. I. Budyko. The effect of solar radiation variations on the climate of the Earth. *Tellus*, 21(5), Jan. 1969. ISSN 3035-9554. doi: 10.3402/tellusa.v21i5.10109. URL <https://tellusjournal.org/articles/10.3402/tellusa.v21i5.10109>.

Y. Cao, S. Liang, X. Chen, T. He, D. Wang, and X. Cheng. Enhanced wintertime greenhouse effect reinforcing Arctic amplification and initial sea-ice melting. *Scientific Reports*, 7(1):8462, Aug. 2017. ISSN 2045-2322. doi: 10.1038/s41598-017-08545-2. URL <https://www.nature.com/articles/s41598-017-08545-2>.

Y.-C. Chen, S.-H. Wang, Q. Min, S. Lu, P.-L. Lin, N.-H. Lin, K.-S. Chung, and E. Joseph. Aerosol impacts on warm-cloud microphysics and drizzle in a moderately polluted environment. *Atmospheric Chemistry and Physics*, 21(6):4487–4502, Mar. 2021. ISSN 1680-7316. doi: 10.5194/acp-21-4487-2021. URL <https://acp.copernicus.org/articles/21/4487/2021/>.

J. Chylik, D. Chechin, R. Dupuy, B. S. Kulla, C. Lüpkes, S. Mertes, M. Mech, and R. A. J. Neggers. Aerosol impacts on the entrainment efficiency of Arctic mixed-phase convection in a simulated air mass over open water. *Atmospheric Chemistry and Physics*, 23(8):4903–4929, Apr. 2023. ISSN 1680-7316. doi: 10.5194/acp-23-4903-2023. URL <https://acp.copernicus.org/articles/23/4903/2023/>.

E. Clapeyron. Memoire sur la puissance motrice de la chaleur. 34, 1834.

R. Clausius. On a modified form of the second fundamental theorem in the mechanical theory of heat. 12:77:81–98, 1856. doi: 10.1080/14786445608642141. URL <https://doi.org/10.1080/14786445608642141>.

R. Colman. Seasonal contributions to climate feedbacks. *Climate Dynamics*, 20(7):825–841, May 2003. ISSN 1432-0894. doi: 10.1007/s00382-002-0301-5. URL <https://doi.org/10.1007/s00382-002-0301-5>.

G. Corbetta, E. Orlandi, T. Heus, R. Neggers, and S. Crewell. Overlap statistics of shallow boundary layer clouds: Comparing ground-based observations with large-eddy simulations. *Geophysical Research Letters*, 42, 2015. doi: 10.1002/2015GL065140. URL <https://doi.org/10.1002/2015GL065140>.

T. W. Cronin and I. Dutta. How Well do We Understand the Planck Feedback? *Journal of Advances in Modeling Earth Systems*, 15(7):e2023MS003729, 2023. ISSN 1942-2466. doi: 10.1029/2023MS003729. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2023MS003729>.

S. Dahlke, A. Solbès, and M. Maturilli. Cold Air Outbreaks in Fram Strait: Climatology, Trends, and Observations During an Extreme Season in 2020. *Journal of Geophysical Research: Atmospheres*, 127(3):e2021JD035741, 2022. ISSN 2169-8996. doi: 10.1029/2021JD035741. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021JD035741>.

A. Dai and J. Deng. Arctic Amplification Weakens the Variability of Daily Temperatures over Northern Middle-High Latitudes. *Journal of Climate*, 34(7):2591–2609, Apr. 2021. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-20-0514.1. URL <https://journals.ametsoc.org/view/journals/clim/34/7/JCLI-D-20-0514.1.xml>.

A. Das, E. E. Clothiaux, and J. Y. Harrington. Ice Crystal Habit Effects on the Resilience of Arctic Mixed-Phase Stratus Clouds in a One-Dimensional Model. *Journal of Geophysical*

Research: Atmospheres, 130(6):e2024JD041305, 2025. ISSN 2169-8996. doi: 10.1029/2024JD041305. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024JD041305>.

G. de Boer, H. Morrison, M. D. Shupe, and R. Hildner. Evidence of liquid dependent ice nucleation in high-latitude stratiform clouds from surface remote sensors. *Geophysical Research Letters*, 38(1), 2011. ISSN 1944-8007. doi: 10.1029/2010GL046016. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2010GL046016>.

S. R. de Roode, T. Frederikse, A. P. Siebesma, A. S. Ackerman, J. Chylik, P. R. Field, J. Fricke, M. Gryschka, A. Hill, R. Honnert, S. K. Krueger, C. Lac, A. T. Lesage, and L. Tomassini. Turbulent Transport in the Gray Zone: A Large Eddy Model Intercomparison Study of the CONSTRAIN Cold Air Outbreak Case. *Journal of Advances in Modeling Earth Systems*, 11(3):597–623, 2019. ISSN 1942-2466. doi: 10.1029/2018MS001443. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018MS001443>.

J. W. Deardorff. Parameterization of the Planetary Boundary layer for Use in General Circulation Models. *Monthly Weather Review*, 100(2):93–106, Feb. 1972. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1972)100<0093:POTPBL>2.3.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/100/2/1520-0493_1972_100_0093_potpbl_2_3_co_2.xml.

J. W. Deardorff. Stratocumulus-capped mixed layers derived from a three-dimensional model. *Boundary-Layer Meteorology*, 18(4):495–527, 1980. doi: 10.1007/BF00119502. URL <https://doi.org/10.1007/BF00119502>.

J. Degirmendžić and M. Marosz. Wintertime Trends in Directional Variability and Intensity of Westerly Air Flow in the Polar Jet Region. *International Journal of Climatology*, 45(9): e8886, 2025. ISSN 1097-0088. doi: 10.1002/joc.8886. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/joc.8886>.

P. J. DeMott, B. E. Swanson, J. M. Creamean, Y. Tobo, T. C. J. Hill, K. R. Barry, I. F. Beck, G. P. Frietas, D. Heslin-Rees, C. P. Lackner, J. Schmale, R. Krejci, P. Zieger, B. Geerts, and S. M. Kreidenweis. Ice nucleating particle sources and transports between the Central and Southern Arctic regions during winter cold air outbreaks. *Elementa: Science of the Anthropocene*, 13(1):00063, Apr. 2025. ISSN 2325-1026. doi: 10.1525/elementa.2024.00063. URL <https://doi.org/10.1525/elementa.2024.00063>.

R. E. Dickinson, G. A. Meehl, and W. M. Washington. Ice-albedo feedback in a CO₂-doubling simulation. *Climatic Change*, 10(3):241–248, July 1987. ISSN 0165-0009. doi: 10.1007/BF00143904. URL <https://www.scopus.com/pages/publications/0023384639>.

H. Dorff, H. Konow, V. Schemann, and F. Ament. Observability of moisture transport divergence in Arctic atmospheric rivers by dropsondes. *Atmospheric Chemistry and Physics*, 24(15):8771–8795, Aug. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-8771-2024. URL <https://acp.copernicus.org/articles/24/8771/2024/>.

ECMWF. IFS documentation CY43R1 - Part IV: Physical processes—IFS documentation CY43R1, 2016.

U. Egerer, A. Ehrlich, M. Gottschalk, H. Griesche, R. A. J. Neggers, H. Siebert, and M. Wendisch. Case study of a humidity layer above Arctic stratocumulus and potential turbulent coupling with the cloud top. *Atmospheric Chemistry and Physics*, 21(8):6347–6364, 2021. doi: 10.5194/acp-21-6347-2021. URL <https://doi.org/10.5194/acp-21-6347-2021>. Publisher: Copernicus GmbH.

A. Ehrlich, S. Crewell, A. Herber, M. Klingebiel, C. Lüpkes, M. Mech, S. Becker, S. Bormann, H. Bozem, M. Buschmann, H.-C. Clemen, E. De La Torre Castro, H. Dorff, R. Dupuy, O. Eppers, F. Ewald, G. George, A. Giez, S. Grawe, C. Goubeyre, J. Hartmann, E. Jäkel, P. Joppe, O. Jourdan, Z. Jurányi, B. Kirbus, J. Lucke, A. E. Luebke, M. Maahn, N. Maherndl, C. Mallaun, J. Mayer, S. Mertes, G. Mioche, M. Moser, H. Müller, V. Pörtge, N. Risse, G. Roberts, S. Rosenburg, J. Röttenbacher, M. Schäfer, J. Schaefer, A. Schäfer, I. Schirmacher, J. Schneider, S. Schnitt, F. Stratmann, C. Tatzelt, C. Voigt, A. Walbröl, A. Weber, B. Wetzel, M. Wirth, and M. Wendisch. A comprehensive in situ and remote sensing data set collected during the HALO–(AC)³ aircraft campaign. *Earth System Science Data*, 17(3):1295–1328, 2025. doi: 10.5194/essd-17-1295-2025. URL <https://doi.org/10.5194/essd-17-1295-2025>. Publisher: Copernicus GmbH.

G. K. Eirund, A. Possner, and U. Lohmann. Response of Arctic mixed-phase clouds to aerosol perturbations under different surface forcings. *Atmospheric Chemistry and Physics*, 19(15):9847–9864, Aug. 2019. ISSN 1680-7316. doi: 10.5194/acp-19-9847-2019. URL <https://acp.copernicus.org/articles/19/9847/2019/>.

D. Etling and R. A. Brown. Roll vortices in the planetary boundary layer: A review. *Boundary-Layer Meteorology*, 65(3):215–248, Aug. 1993. ISSN 1573-1472. doi: 10.1007/BF00705527. URL <https://doi.org/10.1007/BF00705527>.

M. G. Fearon, J. D. Doyle, D. R. Ryglicki, P. M. Finocchio, and M. Sprenger. The Role of Cyclones in Moisture Transport into the Arctic. *Geophysical Research Letters*, 48(4):e2020GL090353, 2021. ISSN 1944-8007. doi: 10.1029/2020GL090353. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2020GL090353>.

N. Feldl, S. Po-Chedley, H. K. A. Singh, S. Hay, and P. J. Kushner. Sea ice and atmospheric circulation shape the high-latitude lapse rate feedback. *npj Climate and Atmospheric Science*,

- 3(1):41, Oct. 2020. ISSN 2397-3722. doi: 10.1038/s41612-020-00146-7. URL <https://www.nature.com/articles/s41612-020-00146-7>.
- J. Feng, D. Paynter, and R. Menzel. How a Stable Greenhouse Effect on Earth Is Maintained Under Global Warming. *Journal of Geophysical Research: Atmospheres*, 128(9): e2022JD038124, 2023. ISSN 2169-8996. doi: 10.1029/2022JD038124. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD038124>.
- P. R. Field, R. J. Cotton, K. McBeath, A. P. Lock, S. Webster, and R. P. Allan. Improving a convection-permitting model simulation of a cold air outbreak. *Quarterly Journal of the Royal Meteorological Society*, 140(678):124–138, 2014. ISSN 1477-870X. doi: 10.1002/qj.2116. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.2116>.
- J. Fletcher, S. Mason, and C. Jakob. The Climatology, Meteorology, and Boundary Layer Structure of Marine Cold Air Outbreaks in Both Hemispheres. *Journal of Climate*, 29(6): 1999–2014, Mar. 2016a. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0268.1. URL <https://journals.ametsoc.org/view/journals/clim/29/6/jcli-d-15-0268.1.xml>.
- J. K. Fletcher, S. Mason, and C. Jakob. A Climatology of Clouds in Marine Cold Air Outbreaks in Both Hemispheres. *Journal of Climate*, 29(18):6677–6692, Sept. 2016b. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0783.1. URL <https://journals.ametsoc.org/view/journals/clim/29/18/jcli-d-15-0783.1.xml>.
- J. D. Ford, D. Clark, T. Pearce, L. Berrang-Ford, L. Copland, J. Dawson, M. New, and S. L. Harper. Changing access to ice, land and water in Arctic communities. *Nature Climate Change*, 9(4):335–339, Apr. 2019. ISSN 1758-6798. doi: 10.1038/s41558-019-0435-7. URL <https://www.nature.com/articles/s41558-019-0435-7>.
- J. A. Francis and E. Hunter. Changes in the fabric of the Arctic’s greenhouse blanket. *Environmental Research Letters*, 2(4):045011, Nov. 2007. ISSN 1748-9326. doi: 10.1088/1748-9326/2/4/045011. URL <https://doi.org/10.1088/1748-9326/2/4/045011>.
- J. A. Francis and S. J. Vavrus. Evidence for a wavier jet stream in response to rapid Arctic warming. *Environmental Research Letters*, 10(1):014005, Jan. 2015. ISSN 1748-9326. doi: 10.1088/1748-9326/10/1/014005. URL <https://doi.org/10.1088/1748-9326/10/1/014005>.
- A. M. Fridlind, A. S. Ackerman, G. McFarquhar, G. Zhang, M. R. Poellot, P. J. DeMott, A. J. Prenni, and A. J. Heymsfield. Ice properties of single-layer stratocumulus during the Mixed-Phase Arctic Cloud Experiment: 2. Model results. *Journal of Geophysical Research: Atmospheres*, 112(D24), 2007. ISSN 2156-2202. doi: 10.1029/2007JD008646. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2007JD008646>.

G. V. Frost, U. S. Bhatt, H. E. Epstein, I. Myers-Smith, G. K. Phoenix, L. T. Berner, J. W. Bjerke, B. C. Forbes, S. J. Goetz, J. T. Kerby, M. J. Macander, T. Park, M. K. Raynolds, H. Tømmervik, and D. A. Walker. Arctic Report Card 2020: Tundra Greenness, 2020.

Q. Fu and K. N. Liou. Parameterization of the Radiative Properties of Cirrus Clouds. *Journal of Atmospheric Sciences*, July 1993. ISSN 1520-0469. URL https://journals.ametsoc.org/view/journals/atsc/50/13/1520-0469_1993_050_2008_potrpo_2_0_co_2.xml. Section: Journal of the Atmospheric Sciences.

T. Garrett, C. Zhao, and P. Novelli. Assessing the relative contributions of transport efficiency and scavenging to seasonal variability in Arctic aerosol. *Tellus B: Chemical and Physical Meteorology*, 62(3):190–196, Jan. 2010. ISSN null. doi: 10.1111/j.1600-0889.2010.00453.x. URL <https://doi.org/10.1111/j.1600-0889.2010.00453.x>.

B. Geerts, S. E. Giangrande, G. M. McFarquhar, L. Xue, S. J. Abel, J. M. Comstock, S. Crewell, P. J. DeMott, K. Ebell, P. Field, T. C. J. Hill, A. Hunzinger, M. P. Jensen, K. L. Johnson, T. W. Juliano, P. Kollias, B. Kosovic, C. Lackner, E. Luke, C. Lüpkes, A. A. Matthews, R. Neggers, M. Ovchinnikov, H. Powers, M. D. Shupe, T. Spengler, B. E. Swanson, M. Tjernström, A. K. Theisen, N. A. Wales, Y. Wang, M. Wendisch, and P. Wu. The COMBLE Campaign: A Study of Marine Boundary Layer Clouds in Arctic Cold-Air Outbreaks. *Bulletin of the American Meteorological Society*, 103(5):E1371–E1389, May 2022. ISSN 0003-0007, 1520-0477. doi: 10.1175/BAMS-D-21-0044.1. URL <https://journals.ametsoc.org/view/journals/bams/103/5/BAMS-D-21-0044.1.xml>.

G. George, B. Stevens, S. Bony, R. Pincus, C. Fairall, H. Schulz, T. Kölling, Q. T. Kalen, M. Klingebiel, H. Konow, A. Lundry, M. Prange, and J. Radtke. JOANNE: Joint dropsonde Observations of the Atmosphere in tropical North Atlantic meso-scale Environments. *Earth System Science Data*, 13(11):5253–5272, Nov. 2021. ISSN 1866-3508. doi: 10.5194/essd-13-5253-2021. URL <https://essd.copernicus.org/articles/13/5253/2021/>. Publisher: Copernicus GmbH.

G. George, B. Stevens, S. Bony, R. Vogel, and A. K. Naumann. Widespread shallow mesoscale circulations observed in the trades. *Nature Geoscience*, 16(7):584–589, July 2023. ISSN 1752-0908. doi: 10.1038/s41561-023-01215-1. URL <https://www.nature.com/articles/s41561-023-01215-1>. Publisher: Nature Publishing Group.

G. George, A. E. Luebke, M. Klingebiel, M. Mech, and A. Ehrlich. Dropsonde measurements from HALO and POLAR 5 during HALO-(AC)³ in 2022, 2024. URL <https://doi.pangaea.de/10.1594/PANGAEA.968891>.

D. Ghatak and J. Miller. Implications for Arctic amplification of changes in the strength of the water vapor feedback. *Journal of Geophysical Research: Atmospheres*, 118(14):7569–7578, 2013. ISSN 2169-8996. doi: 10.1002/jgrd.50578. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/jgrd.50578>.

- R. Gierens, S. Kneifel, M. D. Shupe, K. Ebell, M. Maturilli, and U. Löhnert. Low-level mixed-phase clouds in a complex Arctic environment. *Atmospheric Chemistry and Physics*, 20(6):3459–3481, Mar. 2020. ISSN 1680-7316. doi: 10.5194/acp-20-3459-2020. URL <https://acp.copernicus.org/articles/20/3459/2020/>.
- H. Goosse, J. E. Kay, K. C. Armour, A. Bodas-Salcedo, H. Chepfer, D. Docquier, A. Jonko, P. J. Kushner, O. Lecomte, F. Massonnet, H.-S. Park, F. Pithan, G. Svensson, and M. Vancoppenolle. Quantifying climate feedbacks in polar regions. *Nature Communications*, 9(1):1919, May 2018. ISSN 2041-1723. doi: 10.1038/s41467-018-04173-0. URL <https://www.nature.com/articles/s41467-018-04173-0>.
- A. A. Grachev, E. L. Andreas, C. W. Fairall, P. S. Guest, and P. O. G. Persson. SHEBA flux–profile relationships in the stable atmospheric boundary layer. *Boundary-Layer Meteorology*, 124(3):315–333, Sept. 2007. ISSN 1573-1472. doi: 10.1007/s10546-007-9177-6. URL <https://doi.org/10.1007/s10546-007-9177-6>.
- R. G. Graversen and P. L. Langen. On the Role of the Atmospheric Energy Transport in $2 \times \text{CO}_2$ -Induced Polar Amplification in CESM1. *Journal of Climate*, 32(13):3941–3956, July 2019. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-18-0546.1. URL <https://journals.ametsoc.org/view/journals/clim/32/13/jcli-d-18-0546.1.xml>.
- M. Gryschka and S. Raasch. Roll convection during a cold air outbreak: A large eddy simulation with stationary model domain. *Geophysical Research Letters*, 32(14), 2005. ISSN 1944-8007. doi: 10.1029/2005GL022872. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2005GL022872>.
- M. Gryschka, C. Drüe, D. Etling, and S. Raasch. On the influence of sea-ice inhomogeneities onto roll convection in cold-air outbreaks. *Geophysical Research Letters*, 35(23), 2008. ISSN 1944-8007. doi: 10.1029/2008GL035845. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2008GL035845>.
- M. Gryschka, J. Fricke, and S. Raasch. On the impact of forced roll convection on vertical turbulent transport in cold air outbreaks. *Journal of Geophysical Research: Atmospheres*, 119(22):12,513–12,532, 2014. ISSN 2169-8996. doi: 10.1002/2014JD022160. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014JD022160>.
- J. Hansen, M. Sato, and R. Ruedy. Radiative forcing and climate response. *Journal of Geophysical Research: Atmospheres*, 102(D6):6831–6864, 1997. ISSN 2156-2202. doi: 10.1029/96JD03436. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/96JD03436>.
- G. Hartfield, J. Blunden, and D. S. Arndt. State of the Climate in 2017. *Bulletin of the American Meteorological Society*, 99(8):Si–S310, Aug. 2018. ISSN 0003-0007, 1520-0477.

doi: 10.1175/2018BAMSStateoftheClimate.1. URL <https://journals.ametsoc.org/view/journals/bams/99/8/2018bamsstateoftheclimate.1.xml>.

J. Hartmann, C. Kottmeier, and S. Raasch. Roll Vortices and Boundary-Layer Development during a Cold Air Outbreak. *Boundary-Layer Meteorology*, 84(1):45–65, July 1997. ISSN 1573-1472. doi: 10.1023/A:1000392931768. URL <https://doi.org/10.1023/A:1000392931768>.

M. Hartmann, T. Blunier, S. Brügger, J. Schmale, M. Schwikowski, A. Vogel, H. Wex, and F. Stratmann. Variation of Ice Nucleating Particles in the European Arctic Over the Last Centuries. *Geophysical Research Letters*, 46(7):4007–4016, 2019. ISSN 1944-8007. doi: 10.1029/2019GL082311. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019GL082311>.

P. F. Hein and R. A. Brown. Observations of longitudinal roll vortices during arctic cold air outbreaks over open water. *Boundary-Layer Meteorology*, 45(1):177–199, Oct. 1988. ISSN 1573-1472. doi: 10.1007/BF00120822. URL <https://doi.org/10.1007/BF00120822>.

M. Henry and T. M. Merlis. The Role of the Nonlinearity of the Stefan–Boltzmann Law on the Structure of Radiatively Forced Temperature Change. *Journal of Climate*, 32(2):335–348, Jan. 2019. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-17-0603.1. URL <https://journals.ametsoc.org/view/journals/clim/32/2/jcli-d-17-0603.1.xml>.

H. Hersbach, B. Bell, P. Berrisford, S. Hirahara, A. Horányi, J. Muñoz-Sabater, J. Nicolas, C. Peubey, R. Radu, D. Schepers, A. Simmons, C. Soci, S. Abdalla, X. Abellan, G. Balsamo, P. Bechtold, G. Biavati, J. Bidlot, M. Bonavita, G. De Chiara, P. Dahlgren, D. Dee, M. Diamantakis, R. Dragani, J. Flemming, R. Forbes, M. Fuentes, A. Geer, L. Haimberger, S. Healy, R. J. Hogan, E. Hólm, M. Janisková, S. Keeley, P. Laloyaux, P. Lopez, C. Lupu, G. Radnoti, P. de Rosnay, I. Rozum, F. Vamborg, S. Villaume, and J.-N. Thépaut. The ERA5 global reanalysis. *Quarterly Journal of the Royal Meteorological Society*, 146(730):1999–2049, 2020. ISSN 1477-870X. doi: 10.1002/qj.3803. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.3803>.

T. Heus, C. C. van Heerwaarden, H. J. J. Jonker, A. Pier Siebesma, S. Axelsen, K. van den Dries, O. Geoffroy, A. F. Moene, D. Pino, S. R. de Roode, and J. Vilà-Guerau de Arellano. Formulation of the Dutch Atmospheric Large-Eddy Simulation (DALES) and overview of its applications. *Geoscientific Model Development*, 3(2):415–444, Sept. 2010. ISSN 1991-959X. doi: 10.5194/gmd-3-415-2010. URL <https://gmd.copernicus.org/articles/3/415/2010/>.

T. F. Hock and J. L. Franklin. The NCAR GPS Dropwindsonde. *Bulletin of the American Meteorological Society*, Mar. 1999. ISSN 1520-0477. URL <https://journals.ametsoc.org>

[rg/view/journals/bams/80/3/1520-0477_1999_080_0407_tngd_2_0_co_2.xml](https://doi.org/10.1175/JAM2340.1). Section: Bulletin of the American Meteorological Society.

R. J. Hogan, M. P. Mittermaier, and A. J. Illingworth. The Retrieval of Ice Water Content from Radar Reflectivity Factor and Temperature and Its Use in Evaluating a Mesoscale Model. *Journal of Applied Meteorology and Climatology*, 45(2):301–317, Feb. 2006. ISSN 1558-8424, 1558-8432. doi: 10.1175/JAM2340.1. URL <https://journals.ametsoc.org/view/journals/apme/45/2/jam2340.1.xml>.

J. R. Holton and G. J. Hakim. *An Introduction to Dynamic Meteorology*. Academic Press, 2013. ISBN 9780123848666.

R. Honnert, G. A. Efstathiou, R. J. Beare, J. Ito, A. Lock, R. Neggers, R. S. Plant, H. H. Shin, L. Tomassini, and B. Zhou. The Atmospheric Boundary Layer and the “Gray Zone” of Turbulence: A Critical Review. *Journal of Geophysical Research: Atmospheres*, 125(13):e2019JD030317, 2020. ISSN 2169-8996. doi: 10.1029/2019JD030317. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019JD030317>.

G. K. Hovelsrud, B. Poppel, B. van Oort, and J. D. Reist. Arctic Societies, Cultures, and Peoples in a Changing Cryosphere. *AMBIO*, 40(1):100–110, Dec. 2011. ISSN 1654-7209. doi: 10.1007/s13280-011-0219-4. URL <https://doi.org/10.1007/s13280-011-0219-4>.

C.-Y. Huang and S. Raman. Numerical simulation of January 28 cold air outbreak during GALE part II: The mesoscale circulation and marine boundary layer. *Boundary-Layer Meteorology*, 56(1):51–81, July 1991. ISSN 1573-1472. doi: 10.1007/BF00119962. URL <https://doi.org/10.1007/BF00119962>.

Y. Huang, H. Huang, and A. Shakirova. The Nonlinear Radiative Feedback Effects in the Arctic Warming. *Frontiers in Earth Science*, 9, Aug. 2021. ISSN 2296-6463. doi: 10.3389/feart.2021.693779. URL <https://www.frontiersin.org/journals/earth-science/articles/10.3389/feart.2021.693779/full>. Publisher: Frontiers.

W. Hundsdorfer, B. Koren, M. v. Loon, and J. Verwer. A positive finite-difference advection scheme. 1995. doi: 10.1006/JCPH.1995.1042. URL <https://doi.org/10.1006/jcph.1995.1042>.

Y.-T. Hwang, D. M. W. Frierson, and J. E. Kay. Coupling between Arctic feedbacks and changes in poleward energy transport. *Geophysical Research Letters*, 38(17), 2011. ISSN 1944-8007. doi: 10.1029/2011GL048546. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2011GL048546>.

J. Inoue, Y. Tobo, F. Taketani, and K. Sato. Oceanic Supply of Ice-Nucleating Particles and Its Effect on Ice Cloud Formation: A Case Study in the Arctic Ocean During a Cold-Air Outbreak in Early Winter. *Geophysical Research Letters*, 48(16):e2021GL094646, 2021.

ISSN 1944-8007. doi: 10.1029/2021GL094646. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL094646>.

J. M. Intrieri, C. W. Fairall, M. D. Shupe, P. O. G. Persson, E. L. Andreas, P. S. Guest, and R. E. Moritz. An annual cycle of Arctic surface cloud forcing at SHEBA. *Journal of Geophysical Research: Oceans*, 107(C10):SHE 13–1–SHE 13–14, 2002. ISSN 2156-2202. doi: 10.1029/2000JC000439. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2000JC000439>.

IPCC. *Climate Change 2023: Synthesis Report. Contribution of Working Groups I, II and III to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*. IPCC, Geneva, Switzerland, 2023. doi: 10.59327/IPCC/AR6-9789291691647.

P. E. Isachsen, M. Drivdal, S. Eastwood, Y. Gusdal, G. Noer, and Ø. Saetra. Observations of the ocean response to cold air outbreaks and polar lows over the Nordic Seas. *Geophysical Research Letters*, 40(14):3667–3671, 2013. ISSN 1944-8007. doi: 10.1002/grl.50705. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/grl.50705>.

ISLAS2022. ISLAS2022 field campaign, 2023. URL <https://www.uib.no/en/rg/meten/150202/islas2022-field-campaign>.

M. T. Jenkins and A. Dai. The Impact of Sea-Ice Loss on Arctic Climate Feedbacks and Their Role for Arctic Amplification. *Geophysical Research Letters*, 2021. doi: 10.1029/2021GL094599. URL <https://doi.org/10.1029/2021GL094599>.

E. Johansson, A. Devasthale, M. Tjernström, A. M. L. Ekman, and T. L’Ecuyer. Response of the lower troposphere to moisture intrusions into the Arctic. *Geophysical Research Letters*, 44(5):2527–2536, 2017. ISSN 1944-8007. doi: 10.1002/2017GL072687. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017GL072687>.

C. R. Jones, C. S. Bretherton, and D. Leon. Coupled vs. decoupled boundary layers in VOCALS-REx. *Atmospheric Chemistry and Physics*, 11(14):7143–7153, July 2011. ISSN 1680-7316. doi: 10.5194/acp-11-7143-2011. URL <https://acp.copernicus.org/articles/11/7143/2011/acp-11-7143-2011-discussion.html>.

T. W. Juliano, C. P. Lackner, B. Geerts, B. Kosović, L. Xue, P. Wu, and J. B. Olson. Simulating Mixed-Phase Open Cellular Clouds Observed During COMBLE: Evaluation of Parameterized Turbulence Closure. *Journal of Geophysical Research: Atmospheres*, 129(18):e2024JD040889, 2024. ISSN 2169-8996. doi: 10.1029/2024JD040889. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024JD040889>.

T. W. Juliano, F. Tornow, A. M. Fridlind, A. S. Ackerman, G. S. Elsaesser, B. Geerts, C. P. Lackner, D. Painemal, I. Silber, M. Ovchinnikov, G. Svensson, M. Tjernström,

P. Wu, A. Baró Pérez, P. Bogenschütz, D. Chechin, K. K. Chandrakar, J. Chylik, A. Debolskiy, R. Fadeev, A. Gupta, L. Ickes, M. Karalis, M. Köhler, B. Kosović, P. Kuma, W. Li, E. Mortikov, H. Morrison, R. A. J. Neggers, A. Possner, T. Raatikainen, S. Romakkaniemi, N. Schnierstein, S.-i. Shima, N. Silin, M. Tolstykh, L. Xue, M. Zhang, and X. Zheng. The Cold-Air Outbreaks in the Marine Boundary Layer Experiment model-observation intercomparison project (COMBLE-MIP), Part I: Model specification, observational constraints, and preliminary findings. *EGUsphere*, pages 1–52, Jan. 2026. doi: 10.5194/egusphere-2025-6217. URL <https://egusphere.copernicus.org/preprints/2026/egusphere-2025-6217/>.

M.-L. Kapsch, R. G. Graversen, M. Tjernström, and R. Bintanja. The Effect of Downwelling Longwave and Shortwave Radiation on Arctic Summer Sea Ice. *Journal of Climate*, 29(3):1143–1159, Feb. 2016. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0238.1. URL <https://journals.ametsoc.org/view/journals/clim/29/3/jcli-d-15-0238.1.xml>.

M. Karalis, G. Sotiropoulou, S.J. Abel, E. Bossioli, P. Georgakaki, G. Methymaki, A. Nenes, and M. Tombrou. Effects of secondary ice processes on a stratocumulus to cumulus transition during a cold-air outbreak. *Atmospheric Research*, 277:106302, Oct. 2022. ISSN 0169-8095. doi: 10.1016/j.atmosres.2022.106302. URL <https://www.sciencedirect.com/science/article/pii/S0169809522002885>.

J. E. Kay and A. Gettelman. Cloud influence on and response to seasonal Arctic sea ice loss. *Journal of Geophysical Research: Atmospheres*, 114(D18), 2009. ISSN 2156-2202. doi: 10.1029/2009JD011773. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2009JD011773>.

G. Kellner, C. Wamser, and R. A. Brown. An observation of the planetary boundary layer in the marginal ice zone. *Journal of Geophysical Research: Oceans*, 92(C7):6955–6965, 1987. ISSN 2156-2202. doi: 10.1029/JC092iC07p06955. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/JC092iC07p06955>.

B. Kirbus, J. Chylik, A. Ehrlich, S. Becker, M. Schäfer, R. Neggers, and M. Wendisch. Analysis of an Arctic cold air outbreak during autumn and related air mass transformations forced by surface changes and advection in higher altitudes. *Elementa: Science of the Anthropocene*, 11(1):00079, Nov. 2023a. ISSN 2325-1026. doi: 10.1525/elementa.2023.00079. URL <https://doi.org/10.1525/elementa.2023.00079>.

B. Kirbus, S. Tiedeck, A. Camplani, J. Chylik, S. Crewell, S. Dahlke, K. Ebell, I. Gorodetskaya, H. Griesche, D. Handorf, I. Höschel, M. Lauer, R. Neggers, J. Rückert, M. D. Shupe, G. Spreen, A. Walbröl, M. Wendisch, and A. Rinke. Surface impacts and associated mechanisms of a moisture intrusion into the Arctic observed in mid-April 2020 during MOSAiC. *Frontiers in Earth Science*, 11, Apr. 2023b. ISSN 2296-6463. doi:

10.3389/feart.2023.1147848. URL <https://www.frontiersin.org/journals/earth-science/articles/10.3389/feart.2023.1147848/full>.

B. Kirbus, I. Schirmacher, M. Klingebiel, M. Schäfer, A. Ehrlich, N. Slättberg, J. Lucke, M. Moser, H. Müller, and M. Wendisch. Thermodynamic and cloud evolution in a cold-air outbreak during HALO-(AC)³: quasi-Lagrangian observations compared to the ERA5 and CARRA reanalyses. *Atmospheric Chemistry and Physics*, 24(6):3883–3904, Apr. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-3883-2024. URL <https://acp.copernicus.org/articles/24/3883/2024/>.

S. A. Klein, R. B. McCoy, H. Morrison, A. S. Ackerman, A. Avramov, G. d. Boer, M. Chen, J. N. S. Cole, A. D. Del Genio, M. Falk, M. J. Foster, A. Fridlind, J.-C. Golaz, T. Hashino, J. Y. Harrington, C. Hoose, M. F. Khairoutdinov, V. E. Larson, X. Liu, Y. Luo, G. M. McFarquhar, S. Menon, R. A. J. Neggers, S. Park, M. R. Poellot, J. M. Schmidt, I. Sednev, B. J. Shipway, M. D. Shupe, D. A. Spangenberg, Y. C. Sud, D. D. Turner, D. E. Veron, K. v. Salzen, G. K. Walker, Z. Wang, A. B. Wolf, S. Xie, K.-M. Xu, F. Yang, and G. Zhang. Intercomparison of model simulations of mixed-phase clouds observed during the ARM Mixed-Phase Arctic Cloud Experiment. I: single-layer cloud. *Quarterly Journal of the Royal Meteorological Society*, 135(641):979–1002, 2009. ISSN 1477-870X. doi: 10.1002/qj.416. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.416>.

M. Klingebiel, A. Ehrlich, M. Gryschka, N. Risse, N. Maherndl, I. Schirmacher, S. Rosenberg, S. Hörnig, M. Moser, E. Jäkel, M. Schäfer, H. Deneke, M. Mech, C. Voigt, and M. Wendisch. Airborne observations of cloud properties during their evolution from organized streets to isotropic cloud structures along an Arctic cold-air outbreak. *Atmospheric Chemistry and Physics*, 25(17):9787–9801, Sept. 2025. ISSN 1680-7316. doi: 10.5194/acp-25-9787-2025. URL <https://acp.copernicus.org/articles/25/9787/2025/>.

E. M. Knudsen, B. Heinold, S. Dahlke, H. Bozem, S. Crewell, I. V. Gorodetskaya, G. Heygster, D. Kunkel, M. Maturilli, M. Mech, C. Viceto, A. Rinke, H. Schmithüsen, A. Ehrlich, A. Macke, C. Lüpkes, and M. Wendisch. Meteorological conditions during the ACLOUD/PASCAL field campaign near Svalbard in early summer 2017. *Atmospheric Chemistry and Physics*, 18(24):17995–18022, Dec. 2018. ISSN 1680-7316. doi: 10.5194/acp-18-17995-2018. URL <https://acp.copernicus.org/articles/18/17995/2018/>.

D. Kochkov, J. Yuval, I. Langmore, P. Norgaard, J. Smith, G. Mooers, M. Klöwer, J. Lottes, S. Rasp, P. Düben, S. Hatfield, P. Battaglia, A. Sanchez-Gonzalez, M. Willson, M. P. Brenner, and S. Hoyer. Neural general circulation models for weather and climate. *Nature*, 632(8027):1060–1066, Aug. 2024. ISSN 1476-4687. doi: 10.1038/s41586-024-07744-y. URL <https://www.nature.com/articles/s41586-024-07744-y>.

M. Koike, J. Ukita, J. Ström, P. Tunved, M. Shiobara, V. Vitale, A. Lupi, D. Baumgardner, C. Ritter, O. Hermansen, K. Yamada, and C. A. Pedersen. Year-Round In Situ Mea-

surements of Arctic Low-Level Clouds: Microphysical Properties and Their Relationships With Aerosols. *Journal of Geophysical Research: Atmospheres*, 124(3):1798–1822, 2019. ISSN 2169-8996. doi: 10.1029/2018JD029802. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018JD029802>.

E. W. Kolstad. Higher ocean wind speeds during marine cold air outbreaks. *Quarterly Journal of the Royal Meteorological Society*, 143(706):2084–2092, 2017. ISSN 1477-870X. doi: 10.1002/qj.3068. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.3068>.

E. W. Kolstad and T. J. Bracegirdle. Marine cold-air outbreaks in the future: an assessment of IPCC AR4 model results for the Northern Hemisphere. *Climate Dynamics*, 30(7):871–885, June 2008. ISSN 1432-0894. doi: 10.1007/s00382-007-0331-0. URL <https://doi.org/10.1007/s00382-007-0331-0>.

E. W. Kolstad, T. J. Bracegirdle, and I. A. Seierstad. Marine cold-air outbreaks in the North Atlantic: temporal distribution and associations with large-scale atmospheric circulation. *Climate Dynamics*, 33(2):187–197, Aug. 2009. ISSN 1432-0894. doi: 10.1007/s00382-008-0431-5. URL <https://doi.org/10.1007/s00382-008-0431-5>.

T. W. v. Laar, V. Schemann, and R. A. J. Neggers. Investigating the Diurnal Evolution of the Cloud Size Distribution of Continental Cumulus Convection Using Multiday LES. *Journal of Atmospheric Sciences*, Mar. 2019. doi: 10.1175/JAS-D-18-0084.1. URL <https://journals.ametsoc.org/view/journals/atsc/76/3/jas-d-18-0084.1.xml>. Section: Journal of the Atmospheric Sciences.

C. P. Lackner, B. Geerts, T. W. Juliano, L. Xue, and B. Kosovic. Vertical Structure of Clouds and Precipitation During Arctic Cold-Air Outbreaks and Warm-Air Intrusions: Observations From COMBLE. *Journal of Geophysical Research: Atmospheres*, 128(13):e2022JD038403, 2023. ISSN 2169-8996. doi: 10.1029/2022JD038403. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD038403>.

F. H. Lambert and P. C. Taylor. Regional variation of the tropical water vapor and lapse rate feedbacks. *Geophysical Research Letters*, 41(21):7634–7641, 2014. ISSN 1944-8007. doi: 10.1002/2014GL061987. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2014GL061987>.

S. Lance, M. D. Shupe, G. Feingold, C. A. Brock, J. Cozic, J. S. Holloway, R. H. Moore, A. Nenes, J. P. Schwarz, J. R. Spackman, K. D. Froyd, D. M. Murphy, J. Brioude, O. R. Cooper, A. Stohl, and J. F. Burkhardt. Cloud condensation nuclei as a modulator of ice processes in Arctic mixed-phase clouds. *Atmospheric Chemistry and Physics*, 11(15):8003–8015, Aug. 2011. ISSN 1680-7316. doi: 10.5194/acp-11-8003-2011. URL <https://acp.copernicus.org/articles/11/8003/2011/>.

[//journals.ametsoc.org/view/journals/atot/16/10/1520-0426_1999_016_1384_medavo_2_0_co_2.xml](https://journals.ametsoc.org/view/journals/atot/16/10/1520-0426_1999_016_1384_medavo_2_0_co_2.xml). Section: Journal of Atmospheric and Oceanic Technology.

D. H. Lenschow, V. Savic-Jovicic, and B. Stevens. Divergence and Vorticity from Aircraft Air Motion Measurements. *Journal of Atmospheric and Oceanic Technology*, Dec. 2007. doi: 10.1175/2007JTECHA940.1. URL https://journals.ametsoc.org/view/journals/atot/24/12/2007jtecha940_1.xml. Section: Journal of Atmospheric and Oceanic Technology.

A. Leonard. Energy Cascade in Large-Eddy Simulations of Turbulent Fluid Flows. In *Advances in Geophysics*, volume 18, pages 237–248. Elsevier, Jan. 1975. doi: 10.1016/S0065-2687(08)60464-1. URL <https://www.sciencedirect.com/science/chapter/bookseries/abs/pii/S0065268708604641>.

Q. Li, M. Rapp, G. Stober, and R. Latteck. High-resolution vertical velocities and their power spectrum observed with the MAARSY radar – Part 1: frequency spectrum. *Annales Geophysicae*, 36(2):577–586, Apr. 2018. ISSN 0992-7689. doi: 10.5194/angeo-36-577-2018. URL <https://angeo.copernicus.org/articles/36/577/2018/>. Publisher: Copernicus GmbH.

X.-Y. Li, H. Wang, J. Chen, S. Endo, G. George, B. Cairns, S. Chellappan, X. Zeng, S. Kirschler, C. Voigt, A. Sorooshian, E. Crosbie, G. Chen, R. A. Ferrare, W. I. Gustafson, J. W. Hair, M. M. Kleb, H. Liu, R. Moore, D. Painemal, C. Robinson, A. J. Scarino, M. Shook, T. J. Shingler, K. L. Thornhill, F. Tornow, H. Xiao, L. D. Ziemba, and P. Zuidema. Large-Eddy Simulations of Marine Boundary Layer Clouds Associated with Cold-Air Outbreaks during the ACTIVATE Campaign. Part I: Case Setup and Sensitivities to Large-Scale Forcings. *Journal of the Atmospheric Sciences*, 79(1):73–100, Dec. 2021. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-21-0123.1. URL <https://journals.ametsoc.org/view/journals/atasc/79/1/JAS-D-21-0123.1.xml>.

X.-Y. Li, H. Wang, J. Chen, S. Endo, S. Kirschler, C. Voigt, E. Crosbie, L. D. Ziemba, D. Painemal, B. Cairns, J. W. Hair, A. F. Corral, C. Robinson, H. Dadashazar, A. Sorooshian, G. Chen, R. A. Ferrare, M. M. Kleb, H. Liu, R. Moore, A. J. Scarino, M. A. Shook, T. J. Shingler, K. L. Thornhill, F. Tornow, H. Xiao, and X. Zeng. Large-Eddy Simulations of Marine Boundary Layer Clouds Associated with Cold-Air Outbreaks during the ACTIVATE Campaign. Part II: Aerosol–Meteorology–Cloud Interaction. *Journal of the Atmospheric Sciences*, 80(4):1025–1045, Mar. 2023. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-21-0324.1. URL <https://journals.ametsoc.org/view/journals/atasc/80/4/JAS-D-21-0324.1.xml>. Publisher: American Meteorological Society Section: Journal of the Atmospheric Sciences.

O. Linke, J. Quaas, F. Baumer, S. Becker, J. Chylik, S. Dahlke, A. Ehrlich, D. Handorf, C. Jacob, H. Kalesse-Los, L. Lelli, S. Mehrdad, R. Neggers, J. Riebold, P. S. Garfias, N. Schnier-

stein, M. D. Shupe, C. J. Smith, G. Spreen, B. Verneuil, K. S. Vinjamuri, M. Vountas, M. Wendisch, and M. Wendisch. Constraints on simulated past Arctic amplification and lapse rate feedback from observations. *Atmospheric Chemistry and Physics*, 2023. doi: 10.5194/ACP-23-9963-2023. URL <https://doi.org/10.5194/acp-23-9963-2023>.

A. Q. Liu, G. W. K. Moore, K. Tsuboki, and I. A. Renfrew. The Effect of the Sea-ice Zone on the Development of Boundary-layer Roll Clouds During Cold Air Outbreaks. *Boundary-Layer Meteorology*, 118(3):557–581, Mar. 2006. ISSN 1573-1472. doi: 10.1007/s10546-005-6434-4. URL <https://doi.org/10.1007/s10546-005-6434-4>.

G. Lloyd, T. W. Choularton, K. N. Bower, M. W. Gallagher, J. Crosier, S. O’Shea, S. J. Abel, S. Fox, R. Cotton, and I. A. Boutle. In situ measurements of cloud microphysical and aerosol properties during the break-up of stratocumulus cloud layers in cold air outbreaks over the North Atlantic. *Atmospheric Chemistry and Physics*, 18(23):17191–17206, Dec. 2018. ISSN 1680-7316. doi: 10.5194/acp-18-17191-2018. URL <https://acp.copernicus.org/articles/18/17191/2018/>.

J.-F. Louis. A parametric model of vertical eddy fluxes in the atmosphere. *Boundary-Layer Meteorology*, 17(2):187–202, Sept. 1979. ISSN 1573-1472. doi: 10.1007/BF00117978. URL <https://doi.org/10.1007/BF00117978>.

N. J. Lutsko, M. T. Luongo, C. J. Wall, and T. A. Myers. Correlation Between Cloud Adjustments and Cloud Feedbacks Responsible for Larger Range of Climate Sensitivities in CMIP6. *Journal of Geophysical Research: Atmospheres*, 127(23):e2022JD037486, 2022. ISSN 2169-8996. doi: 10.1029/2022JD037486. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD037486>.

S. Løklingholm, A. M. Johanessen, H. Sodemann, L. R. Hole, and P. Voss. Evaluating the quality and additional value of Controlled meteorological Balloons to observe air mass transformation during marine cold-air outbreak. 2026. to be submitted to JGR-Atmospheres.

A. R. Macfarlane, M. M. Smith, R. Calmer, E. S. Droste, S. Tippenhauer, H. Angot, and V. Mohaupt. Environmental impact assessments of scientific fieldwork as a path to sustainability: A case study from the MOSAiC expedition. *Elementa: Science of the Anthropocene*, 13(1):00035, Nov. 2025. ISSN 2325-1026. doi: 10.1525/elementa.2025.00035. URL <https://doi.org/10.1525/elementa.2025.00035>.

Z. Mages, P. Kollias, Z. Zhu, and E. P. Luke. Surface-based observations of cold-air outbreak clouds during the COMBLE field campaign. *Atmospheric Chemistry and Physics*, 23(6):3561–3574, Mar. 2023. ISSN 1680-7316. doi: 10.5194/acp-23-3561-2023. URL <https://acp.copernicus.org/articles/23/3561/2023/>.

L. Mahrt and D. H. Lenschow. Growth Dynamics of the Convectively Mixed Layer. *Journal of the Atmospheric Sciences*, 33(1):41–51, Jan. 1976. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(1976)033<0041:GDOTCM>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/33/1/1520-0469_1976_033_0041_gdotcm_2_0_co_2.xml.

S. Manabe and R. T. Wetherald. The Effects of Doubling the CO₂ Concentration on the climate of a General Circulation Model. *Journal of the Atmospheric Sciences*, 32(1):3–15, Jan. 1975. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(1975)032<0003:TEODTC>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/32/1/1520-0469_1975_032_0003_teodtc_2_0_co_2.xml.

B. E. Mapes and J. Lin. Doppler Radar Observations of Mesoscale Wind Divergence in Regions of Tropical Convection. *Monthly Weather Review*, 133(7):1808–1824, July 2005. ISSN 1520-0493, 0027-0644. doi: 10.1175/MWR2941.1. URL <https://journals.ametsoc.org/view/journals/mwre/133/7/mwr2941.1.xml>.

B. E. Mapes, P. E. Ciesielski, and R. H. Johnson. Sampling Errors in Rawinsonde-Array Budgets. *Journal of the Atmospheric Sciences*, 60(21):2697–2714, Nov. 2003. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(2003)060<2697:SEIRB>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/60/21/1520-0469_2003_060_2697_seirb_2_0_co_2.xml.

A. Marcheggiani and T. Spengler. Cold air outbreaks drive near-surface baroclinicity variability over storm track entrance regions in the Northern Hemisphere. *Weather and Climate Dynamics*, 6(4):1479–1489, Nov. 2025. doi: 10.5194/wcd-6-1479-2025. URL <https://wcd.copernicus.org/articles/6/1479/2025/>.

P. J. Mason. A numerical study of cloud streets in the planetary boundary layer. *Boundary-Layer Meteorology*, 32(3):281–304, July 1985. ISSN 1573-1472. doi: 10.1007/BF00121884. URL <https://doi.org/10.1007/BF00121884>.

M. E. Mateling, C. Pettersen, M. S. Kulie, and T. S. L’Ecuyer. Marine Cold-Air Outbreak Snowfall in the North Atlantic: A CloudSat Perspective. *Journal of Geophysical Research: Atmospheres*, 128(10):e2022JD038053, 2023. ISSN 2169-8996. doi: 10.1029/2022JD038053. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022JD038053>.

T. Mauritsen, J. Sedlar, M. Tjernström, C. Leck, M. Martin, M. Shupe, S. Sjogren, B. Sierau, P. O. G. Persson, I. M. Brooks, and E. Swietlicki. An Arctic CCN-limited cloud-aerosol regime. *Atmospheric Chemistry and Physics*, 11(1):165–173, Jan. 2011. ISSN 1680-7316. doi: 10.5194/acp-11-165-2011. URL <https://acp.copernicus.org/articles/11/165/2011/>.

I. L. McCoy, R. Wood, and J. K. Fletcher. Identifying Meteorological Controls on Open and Closed Mesoscale Cellular Convection Associated with Marine Cold Air Outbreaks. *Journal of Geophysical Research: Atmospheres*, 122(21):11,678–11,702, 2017. ISSN 2169-8996. doi: 10.1002/2017JD027031. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017JD027031>.

G. M. McFarquhar and A. J. Heymsfield. The Definition and Significance of an Effective Radius for Ice Clouds. *Journal of Atmospheric Sciences*, June 1998. ISSN 1520-0469. URL https://journals.ametsoc.org/view/journals/atmsc/55/11/1520-0469_1998_055_2039_tdasoa_2.0.co_2.xml. Section: Journal of the Atmospheric Sciences.

A. D. McGuire, L. G. Anderson, T. R. Christensen, S. Dallimore, L. Guo, D. J. Hayes, M. Heimann, T. D. Lorenson, R. W. Macdonald, and N. Roulet. Sensitivity of the carbon cycle in the Arctic to climate change. *Ecological Monographs*, 79(4):523–555, 2009. ISSN 1557-7015. doi: 10.1890/08-2025.1. URL <https://onlinelibrary.wiley.com/doi/abs/10.1890/08-2025.1>.

M. Mech, E. Orlandi, S. Crewell, F. Ament, L. Hirsch, M. Hagen, G. Peters, and B. Stevens. HAMP – the microwave package on the High Altitude and LOng range research aircraft (HALO). *Atmospheric Measurement Techniques*, 7(12):4539–4553, Dec. 2014. ISSN 1867-1381. doi: 10.5194/amt-7-4539-2014. URL <https://amt.copernicus.org/articles/7/4539/2014/>. Publisher: Copernicus GmbH.

M. Mech, L.-L. Kliesch, A. Anhäuser, T. Rose, P. Kollias, and S. Crewell. Microwave Radar/radiometer for Arctic Clouds (MiRAC): first insights from the ACLOUD campaign. *Atmospheric Measurement Techniques*, 12(9):5019–5037, Sept. 2019. ISSN 1867-1381. doi: 10.5194/amt-12-5019-2019. URL <https://amt.copernicus.org/articles/12/5019/2019/>. Publisher: Copernicus GmbH.

M. Mech, N. Risse, G. Marrollo, and D. Paul. ac3airborne, Aug. 2022. URL <https://zenodo.org/records/7305586>.

M. Mech, N. Risse, P. Krobot, D. Paul, I. Schirmacher, S. Schnitt, and S. Crewell. Radar reflectivities at 94 GHz and microwave brightness temperature measurements at 89 GHz during the HALO-(AC)³ Arctic airborne campaign, 2024. URL <https://doi.pangaea.de/10.1594/PANGAEA.964977>.

N. Melia, K. Haines, and E. Hawkins. Sea ice decline and 21st century trans-Arctic shipping routes. *Geophysical Research Letters*, 43(18):9720–9728, 2016. ISSN 1944-8007. doi: 10.1002/2016GL069315. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016GL069315>.

J. P. Mellado. Direct numerical simulation of free convection over a heated plate. *Journal of Fluid Mechanics*, 712:418–450, Dec. 2012. ISSN 0022-1120, 1469-7645. doi: 10.1017/

jfm.2012.428. URL <https://www.cambridge.org/core/journals/journal-of-fluid-mechanics/article/abs/direct-numerical-simulation-of-free-convection-over-a-heated-plate/18BCBDA83EB33D896A9CB06ED524513D>.

J. Michaelis, A. U. Schmitt, C. Lüpkes, J. Hartmann, G. Birnbaum, and T. Vihma. Observations of marine cold-air outbreaks: a comprehensive data set of airborne and dropsonde measurements from the Springtime Atmospheric Boundary Layer Experiment (STABLE). *Earth System Science Data*, 14(4):1621–1637, Apr. 2022. ISSN 1866-3508. doi: 10.5194/essd-14-1621-2022. URL <https://essd.copernicus.org/articles/14/1621/2022/>.

J. D. Mirocha and B. Kosović. A Large-Eddy Simulation Study of the Influence of Subsidence on the Stably Stratified Atmospheric Boundary Layer. *Boundary-Layer Meteorology*, 134(1):1–21, Jan. 2010. ISSN 1573-1472. doi: 10.1007/s10546-009-9449-4. URL <https://doi.org/10.1007/s10546-009-9449-4>.

A. S. Monin and A. M. Obukhov. Basic laws of turbulent mixing in the surface layer of the atmosphere. 24, 1954.

H. Morrison, M. D. Shupe, J. O. Pinto, and J. A. Curry. Possible roles of ice nucleation mode and ice nuclei depletion in the extended lifetime of Arctic mixed-phase clouds. *Geophysical Research Letters*, 32(18), Sept. 2005. ISSN 1944-8007. doi: 10.1029/2005GL023614. URL <https://agupubs.onlinelibrary.wiley.com/doi/10.1029/2005GL023614>.

H. Morrison, P. Zuidema, A. S. Ackerman, A. Avramov, G. de Boer, J. Fan, A. M. Fridlind, T. Hashino, J. Y. Harrington, Y. Luo, M. Ovchinnikov, and B. Shipway. Intercomparison of cloud model simulations of Arctic mixed-phase boundary layer clouds observed during SHEBA/FIRE-ACE. *Journal of Advances in Modeling Earth Systems*, 3(2), 2011. ISSN 1942-2466. doi: 10.1029/2011MS000066. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2011MS000066>.

H. Morrison, G. de Boer, G. Feingold, J. Harrington, M. D. Shupe, and K. Sulia. Resilience of persistent Arctic mixed-phase clouds. *Nature Geoscience*, 5(1):11–17, Jan. 2012. ISSN 1752-0908. doi: 10.1038/ngeo1332. URL <https://www.nature.com/articles/ngeo1332>.

H. Morrison, M. van Lier-Walqui, A. M. Fridlind, W. W. Grabowski, J. Y. Harrington, C. Hoose, A. Korolev, M. R. Kumjian, J. A. Milbrandt, H. Pawlowska, D. J. Posselt, O. P. Prat, K. J. Reimel, S.-I. Shima, B. van Dierenhoven, and L. Xue. Confronting the Challenge of Modeling Cloud and Precipitation Microphysics. *Journal of Advances in Modeling Earth Systems*, 12(8):e2019MS001689, 2020. ISSN 1942-2466. doi: 10.1029/2019MS001689. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019MS001689>.

J. Mortin, G. Svensson, R. G. Graversen, M.-L. Kapsch, J. C. Stroeve, and L. N. Boisvert. Melt onset over Arctic sea ice controlled by atmospheric moisture transport. *Geophysical*

Research Letters, 43(12):6636–6642, 2016. ISSN 1944-8007. doi: 10.1002/2016GL069330. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016GL069330>.

R. J. Murray-Watson and E. Gryspeerdt. Air mass history linked to the development of Arctic mixed-phase clouds. *Atmospheric Chemistry and Physics*, 24(19):11115–11132, Oct. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-11115-2024. URL <https://acp.copernicus.org/articles/24/11115/2024/>.

R. J. Murray-Watson, E. Gryspeerdt, and T. Goren. Investigating the development of clouds within marine cold-air outbreaks. *Atmospheric Chemistry and Physics*, 23(16):9365–9383, Aug. 2023. ISSN 1680-7316. doi: 10.5194/acp-23-9365-2023. URL <https://acp.copernicus.org/articles/23/9365/2023/>.

S. Murto, R. Caballero, G. Svensson, and L. Papritz. Interaction between Atlantic cyclones and Eurasian atmospheric blocking drives wintertime warm extremes in the high Arctic. *Weather and Climate Dynamics*, 3(1):21–44, Jan. 2022. doi: 10.5194/wcd-3-21-2022. URL <https://wcd.copernicus.org/articles/3/21/2022/>.

S. Murto, L. Papritz, G. Messori, R. Caballero, G. Svensson, and H. Wernli. Extreme Surface Energy Budget Anomalies in the High Arctic in Winter. *Journal of Climate*, 36(11):3591–3609, May 2023. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-22-0209.1. URL <https://journals.ametsoc.org/view/journals/clim/36/11/JCLI-D-22-0209.1.xml>.

G. Müller and A. Chlond. Three-dimensional numerical study of cell broadening during cold-air outbreaks. *Boundary-Layer Meteorology*, 81(3):289–323, Dec. 1996. ISSN 1573-1472. doi: 10.1007/BF02430333. URL <https://doi.org/10.1007/BF02430333>.

G. Müller, B. Brümmer, and W. Alpers. Roll Convection within an Arctic Cold-Air Outbreak: Interpretation of In Situ Aircraft Measurements and Spaceborne SAR Imagery by a Three-Dimensional Atmospheric Model. *Monthly Weather Review*, 127(3):363–380, Mar. 1999. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1999)127<0363:RCWAAC>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/127/3/1520-0493_1999_127_0363_rcwaac_2.0.co_2.xml.

J. J. Müller, M. Schäfer, S. Rosenburg, A. Ehrlich, and M. Wendisch. High-resolution maps of Arctic surface skin temperature and type retrieved from airborne thermal infrared imagery collected during the HALO-(AC)³ campaign. *Atmospheric Measurement Techniques*, 18(18):4695–4708, Sept. 2025. ISSN 1867-1381. doi: 10.5194/amt-18-4695-2025. URL <https://amt.copernicus.org/articles/18/4695/2025/>. Publisher: Copernicus GmbH.

M. Nakanishi and T. Michibata. How Does Cloud Emissivity Feedback Affect Present and Future Arctic Warming? *Ocean-Land-Atmosphere Research*, 4:0089, Apr. 2025. doi: 10.34133/olar.0089. URL <https://spj.science.org/doi/full/10.34133/olar.0089>.

NASA. Worldview Snapshots, 2022. URL <https://wvs.earthdata.nasa.gov/>.

S. M. Natali, J. D. Watts, B. M. Rogers, S. Potter, S. M. Ludwig, A.-K. Selbmann, P. F. Sullivan, B. W. Abbott, K. A. Arndt, L. Birch, M. P. Björkman, A. A. Bloom, G. Celis, T. R. Christensen, C. T. Christiansen, R. Commane, E. J. Cooper, P. Crill, C. Czimczik, S. Davydov, J. Du, J. E. Egan, B. Elberling, E. S. Euskirchen, T. Friborg, H. Genet, M. Göckede, J. P. Goodrich, P. Grogan, M. Helbig, E. E. Jafarov, J. D. Jastrow, A. A. M. Kalhori, Y. Kim, J. S. Kimball, L. Kutzbach, M. J. Lara, K. S. Larsen, B.-Y. Lee, Z. Liu, M. M. Lorant, M. Lund, M. Lupascu, N. Madani, A. Malhotra, R. Matamala, J. McFarland, A. D. McGuire, A. Michelsen, C. Minions, W. C. Oechel, D. Olefeldt, F.-J. W. Parmentier, N. Pirk, B. Poulter, W. Quinton, F. Rezanezhad, D. Risk, T. Sachs, K. Schaefer, N. M. Schmidt, E. A. G. Schuur, P. R. Semenchuk, G. Shaver, O. Sonnentag, G. Starr, C. C. Treat, M. P. Waldrop, Y. Wang, J. Welker, C. Wille, X. Xu, Z. Zhang, Q. Zhuang, and D. Zona. Large loss of CO₂ in winter observed across the northern permafrost region. *Nature Climate Change*, 9(11):852–857, Nov. 2019. ISSN 1758-6798. doi: 10.1038/s41558-019-0592-8. URL <https://www.nature.com/articles/s41558-019-0592-8>.

C.-L. Navier. Sur les lois des mouvements des fluides, en ayant égard à l'adhésion des molécules. XIX:244–260, 1821.

C.-L. Navier. Sur les Lois du Mouvement des Fluides. VI:389–440, 1827.

R. a. J. Neggers. Attributing the behavior of low-level clouds in large-scale models to subgrid-scale parameterizations. *Journal of Advances in Modeling Earth Systems*, 7(4): 2029–2043, 2015. ISSN 1942-2466. doi: 10.1002/2015MS000503. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015MS000503>.

R. a. J. Neggers, J. Chylik, U. Egerer, H. Griesche, V. Schemann, P. Seifert, H. Siebert, and A. Macke. Local and Remote Controls on Arctic Mixed-Layer Evolution. *Journal of Advances in Modeling Earth Systems*, 11(7):2214–2237, 2019. ISSN 1942-2466. doi: 10.1029/2019MS001671. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019MS001671>.

R. A. J. Neggers, J. Chylik, and N. Schnierstein. The Entrainment Efficiency of Persistent Arctic Mixed-Phase Clouds as Inferred from Daily Large-Eddy Simulations during the MOSAiC Drift. *Journal of the Atmospheric Sciences*, 82(6):1195–1213, June 2025. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-24-0188.1. URL <https://journals.ametsoc.org/view/journals/atsc/82/6/JAS-D-24-0188.1.xml>.

F. T. M. Nieuwstadt and R. A. Brost. The Decay of Convective Turbulence. *Journal of the Atmospheric Sciences*, 43(6):532–546, Mar. 1986. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(1986)043<0532:TDOCT>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/43/6/1520-0469_1986_043_0532_tdoct_2_0_co_2.xml.

J. M. Novy. Effects of Arctic Amplification on Alaska Native Communities and Ecosystems. In *The Palgrave Handbook of Ecosystems and Wellbeing*, pages 1–6. Palgrave Macmillan, Cham, 2025. ISBN 9783031249525. doi: 10.1007/978-3-031-24952-5_4-1. URL https://link.springer.com/rwe/10.1007/978-3-031-24952-5_4-1.

NSF NCAR Earth Observing Laboratory. NCAR Airborne Vertical Atmospheric Profiling System (AVAPS). 1993. doi: 10.5065/D66W9848. URL <https://doi.org/10.5065/D66W9848>.

P. A. O’Gorman and C. Muller. How closely do changes in surface and column water vapor follow Clausius–Clapeyron scaling in climate change simulations? *Environmental Research Letters*, 5, 2010. doi: 10.1088/1748-9326/5/2/025207. URL <https://consensus.app/papers/how-closely-do-changes-in-surface-and-column-water-vapor-ogorman-muller/bf58f84b274a582aac8202d2d618cbca/>.

Y. Ogura and N. A. Phillips. Scale Analysis of Deep and Shallow Convection in the Atmosphere. *Journal of the Atmospheric Sciences*, 19(2):173–179, Mar. 1962. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(1962)019<0173:SAODAS>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/19/2/1520-0469_1962_019_0173_saodas_2_0_co_2.xml.

J. Olsen, M. Nenasheva, and G. K. Hovelsrud. ‘Road of life’: changing navigation seasons and the adaptation of island communities in the Russian Arctic. *Polar Geography*, 44(1): 1–19, Jan. 2021. ISSN 1088-937X. doi: 10.1080/1088937X.2020.1826593. URL <https://doi.org/10.1080/1088937X.2020.1826593>.

J. E. Overland, B.-M. Kim, and Y. Tachibana. Communicating Arctic-midlatitude weather and ecosystem connections: direct observations and sources of intermittency. *Environmental Research Letters*, 16(10):105006, Sept. 2021. ISSN 1748-9326. doi: 10.1088/1748-9326/ac25bc. URL <https://doi.org/10.1088/1748-9326/ac25bc>.

L. Papritz. Arctic Lower-Tropospheric Warm and Cold Extremes: Horizontal and Vertical Transport, Diabatic Processes, and Linkage to Synoptic Circulation Features. *Journal of Climate*, 33(3):993–1016, Jan. 2020. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-19-0638.1. URL <https://journals.ametsoc.org/view/journals/clim/33/3/jcli-d-19-0638.1.xml>.

L. Papritz and C. M. Grams. Linking Low-Frequency Large-Scale Circulation Patterns to Cold Air Outbreak Formation in the Northeastern North Atlantic. *Geophysical Research Letters*, 45(5):2542–2553, 2018. ISSN 1944-8007. doi: 10.1002/2017GL076921. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2017GL076921>.

L. Papritz and T. Spengler. A Lagrangian Climatology of Wintertime Cold Air Outbreaks in the Irminger and Nordic Seas and Their Role in Shaping Air–Sea Heat Fluxes. *Journal*

of *Climate*, 30(8):2717–2737, Apr. 2017. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-16-0605.1. URL <https://journals.ametsoc.org/view/journals/clim/30/8/jcli-d-16-0605.1.xml>.

L. Papritz, S. Pfahl, H. Sodemann, and H. Wernli. A Climatology of Cold Air Outbreaks and Their Impact on Air–Sea Heat Fluxes in the High-Latitude South Pacific. *Journal of Climate*, 28(1):342–364, Jan. 2015. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-14-00482.1. URL <https://journals.ametsoc.org/view/journals/clim/28/1/jcli-d-14-00482.1.xml>.

L. Papritz, E. Rouges, F. Aemisegger, and H. Wernli. On the Thermodynamic Preconditioning of Arctic Air Masses and the Role of Tropopause Polar Vortices for Cold Air Outbreaks From Fram Strait. *Journal of Geophysical Research: Atmospheres*, 124(21): 11033–11050, 2019. ISSN 2169-8996. doi: 10.1029/2019JD030570. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019JD030570>.

L. Papritz, S. Murto, M. Röthlisberger, R. Caballero, G. Messori, G. Svensson, and H. Wernli. The Role of Local and Remote Processes for Wintertime Surface Energy Budget Extremes over Arctic Sea Ice. *Journal of Climate*, 36(21):7657–7674, Oct. 2023. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-22-0883.1. URL <https://journals.ametsoc.org/view/journals/clim/36/21/JCLI-D-22-0883.1.xml>.

V. K. Patel and J. Kuttippurath. Increase in Tropospheric Water Vapor Amplifies Global Warming and Climate Change. *Ocean-Land-Atmosphere Research*, 2:0015, July 2023. doi: 10.34133/olar.0015. URL <https://spj.science.org/doi/full/10.34133/olar.0015>.

F. Paulus. Impact of Cloud Condensation Nuclei on the Evolution of Clouds in Marine Cold-Air Outbreaks [Model configuration and output], Mar. 2026a. URL <https://zenodo.org/records/19337964>.

F. Paulus. Impacts of mesoscale atmospheric subsidence on cloud glaciation and decoupling in Arctic marine cold air outbreaks [Model configuration and output], Jan. 2026b. URL <https://zenodo.org/records/18232592>.

F. Paulus and M. Karalis. Airborne measurements of mesoscale divergence at high latitudes during HALO-(AC)³ [Code publication], Dec. 2023. URL <https://zenodo.org/records/10402061>.

F. M. Paulus, M. Karalis, G. George, G. Svensson, M. Wendisch, and R. A. J. Neggers. Airborne Measurements of Mesoscale Divergence at High Latitudes during HALO-(AC)³. *Journal of Atmospheric Sciences*, Nov. 2024. doi: 10.1175/JAS-D-24-0034.1. URL <https://journals.ametsoc.org/view/journals/atsc/81/12/JAS-D-24-0034.1.xml>.
Section: Journal of the Atmospheric Sciences.

F. M. Paulus, J. J. Müller, B. Kirbus, H. Sodemann, L. van Gelder, A. Walbröl, M. Wendisch, and R. A. J. Neggers. Impacts of mesoscale atmospheric subsidence on cloud glaciation and decoupling in Arctic marine cold air outbreaks. *EGUsphere*, pages 1–41, Feb. 2026. doi: 10.5194/egusphere-2026-554. URL <https://egusphere.copernicus.org/preprints/2026/egusphere-2026-554/>.

R. A. Pedersen, I. Cvijanovic, P. L. Langen, and B. M. Vinther. The Impact of Regional Arctic Sea Ice Loss on Atmospheric Circulation and the NAO. *Journal of Climate*, 29(2): 889–902, Jan. 2016. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0315.1. URL <https://journals.ametsoc.org/view/journals/clim/29/2/jcli-d-15-0315.1.xml>.

D. K. Perovich, S. V. Nghiem, T. Markus, and A. Schweiger. Seasonal evolution and inter-annual variability of the local solar energy absorbed by the Arctic sea ice–ocean system. *Journal of Geophysical Research: Oceans*, 112(C3), 2007. ISSN 2156-2202. doi: 10.1029/2006JC003558. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2006JC003558>.

P. O. G. Persson, M. D. Shupe, D. Perovich, and A. Solomon. Linking atmospheric synoptic transport, cloud phase, surface energy fluxes, and sea-ice growth: observations of midwinter SHEBA conditions. *Climate Dynamics*, 49(4):1341–1364, Aug. 2017. ISSN 1432-0894. doi: 10.1007/s00382-016-3383-1. URL <https://doi.org/10.1007/s00382-016-3383-1>.

D. Philipp, M. Stengel, and B. Ahrens. Analyzing the Arctic Feedback Mechanism between Sea Ice and Low-Level Clouds Using 34 Years of Satellite Observations. *Journal of Climate*, 33(17):7479–7501, July 2020. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-19-0895.1. URL <https://journals.ametsoc.org/view/journals/clim/33/17/jcliD190895.xml>.

G. K. Phoenix, J. W. Bjerke, R. G. Björk, D. Blok, A. Bryn, T. V. Callaghan, C. T. Christiansen, A. M. Cunliffe, S. J. Davidson, H. E. Epstein, M. M. Lorantý, A. C. Martin, I. H. Myers-Smith, J. Olofsson, T. C. Parker, F.-J. W. Parmentier, F. Stordal, R. Treharne, H. Tømmervik, and C. Voigt. Browning events in Arctic ecosystems: Diverse causes with common consequences. *PLOS Climate*, 4(1):e0000570, Jan. 2025. ISSN 2767-3200. doi: 10.1371/journal.pclm.0000570. URL <https://journals.plos.org/climate/article?id=10.1371/journal.pclm.0000570>.

R. Pincus and B. Stevens. Monte Carlo Spectral Integration: a Consistent Approximation for Radiative Transfer in Large Eddy Simulations. *Journal of Advances in Modeling Earth Systems*, 1(2), 2009. ISSN 1942-2466. doi: 10.3894/JAMES.2009.1.1. URL <https://onlinelibrary.wiley.com/doi/abs/10.3894/JAMES.2009.1.1>.

F. Pithan and T. Mauritsen. Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nature Geoscience*, 7(3):181–184, Mar. 2014. ISSN 1752-0908. doi: 10.1038/ngeo2071. URL <https://www.nature.com/articles/ngeo2071>.

F. Pithan, G. Svensson, R. Caballero, D. Chechin, T. W. Cronin, A. M. L. Ekman, R. Neggers, M. D. Shupe, A. Solomon, M. Tjernström, and M. Wendisch. Role of air-mass transformations in exchange between the Arctic and mid-latitudes. *Nature Geoscience*, 11(11): 805–812, Nov. 2018. ISSN 1752-0908. doi: 10.1038/s41561-018-0234-1. URL <https://www.nature.com/articles/s41561-018-0234-1>.

M. Planck. *The Theory of Heat Radiation*. P. Blakiston's Son & Co., Philadelphia, 1914. URL <https://www.gutenberg.org/ebooks/40030>.

S. B. Pope. Turbulent Flows. *Measurement Science and Technology*, 12(11):2020, Nov. 2001. ISSN 0957-0233. doi: 10.1088/0957-0233/12/11/705. URL <https://doi.org/10.1088/0957-0233/12/11/705>.

B. Poujol and S. Bony. Measuring Clear-Air Vertical Motions From Space. *AGU Advances*, 5(4):e2024AV001267, 2024. ISSN 2576-604X. doi: 10.1029/2024AV001267. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024AV001267>.

M. Previdi, T. Janoski, G. Chiodo, K. L. Smith, and L. M. Polvani. Arctic Amplification: A Rapid Response to Radiative Forcing. *Geophysical Research Letters*, 2020. doi: 10.1029/2020GL089933. URL <https://doi.org/10.1029/2020GL089933>.

M. Previdi, K. L. Smith, and L. M. Polvani. Arctic amplification of climate change: a review of underlying mechanisms. *Environmental Research Letters*, 16(9):093003, Sept. 2021. ISSN 1748-9326. doi: 10.1088/1748-9326/ac1c29. URL <https://doi.org/10.1088/1748-9326/ac1c29>.

H. R. Pruppacher and J. D. Klett. *Microphysics of Clouds and Precipitation: Reprinted 1980*. Springer Science & Business Media, Dec. 2012. ISBN 9789400999053.

S. Raasch. Numerical simulation of the development of the convective boundary layer during a cold air outbreak. *Boundary-Layer Meteorology*, 52(4):349–375, Sept. 1990. ISSN 1573-1472. doi: 10.1007/BF00119429. URL <https://doi.org/10.1007/BF00119429>.

D. A. Randall. *Quick Studies in Atmospheric Science*, volume The Anelastic and Boussinesq Approximations. May 2013. URL <https://hogback.atmos.colostate.edu/group/dave/pdf/AneBous.pdf>.

M. Rantanen, A. Y. Karpechko, A. Lipponen, K. Nordling, O. Hyvärinen, K. Ruosteenoja, T. Vihma, and A. Laaksonen. The Arctic has warmed nearly four times faster than the globe since 1979. *Communications Earth & Environment*, 3(1):168, Aug. 2022. ISSN 2662-4435. doi: 10.1038/s43247-022-00498-3. URL <https://www.nature.com/articles/s43247-022-00498-3>.

E. A. Rasmussen. Polar Lows. In R. H. Johnson and R. A. Houze, editors, *A Half Century of Progress in Meteorology: A Tribute to Richard Reed*, pages 61–78. American Meteorological Society, Boston, MA, 2003. ISBN 9781878220691. doi: 10.1007/978-1-878220-69-1_4. URL https://doi.org/10.1007/978-1-878220-69-1_4.

E. A. Rasmussen, T. S. Pedersen, L. T. Pedersen, and J. Turner. Polar lows and arctic instability lows in the Bear Island region. *Tellus A*, 44(2):133–154, 1992. ISSN 1600-0870. doi: 10.1034/j.1600-0870.1992.t01-1-00004.x. URL <https://onlinelibrary.wiley.com/doi/abs/10.1034/j.1600-0870.1992.t01-1-00004.x>.

R. Rauterkus and C. Ansorge. Cloud-Top Entrainment in Mixed-Phase Stratocumulus and Its Process-Level Representation in Large-Eddy Simulation. *Journal of the Atmospheric Sciences*, 77(12):4109–4127, Dec. 2020. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-19-0221.1. URL <https://journals.ametsoc.org/view/journals/atsc/77/12/JAS-D-19-0221.1.xml>.

S. Reilly, S. D. Gesso, and R. Neggers. Configuring LES Based on Dropsonde Data in Sparsely Sampled Areas in the Subtropical Atlantic. Feb. 2020. doi: 10.1175/JAMC-D-19-0013.1. URL <https://journals.ametsoc.org/view/journals/apme/59/2/jamc-d-19-0013.1.xml>. Section: Journal of Applied Meteorology and Climatology.

reIMAGINE Arctic Research. reIMAGINE Arctic Research | Research Institute for Sustainability, Jan. 2017. URL <https://www.rifs-potsdam.de/en/research/global-change-and-arctic-sustainable-transformations-glocast>.

I. A. Renfrew and G. W. K. Moore. An Extreme Cold-Air Outbreak over the Labrador Sea: Roll Vortices and Air–Sea Interaction. *Monthly Weather Review*, 127(10):2379–2394, Oct. 1999. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1999)127<2379:AECAOO>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/127/10/1520-0493_1999_127_2379_aecaoo_2.0.co_2.xml.

R. M. Robinson, J. Scheff, and N. Golden. CMIP6 captures the satellite-era jet slowdown and Arctic amplification, yet projects future jet speedup and tropical amplification. *Climate Dynamics*, 61(9):4915–4926, Nov. 2023. ISSN 1432-0894. doi: 10.1007/s00382-023-06839-y. URL <https://doi.org/10.1007/s00382-023-06839-y>.

S. R. d. Roode, I. Sandu, J. J. v. d. Dussen, A. S. Ackerman, P. Blossey, D. Jarecka, A. Lock, A. P. Siebesma, and B. Stevens. Large-Eddy Simulations of EUCLIPSE–GASS Lagrangian Stratocumulus-to-Cumulus Transitions: Mean State, Turbulence, and Decoupling. *Journal of the Atmospheric Sciences*, 73(6):2485–2508, June 2016. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-15-0215.1. URL <https://journals.ametsoc.org/view/journals/atsc/73/6/jas-d-15-0215.1.xml>.

R. D. Russotto and M. Biasutti. Polar Amplification as an Inherent Response of a Circulating Atmosphere: Results From the TRACMIP Aquaplanets. *Geophysical Research Letters*, 47(6):e2019GL086771, 2020. ISSN 1944-8007. doi: 10.1029/2019GL086771. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019GL086771>.

Saami Council. Co-Production of Knowledge in the Arctic: Bridging Indigenous and Scientific Perspectives, Dec. 2025. URL <https://arctic-council.org/news/co-production-of-knowledge-in-the-arctic-bridging-indigenous-and-scientific-perspectives/>.

I. Sandu and B. Stevens. On the Factors Modulating the Stratocumulus to Cumulus Transitions. *Journal of the Atmospheric Sciences*, 68(9):1865–1881, Sept. 2011. ISSN 0022-4928, 1520-0469. doi: 10.1175/2011JAS3614.1. URL <https://journals.ametsoc.org/view/journals/atsc/68/9/2011jas3614.1.xml>.

I. Schirmacher, P. Kollias, K. Lamer, M. Mech, L. Pfizenmaier, M. Wendisch, and S. Crewell. Assessing Arctic low-level clouds and precipitation from above – a radar perspective. *Atmospheric Measurement Techniques*, 16(17):4081–4100, Sept. 2023. ISSN 1867-1381. doi: 10.5194/amt-16-4081-2023. URL <https://amt.copernicus.org/articles/16/4081/2023/>.

I. Schirmacher, S. Schnitt, M. Klingebiel, N. Maherndl, B. Kirbus, A. Ehrlich, M. Mech, and S. Crewell. Clouds and precipitation in the initial phase of marine cold-air outbreaks as observed by airborne remote sensing. *Atmospheric Chemistry and Physics*, 24(22):12823–12842, Nov. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-12823-2024. URL <https://acp.copernicus.org/articles/24/12823/2024/>.

T. Schneider von Deimling, H. Lee, T. Ingeman-Nielsen, S. Westermann, V. Romanovsky, S. Lamoureux, D. A. Walker, S. Chadburn, E. Trochim, L. Cai, J. Nitzbon, S. Jacobi, and M. Langer. Consequences of permafrost degradation for Arctic infrastructure – bridging the model gap between regional and engineering scales. *The Cryosphere*, 15(5):2451–2471, May 2021. ISSN 1994-0416. doi: 10.5194/tc-15-2451-2021. URL <https://tc.copernicus.org/articles/15/2451/2021/>.

N. Schnierstein, J. Chylik, M. D. Shupe, and R. a. J. Neggers. Standardized Daily High-Resolution Large-Eddy Simulations of the Arctic Boundary Layer and Clouds During the Complete MOSAiC Drift. *Journal of Advances in Modeling Earth Systems*, 16(11):e2024MS004296, 2024. ISSN 1942-2466. doi: 10.1029/2024MS004296. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024MS004296>.

M. Schäfer, K. Wolf, A. Ehrlich, C. Hallbauer, E. Jäkel, F. Jansen, A. E. Luebke, J. Müller, J. Thoböll, T. Rösenthaller, B. Stevens, and M. Wendisch. VELOX – a new thermal infrared imager for airborne remote sensing of cloud and surface properties. *Atmospheric*

Measurement Techniques, 15(5):1491–1509, Mar. 2022. ISSN 1867-1381. doi: 10.5194/amt-15-1491-2022. URL <https://amt.copernicus.org/articles/15/1491/2022/>. Publisher: Copernicus GmbH.

J. A. Screen. Simulated Atmospheric Response to Regional and Pan-Arctic Sea Ice Loss. *Journal of Climate*, 30(11):3945–3962, June 2017. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-16-0197.1. URL <https://journals.ametsoc.org/view/journals/clim/30/11/jcli-d-16-0197.1.xml>.

J. A. Screen and I. Simmonds. The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature*, 464(7293):1334–1337, Apr. 2010a. ISSN 1476-4687. doi: 10.1038/nature09051. URL <https://www.nature.com/articles/nature09051>.

J. A. Screen and I. Simmonds. Increasing fall-winter energy loss from the Arctic Ocean and its role in Arctic temperature amplification. *Geophysical Research Letters*, 37(16), 2010b. ISSN 1944-8007. doi: 10.1029/2010GL044136. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2010GL044136>.

J. A. Screen, R. Eade, D. M. Smith, S. Thomson, and H. Yu. Net Equatorward Shift of the Jet Streams When the Contribution From Sea-Ice Loss Is Constrained by Observed Eddy Feedback. *Geophysical Research Letters*, 49(23):e2022GL100523, 2022. ISSN 1944-8007. doi: 10.1029/2022GL100523. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2022GL100523>.

C. Seethala, P. Zuidema, J. Edson, M. Brunke, G. Chen, X.-Y. Li, D. Painemal, C. Robinson, T. Shingler, M. Shook, A. Sorooshian, L. Thornhill, F. Tornow, H. Wang, X. Zeng, and L. Ziemba. On Assessing ERA5 and MERRA2 Representations of Cold-Air Outbreaks Across the Gulf Stream. *Geophysical Research Letters*, 48(19):e2021GL094364, 2021. ISSN 1944-8007. doi: 10.1029/2021GL094364. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021GL094364>.

D. J. Seidel, Y. Zhang, A. Beljaars, J.-C. Golaz, A. R. Jacobson, and B. Medeiros. Climatology of the planetary boundary layer over the continental United States and Europe. *Journal of Geophysical Research: Atmospheres*, 117(D17), 2012. ISSN 2156-2202. doi: 10.1029/2012JD018143. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2012JD018143>.

J. S. Seidel, A. A. Kiselev, A. Keinert, F. Stratmann, T. Leisner, and S. Hartmann. Secondary ice production – no evidence of efficient rime-splintering mechanism. *Atmospheric Chemistry and Physics*, 24(9):5247–5263, May 2024. ISSN 1680-7316. doi: 10.5194/acp-24-5247-2024. URL <https://acp.copernicus.org/articles/24/5247/2024/>.

A. Seifert and K. D. Beheng. A two-moment cloud microphysics parameterization for mixed-phase clouds. Part 2: Maritime vs. continental deep convective storms. *Meteorology and Atmospheric Physics*, 92(1):67–82, Feb. 2006. ISSN 1436-5065. doi: 10.1007/s00703-005-0113-3. URL <https://doi.org/10.1007/s00703-005-0113-3>.

W. D. Sellers. A Global Climatic Model Based on the Energy Balance of the Earth-Atmosphere System. *Journal of Applied Meteorology and Climatology*, 8(3):392–400, June 1969. ISSN 1520-0450. doi: 10.1175/1520-0450(1969)008<0392:AGCMBO>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/apme/8/3/1520-0450_1969_008_0392_agcmbo_2_0_co_2.xml.

M. C. Serreze and J. A. Francis. The Arctic Amplification Debate. *Climatic Change*, 2006. doi: 10.1007/S10584-005-9017-Y. URL <https://doi.org/10.1007/s10584-005-9017-y>.

M. C. Serreze, A. P. Barrett, J. C. Stroeve, D. N. Kindig, and M. M. Holland. The emergence of surface-based Arctic amplification. *The Cryosphere*, 3(1):11–19, Feb. 2009. ISSN 1994-0416. doi: 10.5194/tc-3-11-2009. URL <https://tc.copernicus.org/articles/3/11/2009/>.

M. A. Shapiro, L. S. Fedor, and T. Hampel. Research aircraft measurements of a polar low over the Norwegian Sea. *Tellus*, 39(4), Jan. 1987a. ISSN 3035-9554. doi: 10.3402/tellusa.v39i4.11761. URL <https://tellusjournal.org/articles/10.3402/tellusa.v39i4.11761>.

M. A. Shapiro, T. Hampel, and A. J. Krueger. The Arctic Tropopause Fold. *Monthly Weather Review*, 115(2):444–454, Feb. 1987b. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1987)115<0444:TATF>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/115/2/1520-0493_1987_115_0444_tatf_2_0_co_2.xml.

S. C. Sherwood, M. J. Webb, J. D. Annan, K. C. Armour, P. M. Forster, J. C. Hargreaves, G. Hegerl, S. A. Klein, K. D. Marvel, E. J. Rohling, M. Watanabe, T. Andrews, P. Braconnot, C. S. Bretherton, G. L. Foster, Z. Hausfather, A. S. von der Heydt, R. Knutti, T. Mauritsen, J. R. Norris, C. Proistosescu, M. Rugenstein, G. A. Schmidt, K. B. Tokarska, and M. D. Zelinka. An Assessment of Earth’s Climate Sensitivity Using Multiple Lines of Evidence. *Reviews of Geophysics*, 58(4):e2019RG000678, 2020. ISSN 1944-9208. doi: 10.1029/2019RG000678. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019RG000678>.

A. A. Shestakova, D. G. Chechin, C. Lüpkes, J. Hartmann, and M. Maturilli. The foehn effect during easterly flow over Svalbard. *Atmospheric Chemistry and Physics*, 22(2):1529–1548, Feb. 2022. ISSN 1680-7316. doi: 10.5194/acp-22-1529-2022. URL <https://acp.copernicus.org/articles/22/1529/2022/>.

M. D. Shupe and J. M. Intrieri. Cloud Radiative Forcing of the Arctic Surface: The Influence of Cloud Properties, Surface Albedo, and Solar Zenith Angle. *Journal of Climate*, 17(3):616–628, Feb. 2004. ISSN 0894-8755, 1520-0442. doi: 10.1175/1520-0442(2004)017<0616:CRFOTA>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/clim/17/3/1520-0442_2004_017_0616_crfota_2.0.co_2.xml.

M. D. Shupe, P. O. G. Persson, I. M. Brooks, M. Tjernström, J. Sedlar, T. Mauritsen, S. Sjogren, and C. Leck. Cloud and boundary layer interactions over the Arctic sea ice in late summer. *Atmospheric Chemistry and Physics*, 13(18):9379–9399, Sept. 2013. ISSN 1680-7316. doi: 10.5194/acp-13-9379-2013. URL <https://acp.copernicus.org/articles/13/9379/2013/>.

M. D. Shupe, M. Rex, B. Blomquist, P. O. G. Persson, J. Schmale, T. Uttal, D. Althausen, H. Angot, S. Archer, L. Bariteau, I. Beck, J. Bilberry, S. Bucci, C. Buck, M. Boyer, Z. Brasseur, I. M. Brooks, R. Calmer, J. Cassano, V. Castro, D. Chu, D. Costa, C. J. Cox, J. Creamean, S. Crewell, S. Dahlke, E. Damm, G. de Boer, H. Deckelmann, K. Dethloff, M. Dütsch, K. Ebell, A. Ehrlich, J. Ellis, R. Engelmann, A. A. Fong, M. M. Frey, M. R. Gallagher, L. Ganzeveld, R. Gradinger, J. Graeser, V. Greenamyre, H. Griesche, S. Griffiths, J. Hamilton, G. Heinemann, D. Helmig, A. Herber, C. Heuzé, J. Hofer, T. Houchens, D. Howard, J. Inoue, H.-W. Jacobi, R. Jaiser, T. Jokinen, O. Jourdan, G. Jozef, W. King, A. Kirchgaessner, M. Klingebiel, M. Krassovski, T. Krumpen, A. Lampert, W. Landing, T. Laurila, D. Lawrence, M. Lonardi, B. Loose, C. Lüpkes, M. Maahn, A. Macke, W. Maslowski, C. Marsay, M. Maturilli, M. Mech, S. Morris, M. Moser, M. Nicolaus, P. Ortega, J. Osborn, F. Pätzold, D. K. Perovich, T. Petäjä, C. Pilz, R. Pirazzini, K. Posman, H. Powers, K. A. Pratt, A. Preußner, L. Quéléver, M. Radenz, B. Rabe, A. Rinke, T. Sachs, A. Schulz, H. Siebert, T. Silva, A. Solomon, A. Sommerfeld, G. Spreen, M. Stephens, A. Stohl, G. Svensson, J. Uin, J. Viegas, C. Voigt, P. von der Gathen, B. Wehner, J. M. Welker, M. Wendisch, M. Werner, Z. Xie, and F. Yue. Overview of the MOSAiC expedition: Atmosphere. *Elementa: Science of the Anthropocene*, 10(1):00060, Feb. 2022. ISSN 2325-1026. doi: 10.1525/elementa.2021.00060. URL <https://doi.org/10.1525/elementa.2021.00060>.

E. D. Skyllingstad and J. B. Edson. Large-Eddy Simulation of Moist Convection during a Cold Air Outbreak over the Gulf Stream. *Journal of the Atmospheric Sciences*, 66(5):1274–1293, May 2009. ISSN 0022-4928, 1520-0469. doi: 10.1175/2008JAS2755.1. URL <https://journals.ametsoc.org/view/journals/atsc/66/5/2008jas2755.1.xml>.

N. Slättberg, S. Dahlke, and M. Maturilli. Fram Strait Marine Cold Air Outbreaks in CARRA and ERA5: Effects on Surface Turbulent Heat Fluxes and the Vertical Structure of the Troposphere. *Journal of Geophysical Research: Atmospheres*, 130(12):e2024JD042908, 2025. ISSN 2169-8996. doi: 10.1029/2024JD042908. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024JD042908>.

J. Smagorinsky. General Circulation Experiments with the primitive equations: I. The basic experiment. *Monthly Weather Review*, 91(3):99–164, Mar. 1963. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1963)091<0099:GCEWTP>2.3.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/91/3/1520-0493_1963_091_0099_gcewtp_2_3_co_2.xml.

E. T. Smith and S. C. Sheridan. The characteristics of extreme cold events and cold air outbreaks in the eastern United States. *International Journal of Climatology*, 38(S1):e807–e820, 2018. ISSN 1097-0088. doi: 10.1002/joc.5408. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/joc.5408>.

E. T. Smith and S. C. Sheridan. Where Do Cold Air Outbreaks Occur, and How Have They Changed Over Time? *Geophysical Research Letters*, 47(13):e2020GL086983, 2020. ISSN 1944-8007. doi: 10.1029/2020GL086983. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2020GL086983>.

A. Solomon, H. Morrison, O. Persson, M. D. Shupe, and J.-W. Bao. Investigation of Microphysical Parameterizations of Snow and Ice in Arctic Clouds during M-PACE through Model–Observation Comparisons. *Monthly Weather Review*, 137(9):3110–3128, Sept. 2009. ISSN 1520-0493, 0027-0644. doi: 10.1175/2009MWR2688.1. URL <https://journals.ametsoc.org/view/journals/mwre/137/9/2009mwr2688.1.xml>.

A. Solomon, M. D. Shupe, P. O. G. Persson, and H. Morrison. Moisture and dynamical interactions maintaining decoupled Arctic mixed-phase stratocumulus in the presence of a humidity inversion. *Atmospheric Chemistry and Physics*, 11(19):10127–10148, Oct. 2011. ISSN 1680-7316. doi: 10.5194/acp-11-10127-2011. URL <https://acp.copernicus.org/articles/11/10127/2011/>.

A. Solomon, G. de Boer, J. M. Creamean, A. McComiskey, M. D. Shupe, M. Maahn, and C. Cox. The relative impact of cloud condensation nuclei and ice nucleating particle concentrations on phase partitioning in Arctic mixed-phase stratocumulus clouds. *Atmospheric Chemistry and Physics*, 18(23):17047–17059, Dec. 2018. ISSN 1680-7316. doi: 10.5194/acp-18-17047-2018. URL <https://acp.copernicus.org/articles/18/17047/2018/>.

G. Spreen, L. Kaleschke, and G. Heygster. Sea ice remote sensing using AMSR-E 89-GHz channels. *Journal of Geophysical Research: Oceans*, 113(C2), 2008. ISSN 2156-2202. doi: 10.1029/2005JC003384. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2005JC003384>.

M. Sprenger and H. Wernli. The LAGRANTO Lagrangian analysis tool – version 2.0. *Geoscientific Model Development*, 8(8):2569–2586, Aug. 2015. ISSN 1991-959X. doi: 10.5194/gmd-8-2569-2015. URL <https://gmd.copernicus.org/articles/8/2569/2015/>. Publisher: Copernicus GmbH.

I. S. Stachlewska, R. Neuber, A. Lampert, C. Ritter, and G. Wehrle. AMALi – the Airborne Mobile Aerosol Lidar for Arctic research. *Atmospheric Chemistry and Physics*, 10(6):2947–2963, Mar. 2010. ISSN 1680-7316. doi: 10.5194/acp-10-2947-2010. URL <https://acp.copernicus.org/articles/10/2947/2010/>.

L. J. Sterzinger and A. L. Igel. Above-cloud concentrations of cloud condensation nuclei help to sustain some Arctic low-level clouds. *Atmospheric Chemistry and Physics*, 24(6):3529–3540, Mar. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-3529-2024. URL <https://acp.copernicus.org/articles/24/3529/2024/>.

B. Stevens. On the Growth of Layers of Nonprecipitating Cumulus Convection. *Journal of the Atmospheric Sciences*, 64(8):2916–2931, Aug. 2007. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS3983.1. URL <https://journals.ametsoc.org/view/journals/atsc/64/8/jas3983.1.xml>.

B. Stevens and G. Feingold. Untangling aerosol effects on clouds and precipitation in a buffered system. *Nature*, 461(7264):607–613, Oct. 2009. ISSN 1476-4687. doi: 10.1038/nature08281. URL <https://www.nature.com/articles/nature08281>.

B. Stevens, A. S. Ackerman, B. A. Albrecht, A. R. Brown, A. Chlond, J. Cuxart, P. G. Duynkerke, D. C. Lewellen, M. K. Macvean, R. A. J. Neggers, E. Sánchez, A. P. Siebesma, and D. E. Stevens. Simulations of Trade Wind Cumuli under a Strong Inversion. *Journal of the Atmospheric Sciences*, 58(14):1870–1891, July 2001. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(2001)058<1870:SOTWCU>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/58/14/1520-0469_2001_058_1870_sotwcu_2.0.co_2.xml. Publisher: American Meteorological Society Section: Journal of the Atmospheric Sciences.

B. Stevens, F. Ament, S. Bony, S. Crewell, F. Ewald, S. Gross, A. Hansen, L. Hirsch, M. Jacob, T. Kölling, H. Konow, B. Mayer, M. Wendisch, M. Wirth, K. Wolf, S. Bakan, M. Bauer-Pfundstein, M. Brueck, J. Delanoë, A. Ehrlich, D. Farrell, M. Forde, F. Gödde, H. Grob, M. Hagen, E. Jäkel, F. Jansen, C. Klepp, M. Klingebiel, M. Mech, G. Peters, M. Rapp, A. A. Wing, and T. Zinner. A High-Altitude Long-Range Aircraft Configured as a Cloud Observatory: The NARVAL Expeditions. *Bulletin of the American Meteorological Society*, June 2019. doi: 10.1175/BAMS-D-18-0198.1. URL <https://journals.ametsoc.org/view/journals/bams/100/6/bams-d-18-0198.1.xml>. Section: Bulletin of the American Meteorological Society.

B. Stevens, S. Bony, D. Farrell, F. Ament, A. Blyth, C. Fairall, J. Karstensen, P. K. Quinn, S. Speich, C. Acquistapace, F. Aemisegger, A. L. Albright, H. Bellenger, E. Bodenschatz, K.-A. Caesar, R. Chewitt-Lucas, G. de Boer, J. Delanoë, L. Denby, F. Ewald, B. Fildier, M. Forde, G. George, S. Gross, M. Hagen, A. Hausold, K. J. Heywood, L. Hirsch, M. Jacob,

F. Jansen, S. Kinne, D. Klocke, T. Kölling, H. Konow, M. Lothon, W. Mohr, A. K. Naumann, L. Nuijens, L. Olivier, R. Pincus, M. Pöhlker, G. Reverdin, G. Roberts, S. Schnitt, H. Schulz, A. P. Siebesma, C. C. Stephan, P. Sullivan, L. Touzé-Peiffer, J. Vial, R. Vogel, P. Zuidema, N. Alexander, L. Alves, S. Arixi, H. Asmath, G. Bagheri, K. Baier, A. Bailey, D. Baranowski, A. Baron, S. Barrau, P. A. Barrett, F. Batier, A. Behrendt, A. Bendinger, F. Beucher, S. Bigorre, E. Blades, P. Blossey, O. Bock, S. Böing, P. Bosser, D. Bourras, P. Bouruet-Aubertot, K. Bower, P. Branellec, H. Branger, M. Brennek, A. Brewer, P.-E. Brilouet, B. Brüggmann, S. A. Buehler, E. Burke, R. Burton, R. Calmer, J.-C. Canonici, X. Carton, G. Cato Jr., J. A. Charles, P. Chazette, Y. Chen, M. T. Chilinski, T. Choullarton, P. Chuang, S. Clarke, H. Coe, C. Cornet, P. Coutris, F. Couvreur, S. Crewell, T. Cronin, Z. Cui, Y. Cuypers, A. Daley, G. M. Damerell, T. Dauhut, H. Deneke, J.-P. Desbios, S. Dörner, S. Donner, V. Douet, K. Drushka, M. Dütsch, A. Ehrlich, K. Emanuel, A. Emmanouilidis, J.-C. Etienne, S. Etienne-Leblanc, G. Faure, G. Feingold, L. Ferrero, A. Fix, C. Flamant, P. J. Flatau, G. R. Foltz, L. Forster, I. Furtuna, A. Gadian, J. Galewsky, M. Gallagher, P. Gallimore, C. Gaston, C. Gentemann, N. Geyskens, A. Giez, J. Gollop, I. Gouirand, C. Gourbeyre, D. de Graaf, G. E. de Groot, R. Grosz, J. Güttler, M. Gutleben, K. Hall, G. Harris, K. C. Helfer, D. Henze, C. Herbert, B. Holanda, A. Ibanez-Landeta, J. Intrieri, S. Iyer, F. Julien, H. Kalesse, J. Kazil, A. Kellman, A. T. Kidane, U. Kirchner, M. Klingebiel, M. Körner, L. A. Kremper, J. Kretzschmar, O. Krüger, W. Kumala, A. Kurz, P. L'Hégaret, M. Labaste, T. Lachlan-Cope, A. Laing, P. Landschützer, T. Lang, D. Lange, I. Lange, C. Laplace, G. Lavik, R. Laxenaire, C. Le Bihan, M. Leandro, N. Lefevre, M. Lena, D. Lenschow, Q. Li, G. Lloyd, S. Los, N. Losi, O. Lovell, C. Luneau, P. Makuch, S. Malinowski, G. Manta, E. Marinou, N. Marsden, S. Masson, N. Maury, B. Mayer, M. Mayers-Als, C. Mazel, W. McGeary, J. C. McWilliams, M. Mech, M. Mehlmann, A. N. Meroni, T. Mieslinger, A. Minikin, P. Minnett, G. Möller, Y. Morfa Avalos, C. Muller, I. Musat, A. Napoli, A. Neuberger, C. Noisel, D. Noone, F. Nordsiek, J. L. Nowak, L. Oswald, D. J. Parker, C. Peck, R. Person, M. Philippi, A. Plueddemann, C. Pöhlker, V. Pörtge, U. Pöschl, L. Pologne, M. Posyniak, M. Prange, E. Quiñones Meléndez, J. Radtke, K. Ramage, J. Reimann, L. Renault, K. Reus, A. Reyes, J. Ribbe, M. Ringel, M. Ritschel, C. B. Rocha, N. Rochetin, J. Röttenbacher, C. Rollo, H. Royer, P. Sadoulet, L. Saffin, S. Sandiford, I. Sandu, M. Schäfer, V. Schemann, I. Schirmacher, O. Schlenczek, J. Schmidt, M. Schröder, A. Schwarzenboeck, A. Sealy, C. J. Senff, I. Serikov, S. Shohan, E. Siddle, A. Smirnov, F. Späth, B. Spooner, M. K. Stolla, W. Szkółka, S. P. de Szoeki, S. Tarot, E. Tetoni, E. Thompson, J. Thomson, L. Tomassini, J. Totems, A. A. Ubele, L. Villiger, J. von Arx, T. Wagner, A. Walther, B. Webber, M. Wendisch, S. Whitehall, A. Wiltshire, A. A. Wing, M. Wirth, J. Wiskandt, K. Wolf, L. Worbes, E. Wright, V. Wulfmeyer, S. Young, C. Zhang, D. Zhang, F. Ziemann, T. Zinner, and M. Zöger. EUREC⁴A. *Earth System Science Data*, 13(8):4067–4119, Aug. 2021. ISSN 1866-3508. doi: 10.5194/essd-13-4067-2021. URL <https://essd.copernicus.org/articles/13/4067/2021/>. Publisher: Copernicus GmbH.

R. G. Stevens, K. Loewe, C. Dearden, A. Dimitrelos, A. Possner, G. K. Eirund, T. Raatikainen, A. A. Hill, B. J. Shipway, J. Wilkinson, S. Romakkaniemi, J. Tonttila, A. Laaksonen, H. Korhonen, P. Connolly, U. Lohmann, C. Hoose, A. M. L. Ekman, K. S. Carslaw, and P. R. Field. A model intercomparison of CCN-limited tenuous clouds in the high Arctic. *Atmospheric Chemistry and Physics*, 18(15):11041–11071, Aug. 2018. ISSN 1680-7316. doi: 10.5194/acp-18-11041-2018. URL <https://acp.copernicus.org/articles/18/11041/2018/>.

A. Stohl, E. Andrews, J. F. Burkhardt, C. Forster, A. Herber, S. W. Hoch, D. Kowal, C. Lunder, T. Mefford, J. A. Ogren, S. Sharma, N. Spichtinger, K. Stebel, R. Stone, J. Ström, K. Tørseth, C. Wehrli, and K. E. Yttri. Pan-Arctic enhancements of light absorbing aerosol concentrations due to North American boreal forest fires during summer 2004. *Journal of Geophysical Research: Atmospheres*, 111(D22), 2006. ISSN 2156-2202. doi: 10.1029/2006JD007216. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2006JD007216>.

G. Stokes. On the Theories of the Internal Friction of Fluids in Motion and of the Equilibrium and Motion of Elastic Solids. 8:287–391, 1845.

P. J. Stoll, R. G. Graversen, G. Noer, and K. Hodges. An objective global climatology of polar lows based on reanalysis data. *Quarterly Journal of the Royal Meteorological Society*, 144(716):2099–2117, Oct. 2018. ISSN 1477-870X. doi: 10.1002/qj.3309. URL <https://rmets.onlinelibrary.wiley.com/doi/10.1002/qj.3309>.

K. Stramler, A. D. D. Genio, and W. B. Rossow. Synoptically Driven Arctic Winter States. *Journal of Climate*, 24(6):1747–1762, Mar. 2011. ISSN 0894-8755, 1520-0442. doi: 10.1175/2010JCLI3817.1. URL <https://journals.ametsoc.org/view/journals/clim/24/6/2010jcli3817.1.xml>.

R. B. Stull. *An Introduction to Boundary Layer Meteorology*. Springer Netherlands, Dordrecht, 1988. ISBN 978-90-277-2769-5 978-94-009-3027-8. doi: 10.1007/978-94-009-3027-8. URL <http://link.springer.com/10.1007/978-94-009-3027-8>.

G. Svensson, S. Murto, M. D. Shupe, F. Pithan, L. Magnusson, J. J. Day, J. D. Doyle, I. A. Renfrew, T. Spengler, and T. Vihma. Warm air intrusions reaching the MOSAiC expedition in April 2020—The YOPP targeted observing period (TOP). *Elementa: Science of the Anthropocene*, 11(1):00016, June 2023. ISSN 2325-1026. doi: 10.1525/elementa.2023.00016. URL <https://doi.org/10.1525/elementa.2023.00016>.

K. Svingen, A. Brakstad, K. Våge, W.-J. v. Appen, and L. Papritz. The Impact of Cold-Air Outbreaks and Oceanic Lateral Fluxes on Dense-Water Formation in the Greenland Sea from a 10-Year Moored Record (1999–2009). *Journal of Physical Oceanography*, 53(6):1499–1517, June 2023. ISSN 0022-3670, 1520-0485. doi: 10.1175/JPO-D-22-0160.1.

URL <https://journals.ametsoc.org/view/journals/phoc/53/6/JPO-D-22-0160.1.xml>.

A. J. Sweeney, Q. Fu, S. Po-Chedley, H. Wang, and M. Wang. Internal Variability Increased Arctic Amplification During 1980–2022. *Geophysical Research Letters*, 50(24):e2023GL106060, 2023. ISSN 1944-8007. doi: 10.1029/2023GL106060. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2023GL106060>.

I. Tan and T. Storelvmo. Evidence of Strong Contributions From Mixed-Phase Clouds to Arctic Climate Change. *Geophysical Research Letters*, 46(5):2894–2902, 2019. ISSN 1944-8007. doi: 10.1029/2018GL081871. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL081871>.

C. Tao, M. Zhang, and S. Xie. Cloud radiative effect dominates variabilities of surface energy budget in the dark Arctic. *Scientific Reports*, 15(1):2976, Jan. 2025. ISSN 2045-2322. doi: 10.1038/s41598-025-86322-2. URL <https://www.nature.com/articles/s41598-025-86322-2>.

P. C. Taylor, M. Cai, A. Hu, J. Meehl, W. Washington, and G. J. Zhang. A Decomposition of Feedback Contributions to Polar Warming Amplification. *Journal of Climate*, 26(18):7023–7043, Sept. 2013. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-12-00696.1. URL <https://journals.ametsoc.org/view/journals/clim/26/18/jcli-d-12-00696.1.xml>.

P. C. Taylor, R. C. Boeke, L. N. Boisvert, N. Feldl, M. Henry, Y. Huang, P. L. Langen, W. Liu, F. Pithan, S. A. Sejas, and I. Tan. Process Drivers, Inter-Model Spread, and the Path Forward: A Review of Amplified Arctic Warming. *Frontiers in Earth Science*, 9, Feb. 2022. ISSN 2296-6463. doi: 10.3389/feart.2021.758361. URL <https://www.frontiersin.org/journals/earth-science/articles/10.3389/feart.2021.758361/full>.

M. Tengö, E. S. Brondizio, T. Elmqvist, P. Malmer, and M. Spierenburg. Connecting Diverse Knowledge Systems for Enhanced Ecosystem Governance: The Multiple Evidence Base Approach. *AMBIO*, 43(5):579–591, Sept. 2014. ISSN 1654-7209. doi: 10.1007/s13280-014-0501-3. URL <https://doi.org/10.1007/s13280-014-0501-3>.

A. Terpstra, I. A. Renfrew, and D. E. Sergeev. Characteristics of Cold-Air Outbreak Events and Associated Polar Mesoscale Cyclogenesis over the North Atlantic Region. *Journal of Climate*, 34(11):4567–4584, June 2021. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-20-0595.1. URL <https://journals.ametsoc.org/view/journals/clim/34/11/JCLI-D-20-0595.1.xml>.

M. Tjernström, M. D. Shupe, I. M. Brooks, P. O. G. Persson, J. Prytherch, D. J. Salisbury, J. Sedlar, P. Achtert, B. J. Brooks, P. E. Johnston, G. Sotiropoulou, and D. Wolfe. Warm-air advection, air mass transformation and fog causes rapid ice melt. *Geophysical Research*

Letters, 42(13):5594–5602, 2015. ISSN 1944-8007. doi: 10.1002/2015GL064373. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2015GL064373>.

L. Tomassini, P. R. Field, R. Honnert, S. Malardel, R. McTaggart-Cowan, K. Saitou, A. T. Noda, and A. Seifert. The “Grey Zone” cold air outbreak global model intercomparison: A cross evaluation using large-eddy simulations. *Journal of Advances in Modeling Earth Systems*, 9(1):39–64, 2017. ISSN 1942-2466. doi: 10.1002/2016MS000822. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016MS000822>.

F. Tornow, A. S. Ackerman, and A. M. Fridlind. Preconditioning of overcast-to-broken cloud transitions by riming in marine cold air outbreaks. *Atmospheric Chemistry and Physics*, 21(15):12049–12067, Aug. 2021. ISSN 1680-7316. doi: 10.5194/acp-21-12049-2021. URL <https://acp.copernicus.org/articles/21/12049/2021/>.

F. Tornow, A. S. Ackerman, A. M. Fridlind, G. Tselioudis, B. Cairns, D. Painemal, and G. Elsaesser. On the Impact of a Dry Intrusion Driving Cloud-Regime Transitions in a Midlatitude Cold-Air Outbreak. *Journal of the Atmospheric Sciences*, 80(12):2881–2896, Dec. 2023. ISSN 0022-4928, 1520-0469. doi: 10.1175/JAS-D-23-0040.1. URL <https://journals.ametsoc.org/view/journals/atsc/80/12/JAS-D-23-0040.1.xml>.

I. B. Troen and L. Mahrt. A simple model of the atmospheric boundary layer; sensitivity to surface evaporation. *Boundary-Layer Meteorology*, 37(1):129–148, Oct. 1986. ISSN 1573-1472. doi: 10.1007/BF00122760. URL <https://doi.org/10.1007/BF00122760>.

S. Twomey. Pollution and the planetary albedo. *Atmospheric Environment (1967)*, 8(12):1251–1256, Dec. 1974. ISSN 0004-6981. doi: 10.1016/0004-6981(74)90004-3. URL <https://www.sciencedirect.com/science/article/pii/0004698174900043>.

T. Uttal, J. A. Curry, M. G. McPhee, D. K. Perovich, R. E. Moritz, J. A. Maslanik, P. S. Guest, H. L. Stern, J. A. Moore, R. Turenne, A. Heiberg, M. C. Serreze, D. P. Wylie, O. G. Persson, C. A. Paulson, C. Halle, J. H. Morison, P. A. Wheeler, A. Makshtas, H. Welch, M. D. Shupe, J. M. Intrieri, K. Stamnes, R. W. Lindsey, R. Pinkel, W. S. Pegau, T. P. Stanton, and T. C. Grenfeld. Surface Heat Budget of the Arctic Ocean. *Bulletin of the American Meteorological Society*, 83(2):255–276, Feb. 2002. ISSN 0003-0007, 1520-0477. doi: 10.1175/1520-0477(2002)083<0255:SHBOTA>2.3.CO;2. URL https://journals.ametsoc.org/view/journals/bams/83/2/1520-0477_2002_083_0255_shbota_2_3_co_2.xml.

Vaisala. NCAR Dropsonde NRD41. *B212680EN-A* © Vaisala 2023, 2023. URL <https://docs.vaisala.com/api/khub/documents/xFqIJ3AzmQb3A1csocKwLQ/content?Ft-Calling-App=ft%2Fturnkey-portal&Ft-Calling-App-Version=5.1.57#>.

J. J. van der Dussen, S. R. de Roode, A. S. Ackerman, P. N. Blossey, C. S. Bretherton, M. J. Kurowski, A. P. Lock, R. a. J. Neggers, I. Sandu, and A. P. Siebesma. The GASS/EU-

CLIPSE model intercomparison of the stratocumulus transition as observed during AS-TEX: LES results. *Journal of Advances in Modeling Earth Systems*, 5(3):483–499, 2013. ISSN 1942-2466. doi: 10.1002/jame.20033. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/jame.20033>.

J. J. van der Dussen, S. R. de Roode, and A. P. Siebesma. How large-scale subsidence affects stratocumulus transitions. *Atmospheric Chemistry and Physics*, 16(2):691–701, Jan. 2016. ISSN 1680-7316. doi: 10.5194/acp-16-691-2016. URL <https://acp.copernicus.org/articles/16/691/2016/>.

S. J. A. van der Linden, J. M. Edwards, C. C. van Heerwaarden, E. Vignon, C. Genton, I. Petenko, P. Baas, H. J. J. Jonker, and B. J. H. van de Wiel. Large-Eddy Simulations of the Steady Wintertime Antarctic Boundary Layer. *Boundary-Layer Meteorology*, 173(2):165–192, Nov. 2019. ISSN 1573-1472. doi: 10.1007/s10546-019-00461-4. URL <https://doi.org/10.1007/s10546-019-00461-4>.

C. C. van Heerwaarden, B. J. H. van Stratum, T. Heus, J. A. Gibbs, E. Fedorovich, and J. P. Mellado. MicroHH 1.0: a computational fluid dynamics code for direct numerical simulation and large-eddy simulation of atmospheric boundary layer flows. *Geoscientific Model Development*, 10(8):3145–3165, Aug. 2017. ISSN 1991-959X. doi: 10.5194/gmd-10-3145-2017. URL <https://gmd.copernicus.org/articles/10/3145/2017/>.

S. Vavrus. The Impact of Cloud Feedbacks on Arctic Climate under Greenhouse Forcing. *Journal of Climate*, 17(3):603–615, Feb. 2004. ISSN 0894-8755, 1520-0442. doi: 10.1175/1520-0442(2004)017<0603:TIOCFO>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/clim/17/3/1520-0442_2004_017_0603_tiocfo_2.0.co_2.xml.

S. Vavrus, D. Waliser, A. Schweiger, and J. Francis. Simulations of 20th and 21st century Arctic cloud amount in the global climate models assessed in the IPCC AR4. *Climate Dynamics*, 33(7):1099–1115, Dec. 2009. ISSN 1432-0894. doi: 10.1007/s00382-008-0475-6. URL <https://doi.org/10.1007/s00382-008-0475-6>.

J. Verlinde, J. Y. Harrington, G. M. McFarquhar, V. T. Yannuzzi, A. Avramov, S. Greenberg, N. Johnson, G. Zhang, M. R. Poellot, J. H. Mather, D. D. Turner, E. W. Eloranta, B. D. Zak, A. J. Prenni, J. S. Daniel, G. L. Kok, D. C. Tobin, R. Holz, K. Sassen, D. Spangenberg, P. Minnis, T. P. Tooman, M. D. Ivey, S. J. Richardson, C. P. Bahrmann, M. Shupe, P. J. DeMott, A. J. Heymsfield, and R. Schofield. The Mixed-Phase Arctic Cloud Experiment. *Bulletin of the American Meteorological Society*, 88(2):205–222, Feb. 2007. ISSN 0003-0007, 1520-0477. doi: 10.1175/BAMS-88-2-205. URL <https://journals.ametsoc.org/view/journals/bams/88/2/bams-88-2-205.xml>.

P. B. Voss, R. A. Zaveri, F. M. Flocke, H. Mao, T. P. Hartley, P. DeAmicis, I. Deonandan, G. Contreras-Jiménez, O. Martínez-Antonio, M. Figueroa Estrada, D. Greenberg,

T. L. Campos, A. J. Weinheimer, D. J. Knapp, D. D. Montzka, J. D. Crounse, P. O. Wennberg, E. Apel, S. Madronich, and B. de Foy. Long-range pollution transport during the MILAGRO-2006 campaign: a case study of a major Mexico City outflow event using free-floating altitude-controlled balloons. *Atmospheric Chemistry and Physics*, 10(15): 7137–7159, Aug. 2010. ISSN 1680-7316. doi: 10.5194/acp-10-7137-2010. URL <https://acp.copernicus.org/articles/10/7137/2010/>.

A. Walbröl, J. Michaelis, S. Becker, H. Dorff, K. Ebell, I. Gorodetskaya, B. Heinold, B. Kirbus, M. Lauer, N. Maherndl, M. Maturilli, J. Mayer, H. Müller, R. A. J. Neggers, F. M. Paulus, J. Röttenbacher, J. Rückert, I. Schirmacher, N. Slättberg, A. Ehrlich, M. Wendisch, and S. Crewell. Contrasting extremely warm and long-lasting cold air anomalies in the North Atlantic sector of the Arctic during the HALO-(AC)³ campaign. *Atmospheric Chemistry and Physics*, 2024. doi: 10.5194/acp-24-8007-2024. URL <https://doi.org/10.5194/acp-24-8007-2024>.

A. Walbröl, H. Dorff, and S. Crewell. Integrated water vapour derived from passive microwave measurements over open ocean during the HALO-(AC)³ Campaign in Spring 2022, 2026a. URL <https://doi.pangaea.de/10.1594/PANGAEA.992898>.

A. Walbröl, H. Dorff, and S. Crewell. Liquid water path derived from passive microwave measurements over open ocean during the HALO-(AC)³ Campaign in Spring 2022, 2026b. URL <https://doi.pangaea.de/10.1594/PANGAEA.992918>.

G. Wallentin, A. Oertel, L. Ickes, P. Achtert, M. Tesche, and C. Hoose. Sensitivities of simulated mixed-phase Arctic multilayer clouds to primary and secondary ice processes. *Atmospheric Chemistry and Physics*, 25(13):6607–6631, July 2025. ISSN 1680-7316. doi: 10.5194/acp-25-6607-2025. URL <https://acp.copernicus.org/articles/25/6607/2025/>.

J. J. Wang, K. Young, T. Hock, D. Lauritsen, D. Behringer, M. Black, P. G. Black, J. Franklin, J. Halverson, J. Molinari, L. Nguyen, T. Reale, J. Smith, B. Sun, Q. Wang, and J. A. Zhang. A Long-Term, High-Quality, High-Vertical-Resolution GPS Dropsonde Dataset for Hurricane and Other Studies. *Bulletin of the American Meteorological Society*, June 2015. doi: 10.1175/BAMS-D-13-00203.1. URL <https://journals.ametsoc.org/view/journals/bams/96/6/bams-d-13-00203.1.xml>. Section: Bulletin of the American Meteorological Society.

T. M. Weckwerth, J. W. Wilson, R. M. Wakimoto, and N. A. Crook. Horizontal Convective Rolls: Determining the Environmental Conditions Supporting their Existence and Characteristics. *Monthly Weather Review*, 125(4):505–526, Apr. 1997. ISSN 1520-0493, 0027-0644. doi: 10.1175/1520-0493(1997)125<0505:HCRDTE>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/mwre/125/4/1520-0493_1997_125_0505_hcrdte_2.0.co_2.xml.

M. Wendisch, D. Handorf, I. Tegen, R. A. J. Neggers, and G. Spreen. Glimpsing the Ins and Outs of the Arctic Atmospheric Cauldron. *Eos*, Mar. 2021. URL <https://eos.org/features/glimpsing-the-ins-and-outs-of-the-arctic-atmospheric-cauldron>.

M. Wendisch, M. Brückner, S. Crewell, A. Ehrlich, J. Notholt, C. Lüpkes, A. Macke, J. P. Burrows, A. Rinke, J. Quaas, M. Maturilli, V. Schemann, M. D. Shupe, E. F. Akansu, C. Barrientos-Velasco, K. Bärfuss, A.-M. Blechschmidt, K. Block, I. Bougoudis, H. Bozem, C. Böckmann, A. Bracher, H. Bresson, L. Bretschneider, M. Buschmann, D. G. Chechin, J. Chylik, S. Dahlke, H. Deneke, K. Dethloff, T. Donth, W. Dorn, R. Dupuy, K. Ebell, U. Egerer, R. Engelmann, O. Eppers, R. Gerdes, R. Gierens, I. V. Gorodetskaya, M. Gottschalk, H. Griesche, V. M. Gryanik, D. Handorf, B. Harm-Altstädter, J. Hartmann, M. Hartmann, B. Heinold, A. Herber, H. Herrmann, G. Heygster, I. Höschel, Z. Hofmann, J. Hölemann, A. Hünnerbein, S. Jafariserajehlou, E. Jäkel, C. Jacobi, M. Janout, F. Jansen, O. Jourdan, Z. Jurányi, H. Kalesse-Los, T. Kanzow, R. Käthner, L. L. Kliesch, M. Klingebiel, E. M. Knudsen, T. Kovács, W. Körtke, D. Krampe, J. Kretzschmar, D. Kreyling, B. Kulla, D. Kunkel, A. Lampert, M. Lauer, L. Lelli, A. v. Lerber, O. Linke, U. Löhnert, M. Lonardi, S. N. Losa, M. Losch, M. Maahn, M. Mech, L. Mei, S. Mertes, E. Metzner, D. Mewes, J. Michaelis, G. Mioche, M. Moser, K. Nakoudi, R. Neggers, R. Neuber, T. Nomokonova, J. Oelker, I. Papakonstantinou-Presvelou, F. Pätzold, V. Pefanis, C. Pohl, M. v. Pinxteren, A. Radovan, M. Rhein, M. Rex, A. Richter, N. Risse, C. Ritter, P. Rostosky, V. V. Rozanov, E. R. Donoso, P. S. Garfias, M. Salzmann, J. Schacht, M. Schäfer, J. Schneider, N. Schnierstein, P. Seifert, S. Seo, H. Siebert, M. A. Soppa, G. Spreen, I. S. Stachlewska, J. Stapf, F. Stratmann, I. Tegen, C. Viceto, C. Voigt, M. Vountas, A. Walbröl, M. Walter, B. Wehner, H. Wex, S. Willmes, M. Zanatta, and S. Zeppenfeld. Atmospheric and Surface Processes, and Feedback Mechanisms Determining Arctic Amplification: A Review of First Results and Prospects of the (AC)³ Project. *Bulletin of the American Meteorological Society*, 104(1):E208–E242, Jan. 2023. ISSN 0003-0007, 1520-0477. doi: 10.1175/BAMS-D-21-0218.1. URL <https://journals.ametsoc.org/view/journals/bams/104/1/BAMS-D-21-0218.1.xml>.

M. Wendisch, S. Crewell, A. Ehrlich, A. Herber, B. Kirbus, C. Lüpkes, M. Mech, S. J. Abel, E. F. Akansu, F. Ament, C. Aubry, S. Becker, S. Borrmann, H. Bozem, M. Brückner, H.-C. Clemen, S. Dahlke, G. Dekoutsidis, J. Delanoë, E. De La Torre Castro, H. Dorff, R. Dupuy, O. Eppers, F. Ewald, G. George, I. V. Gorodetskaya, S. Grawe, S. Groß, J. Hartmann, S. Henning, L. Hirsch, E. Jäkel, P. Joppe, O. Jourdan, Z. Jurányi, M. Karalis, M. Kellermann, M. Klingebiel, M. Lonardi, J. Lucke, A. E. Luebke, M. Maahn, N. Mahernndl, M. Maturilli, B. Mayer, J. Mayer, S. Mertes, J. Michaelis, M. Michalkov, G. Mioche, M. Moser, H. Müller, R. Neggers, D. Ori, D. Paul, F. M. Paulus, C. Pilz, F. Pithan, M. Pöhlker, V. Pörtge, M. Ringel, N. Risse, G. C. Roberts, S. Rosenburg, J. Röttenbacher, J. Rückert, M. Schäfer, J. Schaefer, V. Schemann, I. Schirmacher, J. Schmidt, S. Schmidt, J. Schneider, S. Schnitt, A. Schwarz, H. Siebert, H. Sodemann, T. Sperzel, G. Spreen,

B. Stevens, F. Stratmann, G. Svensson, C. Tatzelt, T. Tuch, T. Vihma, C. Voigt, L. Volkmer, A. Walbröl, A. Weber, B. Wehner, B. Wetzel, M. Wirth, and T. Zinner. Overview: quasi-Lagrangian observations of Arctic air mass transformations – introduction and initial results of the HALO-(AC)³ aircraft campaign. *Atmospheric Chemistry and Physics*, 24(15):8865–8892, Aug. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-8865-2024. URL <https://acp.copernicus.org/articles/24/8865/2024/>.

M. Wendisch, B. Kirbus, D. Ori, M. D. Shupe, S. Crewell, H. Sodemann, and V. Schemann. Observed and modeled Arctic airmass transformations during warm air intrusions and cold air outbreaks. *Atmospheric Chemistry and Physics*, 25(21):15047–15076, Nov. 2025. ISSN 1680-7316. doi: 10.5194/acp-25-15047-2025. URL <https://acp.copernicus.org/articles/25/15047/2025/>.

H. Wernli and H. C. Davies. A Lagrangian-based analysis of extratropical cyclones. I: The method and some applications. *Quarterly Journal of the Royal Meteorological Society*, 123(538):467–489, 1997. ISSN 1477-870X. doi: 10.1002/qj.49712353811. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/qj.49712353811>.

C. Wesche, D. Steinhage, and U. Nixdorf. Polar aircraft Polar5 and Polar6 operated by the Alfred Wegener Institute. *Journal of large-scale research facilities JLSRF*, 2:A87–A87, Sept. 2016. ISSN 2364-091X. doi: 10.17815/jlsrf-2-153. URL <https://jlsrf.org/index.php/lfsf/article/view/153>.

H. Wex, L. Huang, W. Zhang, H. Hung, R. Traversi, S. Becagli, R. J. Sheesley, C. E. Moffett, T. E. Barrett, R. Bossi, H. Skov, A. Hünnerbein, J. Lubitz, M. Löffler, O. Linke, M. Hartmann, P. Herenz, and F. Stratmann. Annual variability of ice-nucleating particle concentrations at different Arctic locations. *Atmospheric Chemistry and Physics*, 19(7):5293–5311, Apr. 2019. ISSN 1680-7316. doi: 10.5194/acp-19-5293-2019. URL <https://acp.copernicus.org/articles/19/5293/2019/>.

A. S. Williams, J. L. Dedrick, L. M. Russell, F. Tornow, I. Silber, A. M. Fridlind, B. Swanson, P. J. DeMott, P. Zieger, and R. Krejci. Aerosol size distribution properties associated with cold-air outbreaks in the Norwegian Arctic. *Atmospheric Chemistry and Physics*, 24(20):11791–11805, Oct. 2024. ISSN 1680-7316. doi: 10.5194/acp-24-11791-2024. URL <https://acp.copernicus.org/articles/24/11791/2024/>.

C. Woods and R. Caballero. The Role of Moist Intrusions in Winter Arctic Warming and Sea Ice Decline. *Journal of Climate*, 29(12):4473–4485, June 2016. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0773.1. URL <https://journals.ametsoc.org/view/journals/clim/29/12/jcli-d-15-0773.1.xml>.

D. L. Wu and J. N. Lee. Arctic low cloud changes as observed by MISR and CALIOP: Implication for the enhanced autumnal warming and sea ice loss. *Journal of Geophysical*

Research: Atmospheres, 117(D7), Apr. 2012. ISSN 0148-0227. doi: 10.1029/2011JD017050. URL <https://agupubs.onlinelibrary.wiley.com/doi/full/10.1029/2011JD017050>.

P. Wu and M. Ovchinnikov. Cloud Morphology Evolution in Arctic Cold-Air Outbreak: Two Cases During COMBLE Period. *Journal of Geophysical Research: Atmospheres*, 127(10):e2021JD035966, 2022. ISSN 2169-8996. doi: 10.1029/2021JD035966. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2021JD035966>.

P. Wu, M. Ovchinnikov, H. Xiao, C. P. Lackner, B. Geerts, F. Tornow, and G. Elsaesser. Effect of Ice Number Concentration on the Evolution of Boundary Layer Clouds During Arctic Marine Cold-Air Outbreaks. *Journal of Geophysical Research: Atmospheres*, 130(3):e2024JD041282, 2025. ISSN 2169-8996. doi: 10.1029/2024JD041282. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2024JD041282>.

Y. Wu and K. L. Smith. Response of Northern Hemisphere Midlatitude Circulation to Arctic Amplification in a Simple Atmospheric General Circulation Model. *Journal of Climate*, 29(6):2041–2058, Mar. 2016. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-15-0602.1. URL <https://journals.ametsoc.org/view/journals/clim/29/6/jcli-d-15-0602.1.xml>.

J. C. Wyngaard. Toward Numerical Modeling in the “Terra Incognita”. *Journal of the Atmospheric Sciences*, 61(14):1816–1826, July 2004. ISSN 0022-4928, 1520-0469. doi: 10.1175/1520-0469(2004)061<1816:TNMITT>2.0.CO;2. URL https://journals.ametsoc.org/view/journals/atsc/61/14/1520-0469_2004_061_1816_tnmitt_2.0.co_2.xml.

W. Yang and G. Magnusdottir. Springtime extreme moisture transport into the Arctic and its impact on sea ice concentration. *Journal of Geophysical Research: Atmospheres*, 122(10):5316–5329, 2017. ISSN 2169-8996. doi: 10.1002/2016JD026324. URL <https://onlinelibrary.wiley.com/doi/abs/10.1002/2016JD026324>.

G. Young, P. J. Connolly, C. Dearden, and T. W. Choularton. Relating large-scale subsidence to convection development in Arctic mixed-phase marine stratocumulus. *Atmospheric Chemistry and Physics*, 18(3):1475–1494, Feb. 2018. ISSN 1680-7316. doi: 10.5194/acp-18-1475-2018. URL <https://acp.copernicus.org/articles/18/1475/2018/>.

M. D. Zelinka, T. A. Myers, D. T. McCoy, S. Po-Chedley, P. M. Caldwell, P. Ceppi, S. A. Klein, and K. E. Taylor. Causes of Higher Climate Sensitivity in CMIP6 Models. *Geophysical Research Letters*, 47(1):e2019GL085782, 2020. ISSN 1944-8007. doi: 10.1029/2019GL085782. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2019GL085782>.

R. Zhang, H. Wang, Q. Fu, A. G. Pendergrass, M. Wang, Y. Yang, P.-L. Ma, and P. J. Rasch. Local Radiative Feedbacks Over the Arctic Based on Observed Short-Term Climate Variations. *Geophysical Research Letters*, 45(11):5761–5770, 2018. ISSN 1944-8007. doi: 10.1029/2018GL077852. URL <https://onlinelibrary.wiley.com/doi/abs/10.1029/2018GL077852>.

X. Zhao and X. Liu. Primary and secondary ice production: interactions and their relative importance. *Atmospheric Chemistry and Physics*, 22(4):2585–2600, Feb. 2022. ISSN 1680-7316. doi: 10.5194/acp-22-2585-2022. URL <https://acp.copernicus.org/articles/22/2585/2022/>.

X. Zhao, X. Liu, V. T. J. Phillips, and S. Patade. Impacts of secondary ice production on Arctic mixed-phase clouds based on ARM observations and CAM6 single-column model simulations. *Atmospheric Chemistry and Physics*, 21(7):5685–5703, Apr. 2021. ISSN 1680-7316. doi: 10.5194/acp-21-5685-2021. URL <https://acp.copernicus.org/articles/21/5685/2021/>.

W. Zhou, L. R. Leung, and J. Lu. Steady threefold Arctic amplification of externally forced warming masked by natural variability. *Nature geoscience*, 17, 2024. doi: 10.1038/s41561-024-01441-1. URL <https://doi.org/10.1038/s41561-024-01441-1>.

X. Zhou, P. Kollias, and E. R. Lewis. Clouds, Precipitation, and Marine Boundary Layer Structure during the MAGIC Field Campaign. *Journal of Climate*, 28(6):2420–2442, Mar. 2015. ISSN 0894-8755, 1520-0442. doi: 10.1175/JCLI-D-14-00320.1. URL <https://journals.ametsoc.org/view/journals/clim/28/6/jcli-d-14-00320.1.xml>.

P. Zuidema, B. Geerts, G. Mcfarquhar, A. Bailey, J. Cassano, J. Doyle, J. French, P. J. Demott, A. Dzambo, C. Grasmick, E. Jarvinen, P. Romashkin, E. Rosky, C. Wolff, and D. Zietlow. Bulletin of the American Meteorological Society Flying into the Cold: The NSF Cold Air outbreak Experiment in the Sub-Arctic (CAESAR). 2026. URL <https://www.authorea.com/users/528833/articles/1377961-bulletin-of-the-american-meteorological-society-flying-into-the-cold-the-nsf-cold-air-outbreak-experiment-in-the-sub-arctic-caesar?commit=f91d082c754dddbfa06c7b722a2709c1530769>.

THIS thesis was typeset using $\text{\LaTeX}$, originally developed by Leslie Lamport and based on Donald Knuth's $\text{\TeX}$. The body text is set in 11 point Egenolff-Berner Garamond, a revival of Claude Garamont's humanist typeface. The above illustration, "Science Experiment 02", was created by Ben Schlitter and released under [CC BY-NC-ND 3.0](#). A template that can be used to format a PhD thesis with this look and feel has been released under the permissive MIT (X11) license, and can be found online at github.com/suchow/Dissertate or from its author, Jordan Suchow, at suchow@post.harvard.edu.