
On the Physical Processes Controlling Warm Conveyor Belt Moisture Transport into the Upper Troposphere and Lower Stratosphere

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Abstract

Warm conveyor belts (WCBs) are large-scale, coherently ascending airstreams that greatly influence extratropical weather. They are also a key contributor to upper tropospheric and lower stratospheric (UTLS) water vapor, which is a potent greenhouse gas. Therefore, accurately representing WCBs in models is crucial for weather and climate predictions. However, three key uncertainties in WCB-driven moisture transport into the UTLS remain, and they concern (i) the microphysical processes that regulate this transport and the differences between convective and slantwise ascent; (ii) the effect of the representation of convection (parameterized vs. explicit) on modeled moisture transport; and (iii) the sensitivity of this transport to microphysical parameterizations.

In Chapter 3, we investigate the microphysical processes involved in WCB moisture transport and the role of convection using a convection-permitting ICON simulation of a WCB case study. We find that the moisture content at the end of the WCB ascent is strongly dependent on the ascent timescale. Trajectories undergoing rapid ascent tend to glaciate later and contain a large number of small ice crystals by the end of the ascent. In contrast, slowly ascending trajectories glaciate earlier and contain fewer ice crystals. For both ascent types, the final water vapor content is primarily governed by temperature and pressure.

In Chapters 4, 5 and 6, we investigate the role of simulation scale by examining the same WCB with an additional, convection parameterizing simulation. We find that too much vapor and too few ice crystals are transported into the UTLS by the convection parameterization scheme, resulting in less outgoing longwave radiation. Comparisons with satellite observations suggest that the convection-permitting simulation yields more accurate results. This suggests that convection parameterizing simulations may not accurately represent the radiative effects of WCBs, among other things.

Finally, in Chapter 7, we analyze 70 convection permitting simulations with perturbed parameter choices of one WCB case study. We find that ice content and relative humidity at the end of the ascent depend strongly on the choice for the ice capacitance and INP concentrations. Temperature, pressure and specific humidity remain mostly unchanged, even when the saturation adjustment scheme allows for the production of supersaturated conditions in liquid clouds, or when sea surface temperatures are increased by 2 K. These results provide key insights into longstanding uncertainties concerning WCB moisture transport into the UTLS and together with the methods we developed, provide a solid foundation for future investigations into this important topic.

Zusammenfassung

Warm conveyor belts (WCBs) sind großräumige, kohärent aufsteigende Luftströme, die das extratropische Wetter stark beeinflussen. Sie tragen auch wesentlich zum Wasserdampfgehalt in der oberen Troposphäre und unteren Stratosphäre (UTLS) bei, was für den Treibhauseffekt wichtig ist. Daher ist die genaue Darstellung von WCBs in Modellen für Wetter- und Klimaprognosen von entscheidender Bedeutung. Allerdings bestehen weiterhin drei wesentliche Unsicherheiten hinsichtlich des durch WCBs verursachten Feuchtigkeitstransports in die UTLS, die folgende Punkte betreffen: (i) die mikrophysikalischen Prozesse, die diesen Transport regulieren, und die Unterschiede zwischen konvektivem und schrägem Aufstieg; (ii) die Auswirkungen der Darstellung der Konvektion (parametrisiert vs. explizit berechnet) auf den modellierten Feuchtigkeitstransport; und (iii) die Empfindlichkeit dieses Transports gegenüber der Auswahl von Parametern mikrophysikalischen Parametrisierungen.

In Kapitel 3 untersuchen wir anhand einer konvektionszulässigen ICON-Simulation einer WCB-Fallstudie die mikrophysikalischen Prozesse, die beim Feuchtigkeitstransport in WCBs eine Rolle spielen, sowie die Rolle der Konvektion dabei. Wir stellen fest, dass der Feuchtigkeitsgehalt am Ende des WCB-Aufstiegs stark von der Aufstiegszeit abhängt. Trajektorien, die einen schnellen Aufstieg durchlaufen, neigen dazu, später zu vereisen und enthalten am Ende des Aufstiegs eine große Anzahl kleiner Eiskristalle. Im Gegensatz dazu vereisen langsam aufsteigende Trajektorien früher und enthalten weniger Eiskristalle. Bei beiden Aufstiegstypen wird der endgültige Wasserdampfgehalt in erster Linie von Temperatur und Druck bestimmt. In den Kapiteln 4, 5 und 6 untersuchen wir die Rolle der Simulationsskala, indem wir denselben WCB mit einer zusätzlichen Simulation, die Konvektion parametrisiert, untersuchen. Wir stellen fest, dass durch das Konvektionsparametrisierungsschema zu viel Wasserdampf und zu wenige Eiskristalle in die UTLS transportiert werden, was zu einer geringeren ausgehenden Langwellenstrahlung führt.

Vergleiche mit Satellitenbeobachtungen deuten darauf hin, dass die konvektionsfähige Simulation genauere Ergebnisse liefert. Dies lässt vermuten, dass Simulationen, die Konvektion parametrisieren, unter anderem die Strahlungseffekte von WCBs nicht genau wiedergeben. Zu aller letzt analysieren wir in Kapitel 7 70 konvektionsfähige Simulationen mit gestörten Parameterauswahlen einer WCB-Fallstudie. Wir stellen fest, dass der Eisgehalt und die relative Luftfeuchtigkeit am Ende des Aufstiegs stark von der Wahl der Eiskapazität und der INP-Konzentrationen abhängen. Temperatur, Druck und spezifische Feuchte bleiben weitgehend unverändert, selbst wenn das Schema zur Anpassung der Sättigung ("saturation adjustment scheme") die Entstehung übersättigter Bedingungen in Flüssigwolken zulässt oder wenn die Meeresoberflächentemperaturen um 2 K erhöht werden. Diese Ergebnisse liefern wichtige Erkenntnisse zu den seit langem bestehenden Unsicherheiten hinsichtlich des Feuchtetransports von WCB in die UTLS und bilden zusammen mit den von uns entwickelten Methoden eine solide Grundlage für zukünftige Untersuchungen zu diesem wichtigen Thema.

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Chapter 1

Motivation

The atmosphere is one of the most complex and dynamic systems in nature. Across scales spanning more than fifteen orders of magnitude—from planetary Rossby waves meandering over hemispheres to molecular vibrations producing spectral lines—a symphony of natural phenomena produces storms, floods, tornadoes, hurricanes, rainbows, droughts, noctilucent clouds, hail, lightning, the aurora borealis, heat waves, and haboobs. The atmosphere is bombarded from above by cosmic rays, meteoroid dust, and the charged particles of the solar wind, while volcanoes, oceans, forests, factory stacks, tires, ship exhausts, wildfires, and penguin guano feed the atmosphere from below. Within this atmospheric orchestra, atmospheric constituents such as dust grains, ice crystals, and organic molecules are continually lofted, transformed, and returned to Earth in an endless cycle. Throughout history, people attributed these marvels to supernatural forces and found themselves utterly at the mercy of the sky. A stretch of gentle seasons could fill granaries and allow cultures to flourish; a run of failed monsoons could ruin empires. The same rivers that nourished cities one year could destroy them the next; the winds that carried spice ships across the world turned into storms that sank entire navies.

Today, we don't rely on the supernatural to explain these marvels. Instead, we deploy an international corps of meteorologists who radiosonde the troposphere at dawn, tend instruments from Svalbard to the Sahara, launch satellites into orbit, and run models on petascale computers. We have evolved from mere observers of nature's forces into a force within the system itself. We reshape the land, redirecting water and altering nutrient cycles. We pollute the air, transforming clouds. We release carbon dioxide and methane at levels unseen for millennia, potentially triggering our own undoing and threatening much of life on Earth. However, despite our influence, engineering achievements, and scientific breakthroughs, we are still vulnerable to droughts, floods, fires, hurricanes, and hailstorms, or even a simple sandstorm that stranded a ship in the Suez Canal and brought global trade to a halt. As atmospheric scientists we strive for a more complete understanding of the atmosphere, but that does not guarantee control over it. However, it offers the possibility of anticipating and mitigating the consequences of our actions. It is in pursuit of this understanding that we carry out our research; even if after many years of hard work, each individual and every dissertation can only contribute a minuscule amount of knowledge to this vast field of research.

Chapter 2

Introduction

Of the many dynamic phenomena that atmospheric scientists aim to understand, extratropical cyclones stand out as some of the most influential. These vast low-pressure systems produce much of the cloud cover, precipitation, and severe weather across the mid-latitudes. Their study has played a central role in advancing atmospheric science throughout the 20th century and they remain a cornerstone of meteorological research today. The most consequential process that unfolds within extratropical cyclones is the movement of water vapor. This movement fuels storms, governs cloud and precipitation formation, and stores and releases latent heat, which facilitates the transport of energy from the tropics to the poles.

While much attention has been given to water’s role in the lower atmosphere, a subtler but no less important story unfolds higher up, near the boundary between the upper troposphere and lower stratosphere (UTLS). Here, small changes in moisture can have a large impact on cloud formation, Earth’s radiative balance, and stratospheric chemistry. One of the key players in this vertical transport is the *warm conveyor belt* (WCB)—a coherent stream of warm, moist, and ascending air that forms within extratropical cyclones. These airstreams can carry moisture from the boundary layer to the UTLS in a matter of hours, injecting water vapor and ice particles into regions where even trace amounts influence the broader climate system. Investigations of this process and the factors that govern it form the core of this thesis.

In this introduction, we begin with a general overview of Earth’s atmosphere, focusing on the regions and phenomena most relevant to UTLS moisture and WCBs, and review the current state of knowledge (Section 2.1). We then identify key knowledge gaps that motivate our research, and state the main research questions (Section 2.2). Next, we introduce the theoretical background on thermodynamics and cloud physics necessary for understanding the processes investigated (Section 2.3), and provide an overview of the numerical weather model used in this work, including the relevant physical parameterizations (Section 2.4). Finally, we give an overview of the results and the publications that form this thesis (Section 2.5).

2.1 Earth’s Atmosphere

The Earth’s atmosphere is a continuous, compressible fluid that extends from Earth’s surface into outer space, and consists mostly of nitrogen, oxygen, argon, water vapor, and various other gases, such as carbon dioxide. While it lacks a distinct boundary beyond the surface, different regions of the atmosphere have vastly different temperatures, pressures, and concentrations of constituents, giving rise to distinct phenomena. For this reason, atmospheric scientists have found it useful to categorize the atmosphere into separate regions according to changes in temperature with altitude (Fig. 2.1).

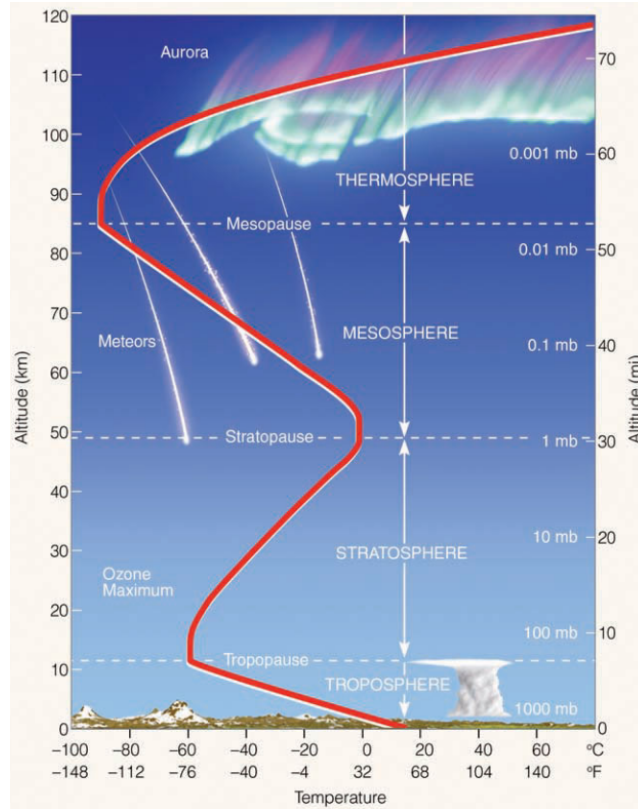


Figure 2.1: Division of the atmosphere into the troposphere, stratosphere, mesosphere and thermosphere according to changes in temperature with altitude (taken from Ahrens (2009))

2.1.1 The Troposphere and Stratosphere

In the troposphere, the layer closest to the surface ranging from 0 km to ~ 12 km, temperatures decrease with altitude due to decreasing pressure. The troposphere may seem like a thin slice of the atmosphere, but it contains more than half of its total mass and is the most chaotic and dynamic region, driving most of Earth's weather (Häckel, 2021). In the stratosphere directly above, ranging from approximately 10 km to 50 km, temperature increases with altitude due to localized heating resulting from the absorption of high energy radiation by ozone molecules (Ackerman and Knox, 2012). Unlike the turbulent, convectively driven processes of the troposphere, the stratosphere is characterized by more stable, stratified conditions and is dominated by radiative and large-scale dynamical processes (Baldwin et al., 2003). Regions beyond the stratosphere are fascinating, but they are not relevant to this thesis and not discussed further.

2.1.2 The Tropopause

The region separating the troposphere and the stratosphere is called the tropopause (at an altitude of approximately 10 km (17 km) in the extratropics (tropics); Fig. 2.1 and 2.4). It is traditionally defined as the lowest level at which the lapse rate decreases to 2 K km^{-1} or less, provided that the average lapse rate between this level and all higher levels within 2 km does not exceed 2 K km^{-1} (World Meteorological Organization, 1957). However, there are several other definitions that offer different benefits for different purposes and/or regions.

Examples include the cold point tropopause (i.e., temperature minimum) which is useful in the tropics (Highwood and Hoskins, 1998); stability-based definitions that use the Brunt–Väisälä frequency (Gettelman and Wang, 2015; Tinney et al., 2022); dynamical definitions that use the potential vorticity (PV; Kunz et al. (2011)); and the chemical tropopause, that makes use of the large composition change between troposphere and stratosphere (Bethan et al., 1996; Prather et al., 2011; Bauchinger et al., 2025).

Naturally, the tropopause is not a single point, but a layer that can be several meters to kilometers deep (Hoinka, 1997; Tinney et al., 2022). The stability of this layer, caused by its inverted temperature profile, makes it difficult for air to rise past this point and effectively makes it a barrier between the troposphere and the stratosphere (Gettelman et al., 2011). This is why almost all clouds reside within the troposphere only (Fig. 2.2). Nevertheless, the tropopause is not an impervious barrier. Some clouds can ascend with such rigor that their tops penetrate into the stratosphere (Pommereau, 2010; Frey et al., 2015; Dauhut et al., 2018), and vertical wind shear as well as gravity wave breaking can induce turbulent mixing and facilitate vertical transport (Hoor et al., 2010; Heller et al., 2017; Kaluza et al., 2021; Lachnitt et al., 2023; Chau et al., 2024; Dörnbrack et al., 2025). Therefore, trace gases, water vapor, hydrometeors, and aerosols can be exchanged between the troposphere and the stratosphere (Cristofanelli et al., 2003; Joppe et al., 2025; Breuninger et al., 2025).

At first glance, the exchange of air masses between the troposphere and stratosphere might not seem that important. After all, the stratosphere is largely decoupled from the turbulent weather systems that directly affect life at the surface. However, the stratosphere plays a vital role in Earth’s climate system. Stratospheric water vapor, though only present in small concentrations, is a potent greenhouse gas (Shindell, 2001; Solomon et al., 2010; Dessler et al., 2013), and stratospheric aerosols can reflect incoming solar radiation and exert a cooling effect on the planet (Kremser et al., 2016). Crucially, both stratospheric water vapor and aerosols are primarily sourced from the troposphere. Thus, understanding how air masses are transported across the tropopause is essential for quantifying the composition and radiative impact of the stratosphere. Additionally, the stratosphere does influence the weather at Earth’s surface to some extent (Baldwin et al., 2003). For example, anomalies in the stratospheric polar vortex can modify the jet stream in ways that favor mid-latitude blocking, which can slow the progression of weather systems in the troposphere and lead to prolonged heatwaves and drought (Lupo, 2020). It has been shown that tropospheric phenomena such as extratropical cyclones can influence these anomalies (Oertel et al., 2020), making it clear that interactions between the troposphere and the stratosphere beyond the simple exchange of mass are also important for weather and climate. Therefore, the region around the tropopause where these interactions occur—the upper troposphere and lower stratosphere (UTLS)—has become a widely studied area (Gettelman et al., 2011; Sharman et al., 2012). It is also the region of focus for this thesis.

2.1.3 The Extratropics

Naturally, any investigation must narrow its scope. In this thesis, we concentrate on the UTLS located in the extratropical regions of the Earth. The extratropics are located between

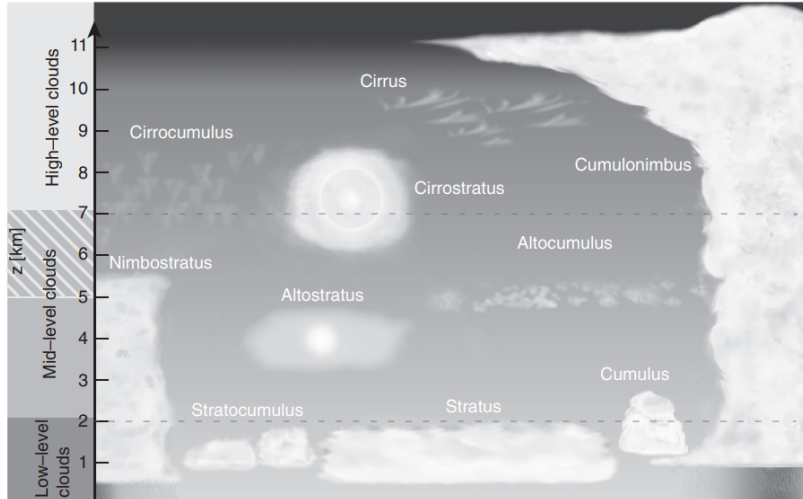


Figure 2.2: Schematic showing the ten cloud types classified by the World Meteorological Association and their approximate altitudes (taken from Lohmann et al. (2016)).

approximately 30° and 60° of latitude in both hemispheres (Fig. 2.3). They are bordered by the tropics towards the equator and the polar regions towards the poles. The tropics ($\approx 23.5^\circ$ S to 23.5° N) experience abundant year round solar heating and a small Coriolis force, which leads to weather that is dominated by heat-driven convection (Bao et al., 2022). In contrast, the polar regions (above 60° N/S) experience weak solar input and continuous surface cooling which creates a strongly stratified atmosphere (Jozef et al., 2023; Dice et al., 2023), and a strong Coriolis force gives rise to planetary waves and powerful polar lows (Stoll, 2022). The extratropics combine these aspects and experience the effects of both solar radiation and the Coriolis force, resulting in sharp temperature gradients, baroclinic instability, westerly jet streams, and fast-moving, synoptic-scale cyclones and anticyclones (Häckel, 2021). Therefore, the weather in the extratropics is highly dynamic and versatile, with many different phenomena that vary geographically and seasonally influencing the UTLS region.

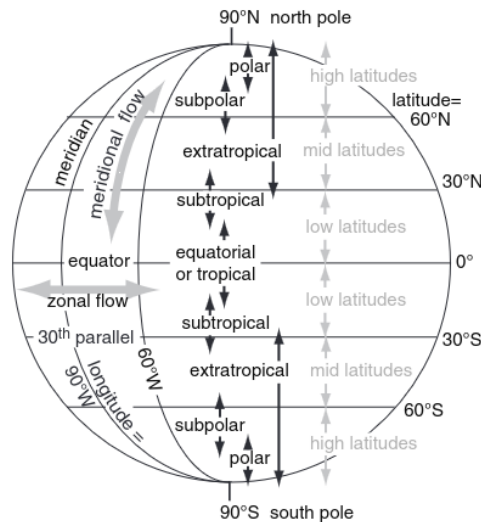


Figure 2.3: Nomenclature for global climate zones and geography, taken from Stull (2015a).

2.1.4 The Extratropical UTLS

The extratropical UTLS (Ex-UTLS) has no rigidly defined boundaries but is generally understood to exist poleward of the subtropical jet (blue shaded region in Fig. 2.4). The tropopause, irrespective of the choice of definition, decreases in altitude from approximately 13 km at the subtropical jet to approximately 7 km at the poles (red dotted contour for thermal tropopause and blue contour for 2PVU tropopause in Fig. 2.4). The region just below the tropopause is the approximate lower boundary of the Ex-UTLS, while the region above—the lowermost stratosphere (LMS)—is defined as the region where isentropes intersect the tropopause and extends up to the 400–440 K isentrope (Gettelman et al., 2011). Above the 400 K isentrope, the Brewer-Dobson circulation drives the transport of stratospheric air poleward in an upward direction over the tropics and in a downward direction over the extratropics and poles (hollow arrows in Fig. 2.4). In the troposphere below the Ex-UTLS, baroclinic waves, weather fronts, gravity waves and extratropical cyclones undulate the tropopause and can produce exotic conditions such as tropopause "folds" (Fig. 2.7 a) and stratospheric cutoffs (Keyser and Shapiro, 1986; Gettelman et al., 2011; Škerlak et al., 2015).

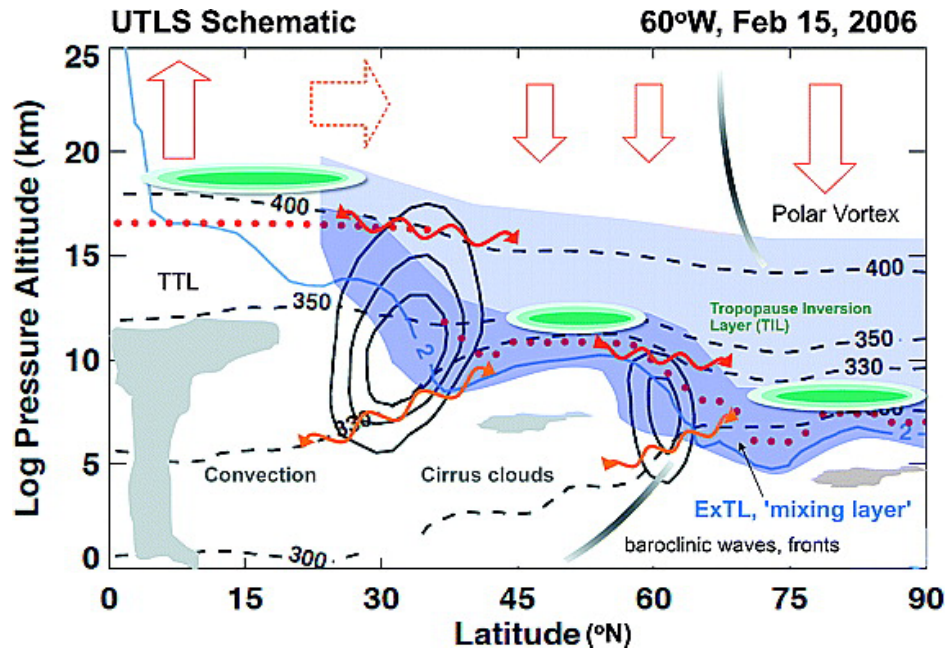


Figure 2.4: Schematic overview of the UTLS using an atmospheric cross-section along 60°W in the NH. The data depicts conditions on 15 February 2006. Illustrated objects represent the following key atmospheric features: the UTLS region (dark and light blue); clouds (grey); regions of enhanced static stability in the tropopause-inversion layer (green); quasi-isentropic exchange (red wavy arrow); cross-isentropic exchange (orange wavy arrow); and the Brewer-Dobson Circulation (hollow arrows). Contoured lines indicated the thermal tropopause (red dotted), potential temperature surfaces (black dashed), potential vorticity surface (2PVU; solid blue), and wind (solid black, intervals of 10 ms⁻¹). Figure taken from Gettelman et al. (2011).

Concentrations of chemical constituents such as Ozone (O₃), Nitrous-Oxide (N₂O), and Carbon Monoxide (CO) exhibit strong vertical gradients across the Ex-UTLS (Prather et al., 2011; Gettelman et al., 2011; Bauchinger et al., 2025). Specifically, O₃ concentrations are high in the stratosphere and low in the troposphere. Therefore, cross-isentropic exchange, which can occur in regions of subtropical and polar jet streams (orange wavy arrows in Fig. 2.4) due

to turbulence induced by the strong wind shear (Lelieveld et al., 1997; Gettelman et al., 2011), contributes to the O_3 budget of the troposphere (Roelofs and Lelieveld, 1997; Gettelman et al., 2011). However, this thesis focuses on water (vapor), which also exhibits a strong concentration gradient across the Ex-UTLS.

Water vapor concentrations are typically high in the troposphere and decrease rapidly across the Ex-UTLS region, becoming very low in the stratosphere (Gettelman et al., 2011). This occurs because the air that feeds the stratosphere through the Brewer–Dobson circulation’s rising branch experiences freeze-drying, causing almost all of the air’s water to form precipitation and sediment, leaving the ascending air extremely dry (Randel and Park, 2019). Cloud content is also high in the troposphere and low in the stratosphere, because the Ex-UTLS region contains local temperature minima and statically stable regions of the atmosphere (green regions in Fig. 2.4), leading to it generally being the highest region in which clouds can form (Gettelman et al., 2011). For these reasons, two-way mixing across the Ex-UTLS and the moisture content of the Ex-UTLS region itself are important factors in stratospheric moisture content.

2.1.5 Moisture in the UTLS and Stratosphere

Overall, water vapor is the most dominant greenhouse gas in the atmosphere (Schneider et al., 2010). Changes to its concentration in the UTLS and the stratosphere in particular are one of the most important positive feedback mechanisms for climate change (Held and Soden, 2000; Dessler et al., 2013; Solomon et al., 2010; Nowack et al., 2023), with even minor changes in water vapor content able to substantially impact the radiative budget of Earth (Hansen et al., 1984; Wang et al., 2001). For example, Forster and Shine (2002) determined that increasing stratospheric water vapor concentrations from 1980 to 2000 might have contributed as much as $+0.29 \text{ Wm}^{-2}$ to global radiative forcing, and Solomon et al. (2010) determined that decreasing stratospheric water vapor concentrations from 2000 to 2006 might have slowed the rate of increase in global surface temperatures by about 25%. Therefore, significant efforts have been made to measure and quantify the amount and regional distribution of moisture in the UTLS, as well as the changes in these quantities over time (Tilmes et al., 2010; Hurst et al., 2011; Zahn et al., 2014; Jeffery et al., 2022; Konjari et al., 2024; Xu et al., 2025).

These efforts are hindered by the fact that the UTLS is a notoriously difficult region to characterize accurately. The challenging measurement conditions—high altitudes, low temperatures, low water vapor concentrations, and strong vertical gradients with high temporal variability—result in sparse data and significant uncertainty (Petzold et al., 2020; Jeffery et al., 2022). While this makes accurate analyses difficult, overall measurements indicate that water vapor concentrations in the stratosphere and UTLS are increasing. For example, Hurst et al. (2011) analyzed 30 years of balloon-borne measurements of stratospheric water vapor from 1980 to 2010 and found overall increasing concentrations (Fig. 2.5). From 2000 to 2006, concentrations decreased (Randel et al., 2006), but have increased again since (Yue et al., 2019) and are projected to keep increasing in the future (Sheese et al., 2025).

In order to solidify the understanding of how UTLS and stratospheric water vapor concentrations will change in the future, it is important to identify and quantify the processes

that transport water vapor into these regions. One major transport pathway for UTLS moisture is overshooting deep convection, which has been studied intensively using both measurements (Danielsen, 1993; Corti et al., 2008; Lee et al., 2019; Gordon et al., 2024) and simulations (Hassim and Lane, 2010; Dauhut et al., 2018; Ueyama et al., 2018, 2023). Convectively ascending air is called "overshooting" when it rises beyond its neutral buoyancy level (Dauhut et al., 2018), and in the tropics, this is estimated to occur in approximately 0.1% of convective systems (Liu and Zipser, 2005).

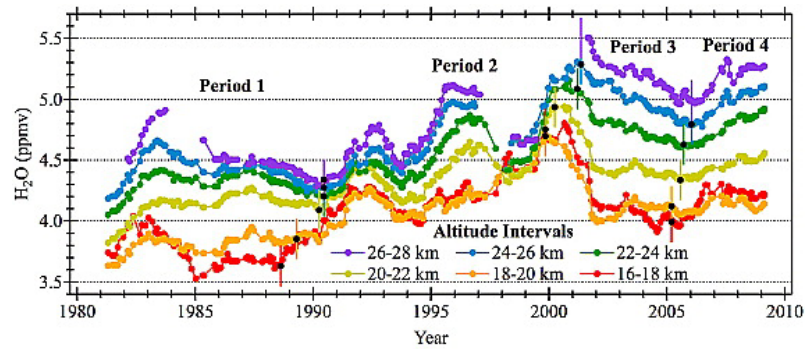


Figure 2.5: Data for 30 years of balloon-borne stratospheric water vapor measurements, taken from 1980 to 2010 over Boulder, Colorado. The colors indicate the layers of the atmosphere, and the colored dots show the moving averages of the measurements for each layer. Figure taken from Hurst et al. (2011).

The mechanism that allows overshooting convective tops to hydrate the stratosphere is explained by Dauhut et al. (2018). Cold air rises into the lower stratosphere (Fig. 2.6 a) and entrains dry, stratospheric air at its top (Fig. 2.6 b). This warms the top of cloud and causes ice particles to sublimate, forming a humid pocket of air at the top of the overshoot (Fig. 2.6 b). The negatively buoyant air below the pocket wants to sink but is pushed outward by rising air from below (black arrows in Fig. 2.6 b), producing wind shear that further enhances mixing of the stratospheric air with the humid and warmer pocket above (Fig. 2.6 c). This process takes no more than a couple of minutes (Dauhut et al., 2018).

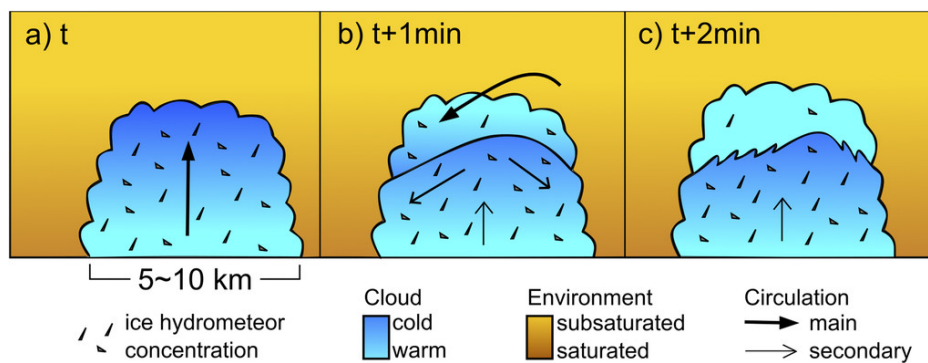


Figure 2.6: Schematic of the mechanism proposed to allow for the hydration of the stratosphere by overshooting convection. In a), cold air rises at the top of a convective cloud beyond its level of neutral buoyancy. In b), the top of the cloud mixes with stratospheric air, which is warmer, causing ice particles to sublimate and humidity to rise. This produces a pocket of more humid and dry air that is cut-off from the rest of the cloud. The negatively buoyant air below wants to sink but is pushed to the side by rising air from below (arrows in c), which enhances shear and further promotes mixing. In c), the two pockets are separated and the top, humid pocket remains in the stratosphere. Figure is taken from Dauhut et al. (2018).

While this process predominantly occurs in the tropics (Liu and Zipser, 2005), evidence for stratospheric hydration by overshooting convection has also been found in extratropical storms over Sweden (Konjari et al., 2025), the United States (Homeyer et al., 2014; Homeyer, 2015; Homeyer et al., 2024; Gordon et al., 2024) and the Tibetan plateau (Fu et al., 2006; Sun et al., 2017). Another pathway that has not been studied in detail is the transport of moisture by gravity waves, for which evidence was found over New Zealand (Heller et al., 2017).

Extratropical cyclones are a common weather feature in the extratropics and are known to transport large masses of air into the UTLS region (Madonna et al., 2014). To pinpoint the primary pathways through which water vapor enters the UTLS from these candidates, a multi-year analysis of monthly UTLS water vapor measurements was conducted by Zahn et al. (2014). They determined that warm conveyor belts (WCBs), which are prominent features of extratropical cyclones (ETCs), are one of the main contributors to UTLS moisture in the extratropical regions of the Earth. This thesis focuses on WCBs and their role for transporting moisture to the UTLS.

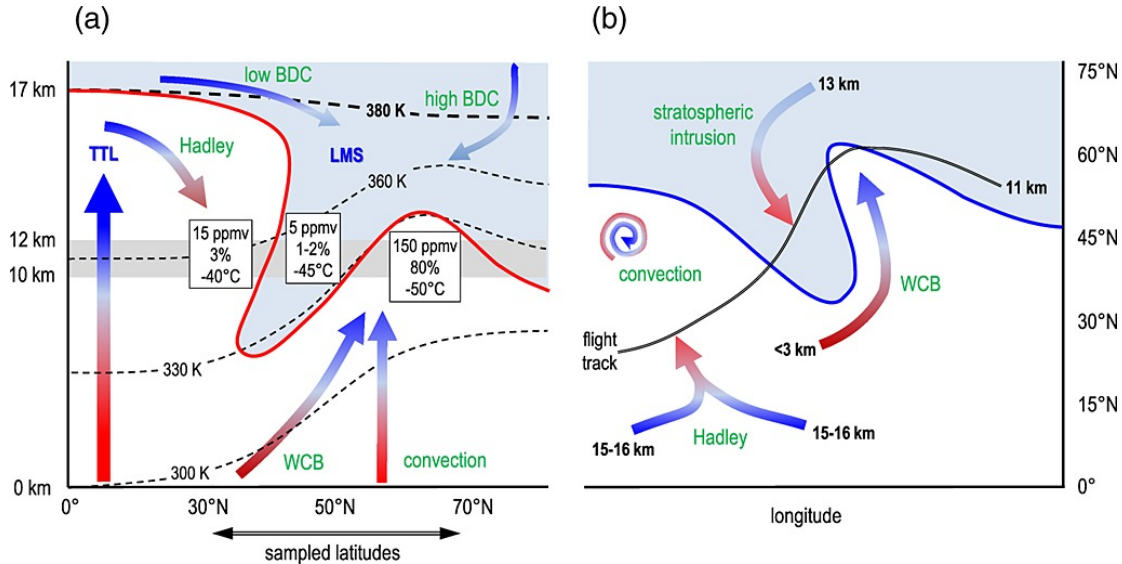


Figure 2.7: Schematic showing the dominant pathways (indicated by arrows) for moisture transport into the UTLS, using a cross section across latitudes in a), and as seen from above in b). In a) the red line shows a tropopause during a folding event; the grey shaded area indicates cruising altitudes for passenger aircraft which were fitted with measurement devices; and dashed lines indicate isentropes. Figure taken from (Zahn et al., 2014).

2.1.6 Extratropical Cyclones

In the extratropics, the latitudinal differences in solar heating set up a meridional temperature gradient, and when winds act frontogenically (i.e., squeeze warm and cold air together, see Fig. 2.8 a), these gradients can become very large and give rise to weather fronts (Bluestein, 1986). Along an elongated, quasi-stationary frontal zone that separates warm subtropical air from cold polar air, the horizontal temperature gradient—and therefore the baroclinicity—is large (Charney, 1947). Additionally, wind direction on the warm and cold sides often show in the opposite direction (Fig. 2.8 b i), producing wind shear and unstable conditions Bjerknes and Solberg (1922). Therefore, almost any perturbation to the front can produce an initial wavelet (Fig. 2.8 b ii), and the movement of air brought about by the wavelet allows the

warmer air to rise over the heavier cold air, which produces an initial low pressure system at the surface (Fig. 2.8 b iii).

The potential energy stored within the warm air is subsequently converted into kinetic energy, which accelerates the cyclonic wind field (Fig. 2.8 b iii; Bjerknes and Solberg (1922)). Often times, this surface low only intensifies into the development stage (Fig. 2.8 b iv) when it lies downstream of an upper-tropospheric trough or jet streak, where positive vorticity advection aloft induces divergence, cyclonic vorticity and upward motion at the surface (Fig. 2.9; Bluestein (1993); Catto (2016)). This is a very rudimentary but generally applicable overview of how low pressure systems and cyclonic winds arise in the mid-latitudes to produce phenomena referred to as "extratropical cyclones".

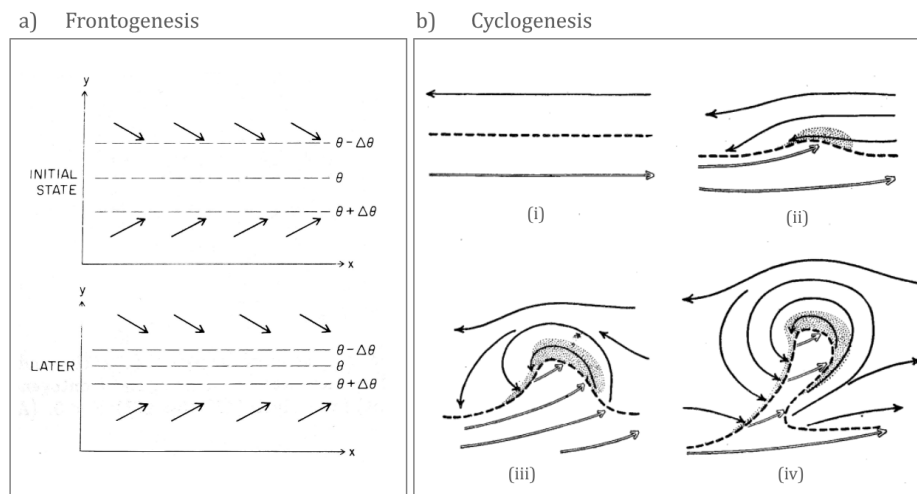


Figure 2.8: a) Schematic showing how frontogenic winds increase temperature (θ) gradients in latitude (y) by "squeezing" cold and warm air together. Image taken from (Bluestein, 1986). b) The idealized process of cyclogenesis as depicted by (Bjerknes and Solberg, 1922). In (i), an initial, quasi-stationary front with opposing wind directions has formed. In (ii), a perturbation causes a wavelet to form, allowing for the warm air to move over the cold air, producing a low pressure system in (iii). This then evolves further, with cyclonic winds and highly dynamic front, to produce an extratropical cyclone in (iv).

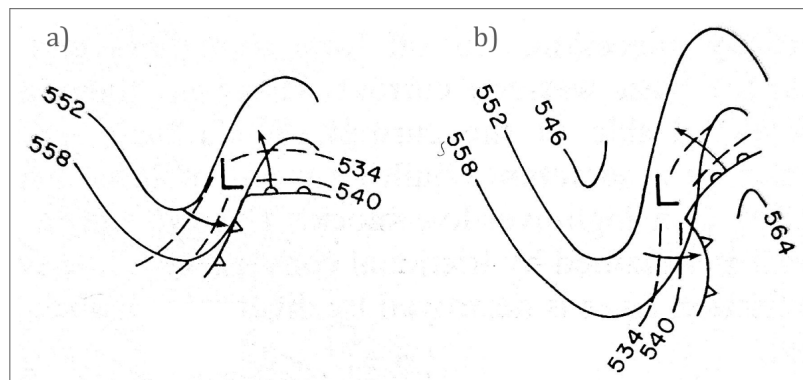


Figure 2.9: Idealized schematic showing how a surface low (dashed lines) is strengthened from one time step a) to a later time step b) by interacting with an upper level wave, depicted by the 500 mb geopotential height contours (solid lines). Image taken from Bluestein (1993).

Extratropical (or mid-latitude) cyclones (ETCs) are large-scale, low-pressure systems with cyclonic winds and a distinct cloud band (Fig. 2.10). They drive much of the day-to-day weather variability poleward of the subtropics (Schultz et al., 2019; Laurila et al., 2021) and transport huge amounts of moisture and energy across latitudes (Catto, 2016). They account for as many as 80% of all extreme precipitation events in the extratropics (Pfahl and Wernli, 2012), and are one of the main drivers for large-scale weather patterns (Rodwell et al., 2018).

In what reads as a love letter to extratropical cyclones, Schultz et al. (2019) portray them as the “centerpiece of meteorology,” celebrating their central role in advancing and developing modern atmospheric science in the 20th century. They emphasize—affectionately—that the intricate and nonlinear interactions of baroclinic and diabatic processes, coupled with the broad spatial variability of ETCs, ensure that each cyclone is unique (Schultz et al., 2019). This statement is supported by the many failed attempts to generalize and predict ETC behavior by classifying them into different types according to the region of origin (Dacre and Gray, 2009), the degree of intensification (Binder et al., 2016), or the conditions experienced during cyclogenesis (Petterssen and Smebye, 1971; Graf et al., 2017). It seems that nature resists our urge to compartmentalize this fascinating phenomenon. As Schultz et al. (2019) point out, perhaps this is because ETCs are not really *things* located at a specific time and place in the atmosphere, but rather *processes* that occur throughout the atmosphere and manifest as surface lows that we label “cyclones.” For the sake of readability, however, this thesis will treat ETCs as a thing.

2.1.7 Warm Conveyor Belts

While it is difficult to categorize different types of ETCs, there are some features all ETCs share. There are four dominant airflows within ETCs (Fig. 2.11): the warm conveyor belt (WCB), the cold conveyor belt (CCB), the dry intrusion (DI), and the sting jet (SJ; Martínez-Alvarado et al. (2014)). Arguably the most prominent feature of ETCs is the WCB (Catto, 2016), which is also the primary focus of this thesis. In general, a WCB is a coherent stream of warm, moist boundary-layer air that rises within an ETC ahead of the cold front and then glides poleward and eastward over the warm front into the upper troposphere (Madonna et al., 2014; Joos et al., 2023). Once the airstream has reached the UTLS region, it branches into one cyclonic and one anticyclonic branch (Fig. 2.11; Martínez-Alvarado et al. (2014)). The ascending air produces clouds, and together with the cyclonic rotation of the ETC this gives rise to the characteristic comma-shaped cloud band that makes WCBs easily recognizable from satellite imagery (Fig. 2.10).

It takes the air masses ascending as part of the WCB approximately two days to travel from the troposphere, where the pressure is around 900 hPa, to the UTLS region, where the pressure is around 300 hPa (see Fig. 2.11; Madonna et al. (2014); Martínez-Alvarado et al. (2014)). This has led to a widely adopted threshold for defining WCB air parcels, which must ascend to 600 hPa in no more than 48 hours (Wernli, 1997). This threshold is used in virtually every WCB study and it has been shown that relaxing this threshold does not substantially change various WCB analyses, indicating that it is a very robust and universal threshold for WCBs (Heitmann et al., 2024). The minimum time required for an air parcel to rise 600 hPa is denoted by τ_{600} ; a variable that will be featured often in this thesis.

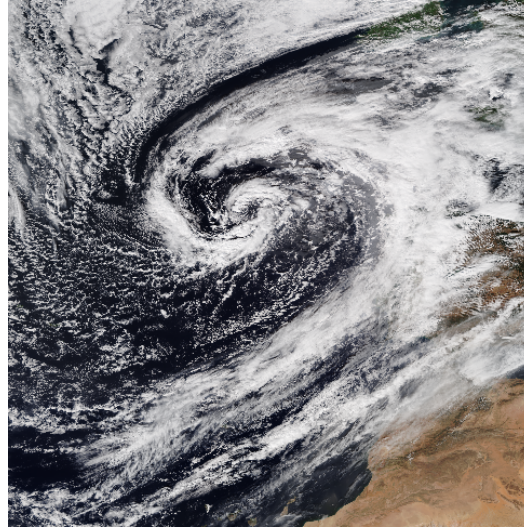


Figure 2.10: Satellite image of an extratropical with a pronounced warm conveyor belt over the Eastern North Atlantic Ocean making landfall over Spain on 22 October, 2022 (National Oceanic and Atmospheric Administration, NESDIS Environmental Visualization Lab).

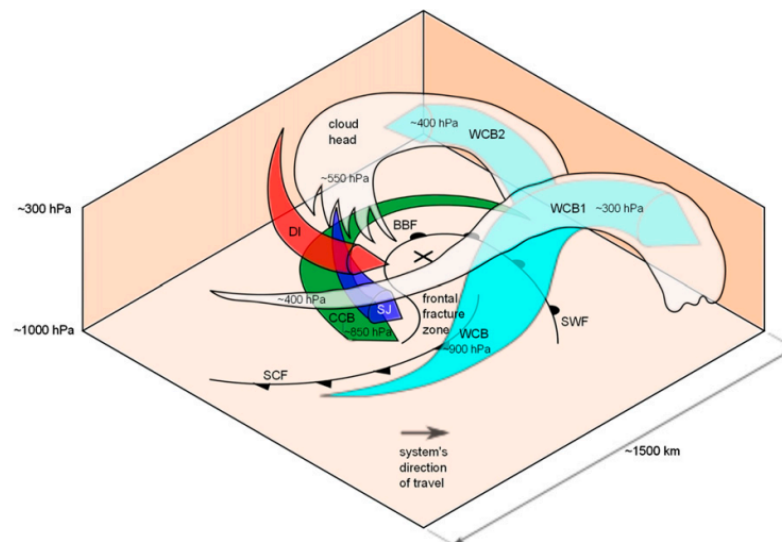


Figure 2.11: Schematic showing the typical structure of an extratropical cyclone in the northern hemisphere during the development stage (Fig. 2.8 b iv). For the surface, the schematic depicts the Surface Warm Front (SWF); Surface Cold Front (SCF); Bent-Back Front (BBF); and the Cold Conveyor Belt (CCB). The two ascending WCB air streams are depicted for the cyclonic branch (WCB1) and anticyclonic branch (WCB2). The Dry Intrusion (DI) is depicted in red, the Sting-Jet (SJ) in blue. Figure taken from Martínez-Alvarado et al. (2014).

Using Lagrangian trajectory data obtained from the ERA-Interim dataset, Madonna et al. (2014) conducted a comprehensive, 32 year global climatology for WCBs. They determined that WCBs occur most frequently in the winter months of the respective hemispheres. In the northern hemisphere and in winter, trajectories mostly begin their τ_{600} ascent over the open western Atlantic and Pacific Oceans, and 48 hours later they sprawl further north and cover a more broad area that includes the continents (Fig. 2.12). Madonna et al. (2014) determined that generally, during a WCB ascent, air parcels convert most of their initial moisture ($\approx 10\text{g kg}^{-1}$) into condensate which precipitates out of the parcels, leaving the specific humidity at the

end of the ascent smaller $\sim 0.1 \text{ g kg}^{-1}$. Due to the diabatic heating experienced during this process, the potential temperature increases dramatically from approximately 290 K on average at the start of the ascent, to approximately 315 K on average at the end (Madonna et al., 2014).

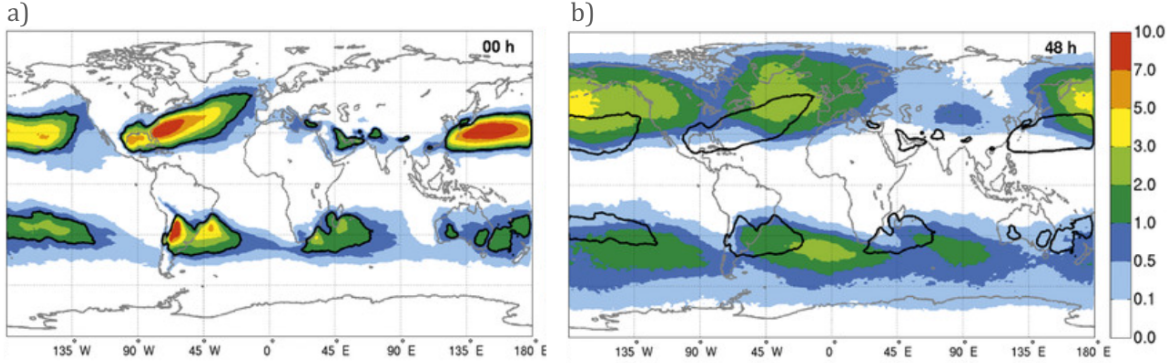


Figure 2.12: 32 year climatology of WCB trajectories, showing frequency of occurrence for the beginning of the τ_{600} ascent (a) and 48 h afterwards (b). Figure taken from Madonna et al. (2014).

In experimental studies that examined the vertical structure and microphysical properties of two WCB case studies (Gehring et al., 2020; Binder et al., 2020), it was determined that WCB air masses are generally part of a deep and strongly precipitating cloud layer, with cloud tops located at approximately 10 km altitude (or 300 hPa, see Fig. 2.13). Within this cloud layer, supercooled liquid water (SLW) exists at concentrations exceeding 0.2 g kg^{-1} along a $\sim 3 \text{ km}$ deep stretch along with ice, resulting in a mixed-phase cloud (Fig. 2.13). High vertical velocities and SLW concentrations are understood to lead to intense riming (see Section 2.3 for details on cloud microphysics). Strong horizontal winds of up to 45 m s^{-1} at 6.5 km above sea level induce a strong vertical wind shear, which increases turbulence and therefore the probability of particle collisions, leading to enhanced aggregation (Gehring et al., 2020).

In the WCB outflow above $\sim 7 \text{ km}$, WCB air masses were found to be located within a $\sim 3 \text{ km}$ deep cirrus layer (Binder et al., 2020). This accompanying "cirrus shield" is common for WCBs (Madonna et al., 2014; Spichtinger et al., 2005; Wernli et al., 2016; Binder et al., 2020). Within the ascending WCB air, liquid cloud droplets freeze and, due to the high vertical velocities, remain suspended up to high altitudes. This process produces "liquid origin" cirrus clouds (Krämer et al., 2016; Lüttmer et al., 2025). The air above the WCB is pushed upward and lifted, which can create supersaturated conditions over ice (Spichtinger et al., 2005). This process can make ice nucleation directly from the vapor phase possible (Wernli et al., 2016), leading to the production of "in situ" cirrus clouds (Krämer et al., 2016).

Given what we have learned about WCBs, it comes as no surprise that they are not only the most prominent feature of ETCs, but also their most influential component. This is because many of the key impacts that ETCs exert on atmospheric dynamics and weather (extreme precipitation, transport of energy and moisture, etc.) can be traced directly to WCBs. In the second part of the comprehensive climatology from Madonna et al. (2014), Pfahl et al. (2014) determined that the ascending WCB air produces large amounts of precipitation, contributing to 80% of extreme precipitation events in the extratropics (Pfahl et al., 2014). The large amounts of diabatic heating experienced during the ascent can produce potential vorticity

(PV) anomalies (such as PV towers) throughout the tropospheric column (Rossa et al., 2000; Madonna et al., 2014) and in the upper levels of the atmosphere (Oertel et al., 2020).

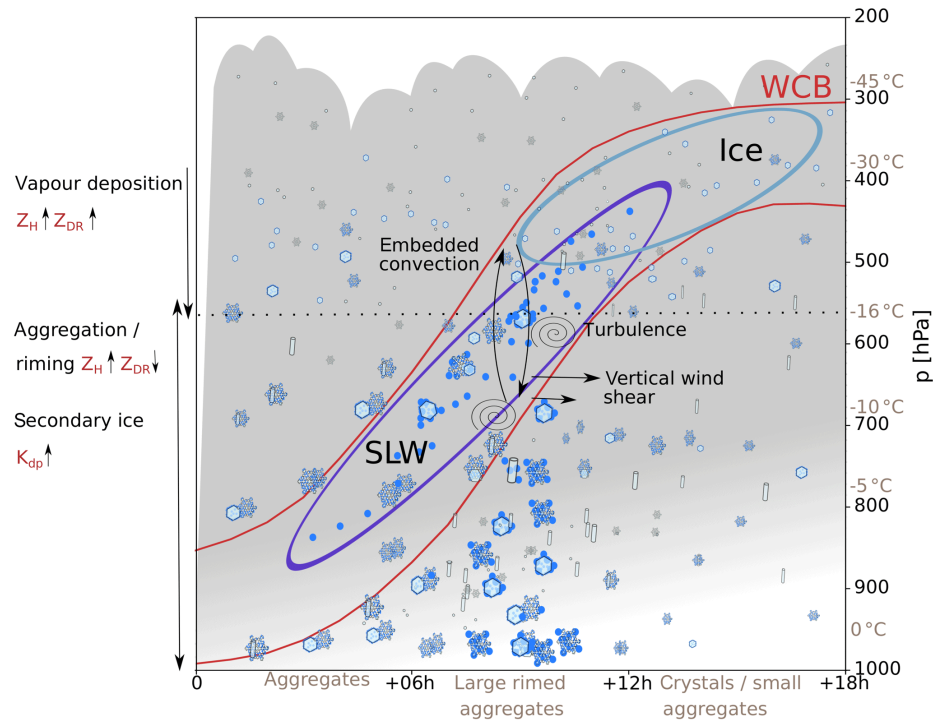


Figure 2.13: Schematic that summarizes the microphysical processes determined to take place in a WCB case study analyzed by Gehring et al. (2020). The frame of reference is Lagrangian, with the x-axis showing time. Figure is taken from Gehring et al. (2020).

This process has far reaching consequences: the production of PV at lower levels of the atmosphere due to latent heating in WCB clouds has been found to significantly contribute to the intensification of cyclones (Binder et al., 2016), and the PV anomalies induced in the upper levels of the atmosphere have been found to create circulation anomalies that affect synoptic-scale weather patterns (Martínez-Alvarado et al., 2015; Oertel et al., 2020). Therefore, the correct representation of WCBs in forecasting models is understood to be crucial for accurately predicting the weather (Pfahl et al., 2014). This is supported by multiple studies that have identified the incorrect representation of WCBs as a key factor in amplifying forecast uncertainties (Berman and Torn, 2019, 2022; Pickl et al., 2022).

However, WCBs are not only important for weather forecasting, but also for modeling the climate. This is mainly due to the fact that the clouds produced by WCBs interact with radiation (Joos, 2019). In the early stages of their lifetime, WCBs produce low-level liquid clouds and are typically located close to the equator and over the open ocean (Fig. 2.12 a; Madonna et al. (2014)). These clouds are highly reflective, and the amount of incident solar radiation at these latitudes is high, meaning that they reflect a large amount of shortwave radiation back into space (Joos, 2019). Additionally, the underlying ocean surface has a low albedo, meaning that without these clouds, the Earth would absorb much of the incoming radiation. Therefore, during this stage of their life cycle, WCBs generally have a net cooling effect on the Earth's climate, also known as negative "net cloud radiative forcing" (NetCRF;

Fig. 2.14; Ramanathan et al. (1989); Joos (2019)).

In contrast, during the later stages of their lifetime, WCBs are typically located further towards the poles and occasionally over land or ice (Fig. 2.12 b; Madonna et al. (2014)). Here, incident solar radiation is small, especially in winter. The cloud tops (and the accompanying cirrus shield) are composed of ice and are located at high altitudes, resulting in low cloud-top temperatures (Joos, 2019). Therefore, only little radiation is either reflected back into space or emitted by the cloud tops, while a lot of outgoing radiation is trapped (Ramanathan et al., 1989). For this reason, WCBs typically exhibit positive NetCRF during the later stages of their life cycle (Fig. 2.14). In between these stages, the net radiative effect from WCBs can be either positive, negative, or neutral (Joos, 2019).

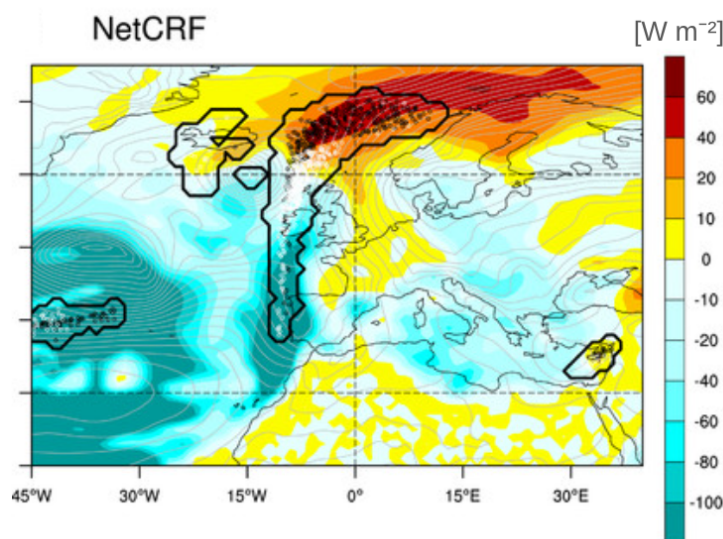


Figure 2.14: Snapshot of the net cloud radiative forcing (NetCRF) of a WCB case, outlined by the black contour. It shows negative NetCRF (a cooling effect) in the south and positive NetCRF (a warming effect) in the north. Figure taken from Joos (2019).

Another process that makes WCBs important for the climate is their transport of water vapor into the UTLS (Zahn et al., 2014), where it acts as a potent greenhouse gas (Schneider et al. (2010); Solomon et al. (2010); see Section 2.1.5). However, it is extremely difficult to quantify how much WCBs contribute to UTLS humidity, and what the net radiative effect of this process is. This difficulty is in part due to challenging measurement conditions in the UTLS (discussed in Section 2.1.5), but also because there exist large uncertainties in the model representation of WCBs. These uncertainties affect the precise determination of WCBs' overall NetCRF, as well the other aspects we have discussed so far, such as their role in extreme precipitation, cyclone intensification, and upper-level wave propagation.

One key reason for the fact that large uncertainties exist in the representation of WCBs in numerical weather prediction (NWP) models and global climate models (GCMs), is that these models use parameterizations to model microphysical processes in clouds (see Section 2.4.1). While these parameterizations are necessary, since fully modeling all cloud microphysical interactions is practically and computationally impossible, the problem is that they cannot fully

represent reality (Posselt and Vukicevic, 2010; van Lier-Walqui et al., 2012; Morrison et al., 2020). Studies have shown that the outcome of simulating WCB cases can change drastically when switching between microphysical parameterizations (Mazoyer et al., 2021, 2023). This is because WCBs undergo various warm-, mixed-, and cold-phase microphysical processes during their ascent (Forbes and Clark, 2003; Gehring et al., 2020), which each contribute differently to diabatic heating and PV production. Therefore, the modeled evolution of WCBs is highly sensitive to the specific calculations of these processes (cloud microphysical processes relevant to this thesis and their parameterization by the two-moment microphysics scheme from Seifert and Beheng (2005) are discussed in Sections 2.3 and 2.4.1, respectively).

Another large uncertainty in the model representation of WCBs stems from the fact that WCBs experience substantial convective activity. While the traditional view asserts that air in a WCB ascends gradually ("slantwise ascent"; Browning (1986)), studies in the past decade have clearly shown that deep convection is common in WCBs (Rasp et al., 2016; Blanchard et al., 2020; Oertel et al., 2020, 2021). This is significant because deep convection produces more extreme precipitation (Flaounas et al., 2017) and transports more tracers into higher regions of the atmosphere (Purvis et al., 2003) than slowly ascending air. Deep convection can also inject water vapor into the stratosphere and therefore influence the climate (see Section 2.1.5; Dauhut et al. (2018)). The uncertainty that comes from this fact is not entirely unrelated to that of microphysical parameterizations, since these parameterizations have been found to produce significant uncertainties specifically when modeling convective systems (Khain et al., 2015; Dearden et al., 2016; Barthlott et al., 2022; Hieronymus et al., 2022). Obviously, this also affects convectively active WCBs. For example, Oertel et al. (2023) determined that WCB air parcels which ascend quickly experience different microphysical processes during their ascent than slowly ascending air parcels, which adds another layer of uncertainty to the microphysical processes experienced by WCBs overall.

Finally, another uncertainty arises from simulation resolution. The study from Oertel et al. (2023) used a simulation with a resolution high enough to explicitly resolve convection, which allowed for the use of a state-of-the-art microphysics parameterization scheme everywhere (Oertel et al., 2023). However, most NWP models and CGMs operate on relatively coarse resolution grids and therefore depend on convection parameterization schemes to model convective activity (see Section 2.4.2). This introduces yet another source of uncertainty, because these schemes often only calculate basic microphysical processes (their primary function is to redistribute water vapor when convection is triggered, not to represent clouds; Tiedtke (1989); Bechtold et al. (2008)). Therefore, common convection parameterization schemes might inadequately represent the microphysical processes experienced by WCBs. A study by Choudhary and Voigt (2022) hints at this, since they determined that WCB air parcels ascend higher and faster, and experience different amounts of diabatic heating when simulation resolution is increased, especially when transitioning from convection parameterizing to convection permitting resolutions.

In this section, we introduced the current state of WCB research as it relates to this thesis and addressed the limitations of existing knowledge. This brings us to the knowledge gaps and research questions that this thesis aims to address.

2.2 Knowledge Gaps and Research Questions

In this thesis, we are primarily concerned with how WCBs transport water vapor and hydrometeors into the UTLS region. The first knowledge gap in this regard that this thesis aims to address concerns the role of convection. As mentioned in the previous section, convective activity is clearly present in WCBs and depending on their convective characteristics, WCB air parcels undergo different microphysical processes related to diabatic heating and hydrometeor production (Choudhary and Voigt, 2022; Oertel et al., 2023). However, it remains unclear (i) which microphysical processes govern the transport and removal of moisture by WCBs overall, and (ii) how these processes differ between convective and non-convective air parcels. Therefore, the first research question is:

Research Question I

Which microphysical processes control WCB moisture transport into the UTLS, and how do these processes differ between regions of convective and non-convective activity?

This research question is addressed in Chapter 3, which appears as a publication in *Atmospheric Chemistry & Physics*, titled "The role of ascent timescales for warm conveyor belt (WCB) moisture transport into the upper troposphere and lower stratosphere (UTLS)" (Schwenk and Miltenberger, 2024).

The second knowledge gap, which naturally follows the first, is that it is unclear how the transport of moisture by WCBs is modeled differently when convection is explicitly resolved in a simulation, versus when it is parameterized and only crude microphysical calculations are performed. It is also unclear whether the crude calculations by convection parameterization schemes are inherently "bad" or worse than convection permitting simulations at representing reality. Therefore, our two-part second research question is:

Research Question II a

How do the modeled moisture transport, the governing microphysical processes, and the radiative effects of WCBs differ when transitioning from a convection-parameterizing simulation to a convection-permitting one?

Research Question II b

Are convection-permitting simulations better at simulating UTLS moisture in the outflow of a WCB than convection-parameterizing simulations?

This research question is addressed in a three-part study, which we present in Chapters 4, 5 and 6. The first two parts address research question II a and the third addresses II b. These publications have been submitted to *Atmospheric Chemistry & Physics*.

The third knowledge gap relates to the uncertainties stemming from microphysical parameterizations, as discussed in the previous section. Microphysical parameterization schemes require fixed choices for various physical constants, such as the capacitance of ice (see Section 2.4.1). However, these constants have natural uncertainty and variability in reality. It remains entirely unclear how changing these constants or other fixed parameters within physically plausible ranges influences the modeled transport of moisture to the UTLS by WCBs. Therefore, our third research question, which partially ties back to the first, is:

Research Question III

In what ways do perturbations to microphysical parameters affect modeled moisture conditions in the outflow of WCBs, and how does this effect vary with ascent timescale?

This research question is addressed in Chapter 7, which appears as a publication in *Atmospheric Chemistry & Physics*, titled "Microphysical Parameter Choices Modulate Ice Content and Relative Humidity in the Outflow of a Warm Conveyor Belt" (Schwenk et al., 2025). Chronologically, the publications listed appeared in a different order than they presented here. However, in this thesis, the chapters are ordered according to the research questions, not the order of appearance.

The remaining sections of the introduction provide an overview of the theory on cloud microphysical processes relevant to this thesis (Section 2.3) and introduce the NWP model used in this study (Section 2.4). They also provide information on the online trajectory module (Section 2.4.3) that produced most of the analyzed data, and the microphysical (Section 2.4.1) as well as the convection parameterization scheme (Section 2.4.2) implemented in the model. Finally, an overview of the results is presented (Section 2.5).

2.3 Theory: Thermodynamics and Cloud Physics

In this section, we provide an overview of physical theory on thermodynamics and clouds as it relates to this thesis. This vast field of atmospheric science allows us to present only a tiny portion of it; for more information the interested reader is directed to the books "Microphysics of Clouds and Precipitation" by Pruppacher and Klett (2010) and "An Introduction to Clouds" by Lohmann et al. (2016), which form the basis for this chapter.

2.3.1 Relevant Thermodynamic Variables

Specific Humidity — q_v

Water is an important component of the atmosphere that, unlike most other components, regularly appears in gaseous, liquid, and solid states. The most dominant form is gaseous and is referred to as "water vapor." The amount of water vapor in an air parcel can be expressed in many different ways, depending on the process that is addressed. One common quantity which is used in this thesis is specific humidity (q_v), which is the mass of water vapor in an air parcel relative to the humid air in this air parcel, expressed in g kg^{-1} .

When air is allowed to exchange water molecules freely with a flat surface of pure liquid water (or ice) until thermodynamic equilibrium is reached, two closely related "saturation" quantities emerge: the saturation vapor pressure and the saturation specific humidity.

Saturation Vapor Pressure — $e_{\text{sat,w/i}}(T)$

At equilibrium, the water-vapor molecules in the gas-phase exert a partial pressure on the surface that exactly balances the flux of molecules returning to the condensed phase. That equilibrium pressure is called the "saturation vapor pressure over water / ice" ($e_{\text{sat,w/i}}$), and it increases with temperature (Fig. 2.15) because when the molecules in the condensed phase

have more kinetic energy, a larger vapor pressure is required to keep them from escaping into the gas phase. Due to the lattice structure of ice, water molecules are more tightly bound in ice than in water, which is why higher partial pressures (larger $e_{\text{sat},w/i}$) are required over water than ice at the same temperature (Fig. 2.15 b). The temperature dependencies of $e_{\text{sat},w/i}$ are governed by the Clausius Clapeyron Relation:

$$\frac{de_{\text{sat},w}(T)}{dT} = \frac{L_v e_{\text{sat},w}}{R_v T^2}, \quad \frac{de_{\text{sat},i}(T)}{dT} = \frac{L_s e_{\text{sat},i}}{R_v T^2} \quad (2.1)$$

with L_v the latent heat of vaporization, L_s the latent heat of sublimation, and R_v the ideal gas constant for water vapor. If $L_{v/s}$ are assumed to be constant with temperature, integrating this function produces the famous *Magnus Formula*. If instead $L_{v/s}$ are assumed to be a function of T , integration gives a lengthy expression that depends on the assumed function, with coefficients that can be determined experimentally. In this thesis, we use the empirical functions presented in Murphy and Koop (2005).

Saturation Specific Humidity — $q_{v,\text{sat},w/i}(T, p)$

For an air parcel of total pressure p , the specific humidity that corresponds to e_{sat} is

$$q_{v,\text{sat},w/i}(T, p) = \frac{\epsilon \cdot e_{\text{sat},w/i}(T)}{p - (1 - \epsilon) \cdot e_{\text{sat},w/i}(T)} \approx \frac{\epsilon \cdot e_{\text{sat},w/i}(T)}{p}, \quad (2.2)$$

where $\epsilon = R_d/R_v \approx 0.622$ is the ratio of the gas constants for dry air and water vapor. Hence, $q_{v,\text{sat},w/i}$ increases with temperature and decreases with pressure. These two saturated quantities are used to define an important and unit-less variable that describes the water vapor conditions in an air parcel with respect to thermodynamic equilibrium.

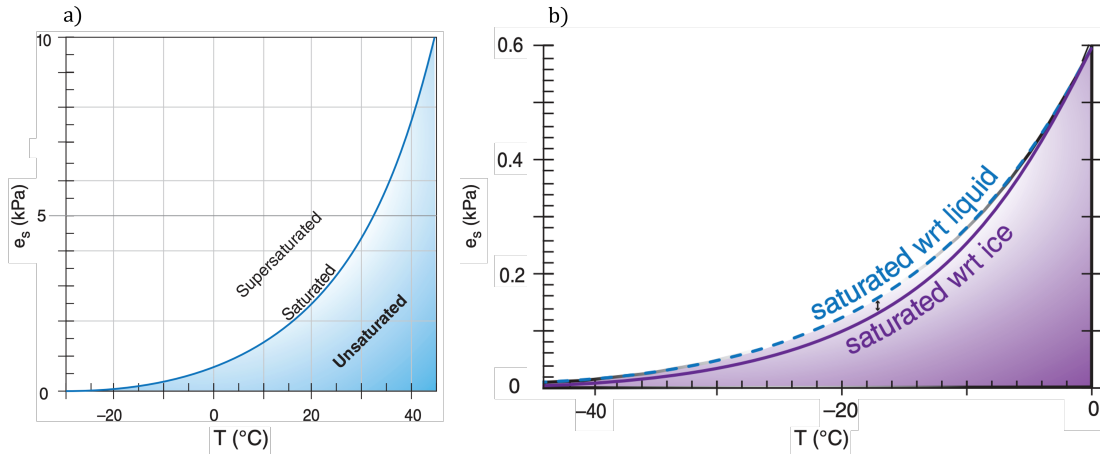


Figure 2.15: $e_{\text{sat},w/i}$ over T for only water (a) and for both water and ice (b) over temperature ranges typically found in the atmosphere. The small differences between $e_{\text{sat},w}$ and $e_{\text{sat},i}$ at identical temperatures, indicated by the small arrow in b), is what leads to the Wegener-Bergeron-Findeisen (WBF) process. Figure taken from Stull (2015b).

Relative Humidity — $RH_{w/i}(T)$

The relative humidity over water / ice ($RH_{w/i}$) is a commonly used variable that describes the percentage of water currently in air, with respect to how much water the air would contain

in thermodynamic equilibrium. It is given by:

$$RH_{w/i} = 100\% \cdot \frac{e_{\text{sat},w/i}(T)}{e} = 100\% \cdot \frac{q_{v,\text{sat},w/i}(T, p)}{q_v}, \quad (2.3)$$

with e the vapor pressure of water vapor. Often times, instead of $RH_{w/i}$, the supersaturation $S_{w/i} = (100\% - RH_{w/i})/100$ is used, which is 1 when $RH_{w/i} = 100\%$, smaller than 1 when $RH_{w/i} < 100\%$, and larger than 1 under when $RH_{w/i} > 100\%$.

In the atmosphere, $RH_{w/i}$ can exceed 100%, producing "supersaturated" conditions. If the air contains liquid water or ice in this case, the water vapor transfers from the gaseous phase to the condensed phase. This releases latent heat into the air, increasing the temperature and buoyancy. Conversely, if $RH_{w/i}$ is below 100% ("subsaturated conditions"), any condensed water will evaporate or sublimate, and the latent heat required for these processes will be taken from the environment, decreasing the air temperature.

The relative humidity of an air parcel can be increased by cooling it down (since then $e_{\text{sat},w/i}$ decreases), which can be achieved by lifting the air parcel, since temperature decreases with altitude in the troposphere (see Section 2.1). Therefore, liquid or ice particles suspended in the air generally grow when the air ascends and evaporate when it descends. For this process to occur, however, there must be some amount of liquid or frozen water in the air. If not, the air can remain in a subsaturated or supersaturated state. This brings us to the production of liquid or ice particles from water vapor.

2.3.2 Formation of Hydrometeors

A "hydrometeor" is a particle consisting of liquid or solid water suspended in or falling through the atmosphere. Commonly, one defines six main types of hydrometeors, which are distinguished by their phase, size, and shape: cloud droplets (c), raindrops (r), ice crystals (i), snowflakes (s), graupel (g), and hail (h ; Tab. 2.1). In this thesis, concentrations of hydrometeor populations are presented using mass mixing ratios (q_x) in g kg^{-1} , and number mixing ratios (N_x) in kg^{-1} . Cloud droplets and ice crystals are the two hydrometeor species that can initially form in an air parcel containing no hydrometeors. In this section, we discuss the mechanisms for their formation. In the section that follows immediately after, we summarize the microphysical processes that allow for the growth and formation of the remaining species.

Species	Symbol	Phase	Size parameter	Typical range [mm]
Cloud droplets	c	Liquid	Radius	0.002 – 0.25
Raindrops	r	Liquid	Radius	0.25 – 5
Ice crystals	i	Frozen	Length	0.001 – 1.5
Snowflakes	s	Frozen	Max. dim.	0.1 – 100
Graupel	g	Frozen	Radius	< 2.5
Hail	h	Frozen	Radius	> 2.5

Table 2.1: Overview of the six main hydrometeor species, listing their symbols and summarizing their typical size characteristics. Adapted from Lohmann et al. (2016).

Formation of Cloud Droplets

The only way to produce cloud droplets from air that consists purely of gas is through homogeneous nucleation. For an air parcel at constant pressure, temperature, and with a large amount of water vapor producing the vapor pressure $e > e_{\text{sat},w}$, the Gibbs free energy G of the purely vapor state is higher than the condensed state. Therefore, $\Delta G < 0$ and the system would prefer to produce condensed water. However, there exists an energy barrier for this process, because work is needed to produce an initial microscopic "cluster" of water molecules. Therefore, this process can only occur spontaneously when $|\Delta G|$ is very large, which in Earth's atmosphere can only happen when $RH_w \gg 100\%$. As is discussed next, when aerosols are present this amount can be greatly reduced, and since the atmosphere contains many aerosols, homogeneous nucleation of water droplets does not occur often in Earth's atmosphere.

The atmosphere contains not only gases, but also many solid and liquid particles called "aerosols." These particles can range in size from a few nanometers (sulfuric acid, soot) to many micrometers (dust, sea salt) and exhibit a variety of properties. According to Raoult's law, if an aerosol is soluble and incorporated into a liquid water droplet, then the necessary vapor pressure to produce equilibrium is smaller than it would be for a droplet of pure water.

This can be understood heuristically by visualizing how the molecules of the dissolved aerosol on the surface of the droplet reduce the number of sites through which water molecules can escape into the gas phase, which reduces the vapor pressure necessary to keep them in the liquid phase. Therefore, the incorporation of a soluble aerosol creates a lower energy barrier for the transition from gas to liquid, enabling the formation of cloud droplets at RH_w only a few percent over 100%. The exact RH_w necessary for droplet formation depends on the size and exact properties of the aerosols, according to Köhler theory. The process described here is simplified, which is sufficient for the discussions in this thesis. For more in-depth theory, see Chapter 6 in Lohmann et al. (2016).

Aerosols that enable heterogeneous droplet formation are called "cloud condensation nuclei" (CCN). In the troposphere and at remote, unpolluted sites over the open ocean or in the arctic regions, CCN concentrations are usually on the order of $\sim 10^2 \text{ cm}^{-3}$ (Bigg and Leck, 2001; Hudson and Noble, 2009; Yu and Luo, 2009). Over non-polluted continental areas such as forests this increases to $\sim 10^3 \text{ cm}^{-3}$ (Sihto et al., 2011), and can exceed $\sim 10^5 \text{ cm}^{-3}$ over polluted, urban areas (Choudhury and Tesche, 2023; Block et al., 2024).

Ice Nucleation

Similarly to cloud droplets, an ice crystal can theoretically form directly from the vapor phase without the help of an aerosol particle. However, compared to water, this requires RH values of more than 1000%, which is why this process does not occur in nature. Instead, the energy barrier for transitioning from a pure liquid or solution droplet to an ice crystal is small enough that this process, known as "homogeneous freezing of cloud droplets," occurs in nature (Fig. 2.16, bottom row). Due to the energy barrier, this process cannot occur directly below 0°C , but only below -38°C , known as the "homogeneous freezing temperature." Below this temperature, clouds consist purely of ice. Homogeneous freezing is a prominent pathway for the formation of ice crystals in cirrus clouds.

All other ice crystal formation pathways require specific types of aerosols, called "ice nucleating particles" (INPs). The conditions for ice crystal formation are more stringent than those for liquid droplet formation. Therefore, far fewer aerosols are INPs than CCNs. While an atmospheric CCN concentration of 10^2 cm^{-3} is considered small, INP concentrations in the troposphere are usually on the order of 10^{-6} cm^{-3} and generally never larger than 10^{-2} cm^{-3} (Wex et al., 2019; Herbert et al., 2025). It is also much more difficult to determine whether a specific aerosol can act as an INP or not, and forecasting INP concentrations remains an elusive task. It is important to note that ice particles themselves act as INPs.

In the presence of INPs, ice crystals can form directly from the vapor phase. This process is called "deposition nucleation" (Fig. 2.16, second row from the bottom). Deposition nucleation is not of primary importance for most clouds, and it also remains unclear exactly which conditions give rise to it. However, it is clear that this process can occur at temperatures above -38°C .

Cloud droplets can freeze at temperatures larger than -38°C if they contain an INP. This process is called "immersion freezing of cloud / solution droplets" (Fig. 2.16 third row from the top). If a CCN also acts as an INP, it can contain some liquid water before reaching $RH_w = 100\%$ and freeze before producing a cloud droplet in a process called "condensation freezing" (Fig. 2.16 second row from the top). However, this is a relatively rare event. Finally, in a process called "contact freezing", an INP might collide with a supercooled cloud droplet, which then freezes (Fig. 2.16, top row).

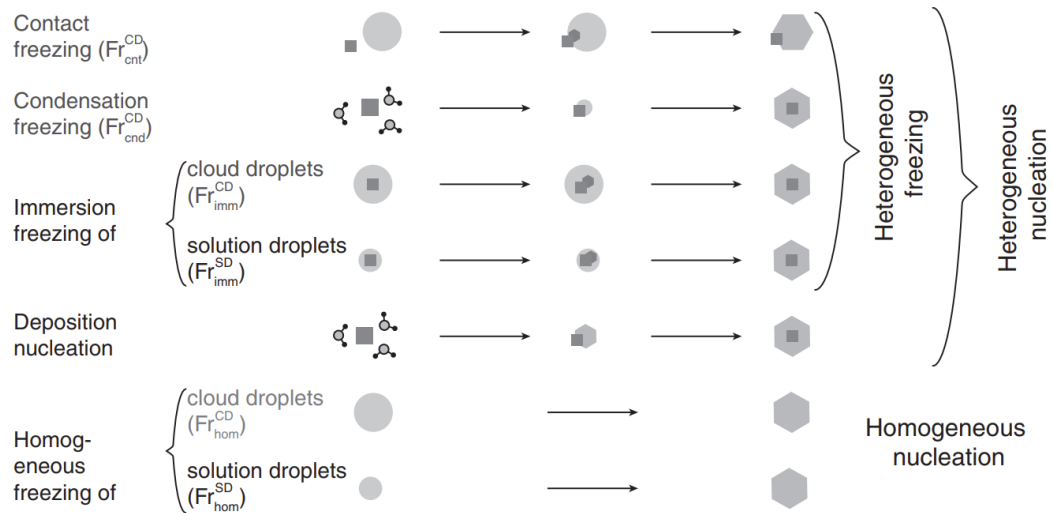


Figure 2.16: Schematic showing the formation pathways of ice in clouds. Figure taken from Lohmann et al. (2016).

2.3.3 Microphysical Processes in Clouds

We now understand how air that is supersaturated with respect to water or ice can form cloud droplets and ice crystals. We have also seen that liquid water can remain in an air parcel even at temperatures well below the freezing point. Since there are so few INPs that can facilitate ice production, this is common in Earth's clouds. Therefore, clouds can exist in a purely liquid

state above 0 °C, in a "mixed-phase" state for temperatures between 0 °C and -38 °C, and in a purely ice phase below -38 °C. In WCBs, mixed-phase clouds are very prominent (Fig. 2.13) and therefore relevant for this thesis. In this section, we will summarize the processes that allow hydrometeors to grow, break-up and transform within liquid, mixed-phase, and ice clouds. A visual overview of these processes is presented in Figure 2.18.

Droplet Growth by Condensation

If a cloud droplet is suspended in air which is supersaturated with respect to water, then water molecules will transfer from the gas phase into the liquid phase until saturated conditions are met. Consequently, the cloud droplet radius will increase with time. This is a non-linear process, since the latent heat released during condensation increases the temperature of the air at the boundary of the droplet, which reduces $e_{\text{sat,w}}$. Additionally, cloud droplet radii in a realistic air parcel will follow a distribution because CCN sizes also vary. Therefore, due to different saturation conditions at the boundary stemming from the Laplace pressure, cloud droplets will grow at different rates. In a typical cloud, this mechanism allows cloud droplets to grow from approximately $\sim 2 \mu\text{m}$ to $\sim 15 \mu\text{m}$ in about 10 minutes. However, to reach raindrop radii through condensational growth alone would require more time than is observed in the atmosphere. Therefore, other mechanisms must contribute to cloud droplet growth as well.

Droplet Growth by Collision–Coalescence

The process of collision-coalescence, also known as "accretion," describes the collision and sticking together of cloud and rain droplets. It occurs once cloud droplet radii exceed $\sim 15 \mu\text{m}$ and is relatively efficient once it is initiated; approximately 10^5 collisions are required to produce a raindrop with a radius of 1 mm. What exactly initiates the collision-coalescence process remains an active topic of research. One theory is that giant CCNs (GCCNs), such as coarse-grained sea salt particles, produce larger cloud droplets with greater terminal velocities. These droplets fall through the air parcel while smaller droplets remain suspended and collide with them. If GCCNs are not present, turbulent motion is required to enhance the probability that cloud droplets collide. Although it is not technically a growth mechanism, we note that upon colliding, two liquid drops may "shatter" instead of sticking together, producing many smaller cloud droplets.

Ice Crystal Growth by Deposition

Similarly to the growth of cloud droplets through condensation under supersaturated conditions with respect to water, ice crystals grow through the deposition of vapor under supersaturated conditions with respect to ice. Like the condensational growth of cloud droplets, ice crystal growth by deposition is a nonlinear process involving latent heat release and changes in $e_{\text{sat,i}}$. The rate of growth for the mass of an ice crystal with time can be expressed as:

$$\frac{dm_i}{dt} = \alpha_m \frac{4\pi C (S_i - 1)}{F_k^i + F_d^i}. \quad (2.4)$$

In this equation, m_i denotes the mass of the ice-crystal, S_i the supersaturation over ice, and C the capacitance. F_k^i and F_d^i are temperature dependent thermodynamic variables: F_k^i includes the thermal conductivity, latent heat of sublimation, temperature, ideal gas constant for water vapor, and air density; F_d^i includes the saturation vapor pressure over ice, diffusivity of water

vapor, temperature, ideal gas constant for water vapor, and air density. The capacitance C (also "CAP" in Chapter 7) depends on the shape of ice crystals. For a sphere, $C = r$, whereas for a disk $C = 2r/\pi$. When C increases, so does the rate of mass growth.

Higher supersaturation values are required for ice nucleation than for cloud droplet nucleation. Therefore, an initially formed ice crystal will encounter larger values of S_i than S_w for cloud droplets. This results in a faster diffusion of water vapor and growth from depositional growth than from condensational growth. Depending on the degree of supersaturation over ice and the temperature, ice crystals will grow into different shapes and forms (Fig. 2.17), each with different values for C .

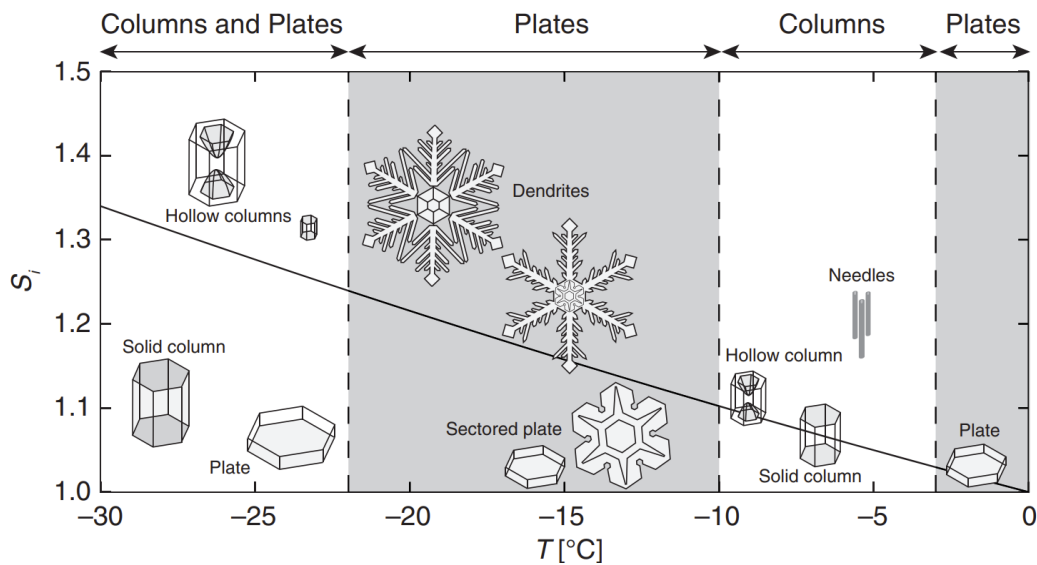


Figure 2.17: Schematic showing the different types of ice crystals and snowflakes that form under varying supersaturation values over ice (S_i) and temperature. Figure taken from Lohmann et al. (2016).

Frozen Hydrometeor Growth by Collisions

Frozen hydrometeors can grow not only by colliding with other frozen hydrometeors, but also by colliding with supercooled liquid drops. Therefore, there are more pathways than just collision-coalescence for the growth of frozen particles. The growth of ice particles through collisions with other ice particles is called "aggregation"; the collision with supercooled cloud droplets or raindrops is called "riming." As we will see in this thesis, riming is an important mechanism for the production of graupel and hail in the mixed-phase clouds of WCBs. Ice particles can shatter upon collision, producing smaller "secondary ice" particles that can act as INPs and facilitate glaciation in mixed-phase clouds by enhancing contact freezing. Technically, this latter process does not count as growth; instead, it increases the number of ice particles (and therefore the overall mass of ice).

The Wegener-Bergeron-Findeisen Process

Because the necessary saturation vapor pressure to maintain equilibrium is higher over water than over ice, a mixed-phase cloud may experience conditions in which frozen hydrometeors are supersaturated, while liquid hydrometeors are subsaturated. In this case, the liquid

droplets shrink due to evaporation, while the frozen particles grow due to deposition. The net result is a transfer of water from the liquid phase to the ice phase via the vapor phase. This process is called the "Wegener-Bergeron-Findeisen" (WBF) process. If an air parcel's ascent velocity and the size, mass, and species distributions of the hydrometeors are just right, then this condition may be experienced for a prolonged time. As it happens, these conditions are regularly met in Earth's atmosphere, making the WBF process an important mechanism for the glaciation of clouds.

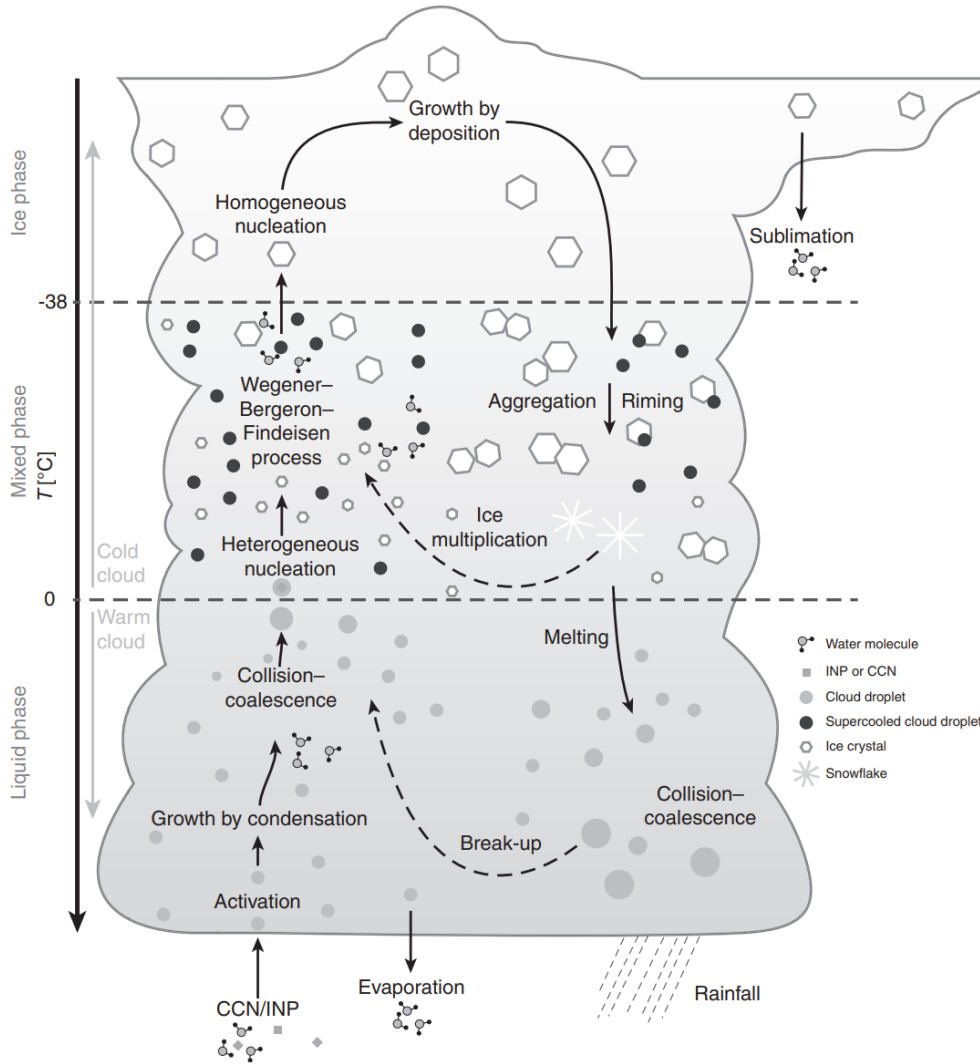


Figure 2.18: A visual summary of the microphysical processes that govern hydrometeor growth, transformation, and destruction in liquid, mixed-phase, and ice clouds. Figure taken from Lohmann et al. (2016).

2.4 The ICON Modelling Framework

The Icosahedral Nonhydrostatic (ICON) Modeling Framework was developed by the German Weather Service (DWD) and the Max Planck Institute for Meteorology (MPI-M). It allows for simulations of the Earth's atmosphere at all relevant spatial resolutions for numerical weather prediction (NWP). A unique and name-giving feature of ICON is that its calculations

are preformed on an icosahedral-triangular Arakawa C grid, instead of a regular longitude-latitude grid (Fig. 2.19 a; (Prill et al., 2020)), which allows for consistent grid spacing globally. ICON also does not assume hydrostatic conditions, but instead solves the full compressible non-hydrostatic equations of motion for the atmosphere (Zängl, 2014). Another important and useful feature in ICON is "grid-nesting", which allows for the seamless incorporation of smaller yet higher-resolution domains within lower-resolution domains (Fig. 2.19 a and b). Through telescoping, one can place multiple "nests" into each other, with each nest utilizing a finer mesh in space and time with half the grid width of the "parent" domain (Prill et al., 2020; Zängl et al., 2022). Simulations on lower-resolution domains are coupled with the next highest-resolution simulation by providing lateral boundary conditions. In turn, they are nudged towards the solutions of the higher-resolution simulation (Fig. 2.19 b; Zängl et al. (2022)).

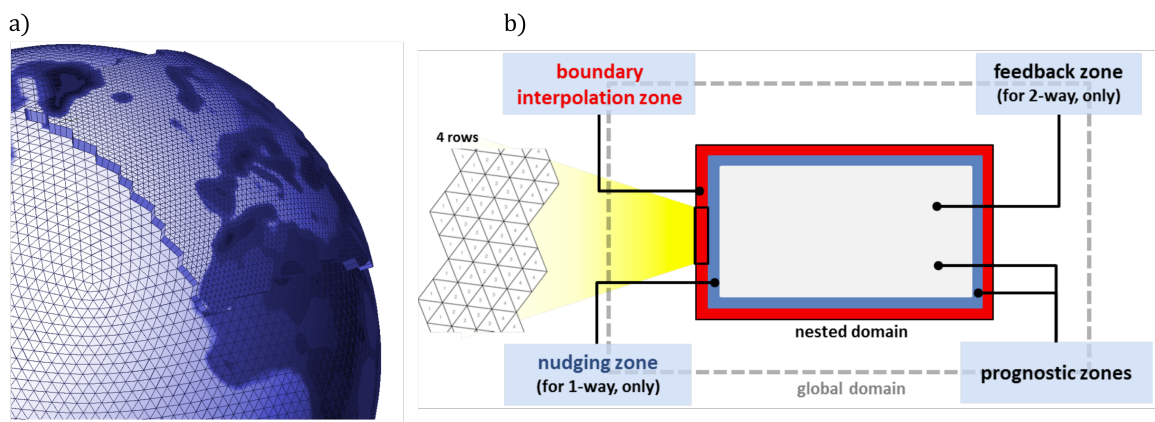


Figure 2.19: Image showing an example of the ICON grid, including a nested, higher resolution domain over Europe (a), and a schematic explaining how nested domains interact with the parent domains (b). This figure is taken from Prill et al. (2020).

In this thesis, we use the ICON version 2.6.2 from the DWD. We make use of the two-moment microphysics parameterization scheme from Seifert and Beheng (2005), and the convection parameterization scheme from Tiedtke (1989), which are discussed next. The ICON model produces Eulerian data. However, in this thesis, we make extensive use of Lagrangian data, which is made possible by the online trajectory module that was incorporated into the ICON model by Miltenberger (2014) and expanded to support grid-nesting by Oertel et al. (2023). Therefore, we finish this section by briefly introducing the online-trajectory module.

2.4.1 Two-Moment Microphysics Parameterization

In NWP models it is impossible to run computations for all of the complex microphysical processes and for all hydrometeors in a grid cell. One grid cell might represent $\sim 1 \text{ km}^2$ in very high resolution simulations, but there can be as many as 10^9 cloud droplets in a single kilogram of air. Therefore, these processes must be parameterized.

In a "two-moment" microphysical parameterization scheme, computations are performed on two integrals ("moments") of hydrometeor size or mass distributions, instead of performing calculations for every hydrometeor. For a given mass (m) distribution $\Gamma_x(m)$ for the

hydrometeor species x , the n -th moment of the distribution is defined as:

$$M_x^n = \int_0^\infty m^n \Gamma(m) dm. \quad (2.5)$$

The zeroth moment gives the number mixing ratio $M_x^0 = N_x$, and the first moment gives the mass mixing ratio $M_x^1 = q_x$. In the parameterization scheme by Seifert and Beheng (2005), which treats the six main hydrometeor species listed in Table 2.1, the mass distributions for all hydrometeors are assumed to be generalized Γ -distributions:

$$\Gamma_x(m) = A m^\mu \exp(-\lambda m^\mu). \quad (2.6)$$

The species-specific values for A , μ and λ , as well as mass thresholds for the categorization of species, are summarized in Table 1 in Seifert and Beheng (2005). In the liquid phase, the parameterization scheme treats cloud droplet nucleation by CCN, droplet growth by collision coalescence, droplet growth by condensation, collisional breakup of rain drops, evaporation of raindrops, and sedimentation. As long as there is liquid water present, if an air parcel experiences supersaturated or sub-saturated conditions, the "saturation adjustment scheme" immediately condenses or evaporates liquid water onto or from cloud or rain droplets to ensure $RH_w = 100\%$. The equilibration process in the presence of liquid in the real atmosphere is assumed to be fast enough that this immediate equilibration by the saturation adjustment scheme is adequate given the typical time resolution (~ 60 s) of an ICON simulation. However, in reality, when vertical velocities are high, supersaturated conditions can persist for an extended period of time, even in liquid clouds. The effects of this potential inadequacy are investigated in Chapter 7.

Whereas liquid drops are assumed to be spherical, in the frozen phase, the diameter-mass relationships are described by power laws in the form $D_x(m) \approx a m^b$, with a and b specific to the hydrometeor species. The parameterization scheme treats deposition nucleation, depositional growth with an equation similar to Eq. 2.4, heterogeneous freezing of cloud and raindrops, aggregation, riming, secondary ice production by "rime-splintering", as well as melting and sedimentation of ice particles. The capacitance C is assumed to be a fixed function of the diameter D and does not take into account that values for C can vary greatly for different shapes and sizes of ice particles. The effects of this potential inadequacy are also investigated in Chapter 7.

For a chronological breakdown of the order in which calculations and adjustments are performed in the implementation of the two-moment microphysics parameterization scheme in ICON, see the supplementary information of Chapter 3. In that summary, references are also given to the publications upon which individual calculations are based.

2.4.2 Convection Parameterization

If the horizontal and vertical grid spacings are sufficiently small, numerical calculations can resolve convective motion in the atmosphere. In this case, the simulation is referred to as "convection permitting." However, in the context of NWP, grid-spacings of about 3 km (and smaller) are necessary to explicitly resolve deep (and shallow) convection. Running such simulations with global coverage is computationally expensive and has only recently become feasible.

However, for climate models that run for hundreds or even thousands of simulation years, the computational cost remains too high. Therefore, many NWP model simulations are preformed using grid-spacings of about 15 km, meaning that they must parameterize convective processes.

In ICON, convection is parameterized by the Tiedtke-Bechtold scheme (Tiedtke, 1989; Bechtold et al., 2008; ECMWF, 2016) which covers deep, shallow, and mid-level convection. In this scheme, clouds are represented by a single pair of entraining/detraining plumes, which describe updraft and downdraft processes, as well as momentum and tracer transport. The scheme assumes equilibrium between large-scale and boundary-layer forcing, which generates convective available potential energy (CAPE), and convection, which reduces CAPE. The scheme establishes budget equations for heat, moisture, momentum, and chemical tracers that only consider zeroth moments, referred to as "bulk" equations.

The most important part of a convection parameterization scheme is the algorithm that determines when to trigger convection. In the ICON implementation of the Tiedtke-Bechtold scheme, computations are performed at every horizontal grid point and at each time step of the model to determine if the air contains sufficient CAPE and if the atmospheric column is sufficiently unstable for convection to occur. When convection is triggered, the calculations for the updrafts treat entrainment rates, turbulent detraining rates, and updraft kinetic energy. For downdrafts, the downward mass flux is assumed to be proportional to the upward mass flux from the updrafts, and turbulent as well as organized entrainment and detraining are calculated.

Cloud microphysical processes are calculated by the convection parameterization scheme, but not with the sophistication of the two-moment scheme by Seifert and Beheng (2005). Condensation rates are calculated using the updraft mass flux, and the fraction of hydrometeors in the ice phase is a simple function of temperature. Auto-conversion rates are assumed to be functions of the fraction of ice phase hydrometeors, which together with various threshold values for liquid and ice water mass mixing ratios are used to calculate precipitation rates. The WBF process is also parameterized as a function of cloud condensate and temperature. The scheme also calculates sedimentation rates, evaporation of rain, and the melting of ice particles. Whether this rudimentary calculation of microphysical processes produces inadequate representations for vertical transport of moisture and hydrometeors is investigated in Chapter 4 and 5.

2.4.3 Online Trajectories

The ICON model runs on an Eulerian grid and produces Eulerian output. However, it is possible to produce Lagrangian data from Eulerian data. Lagrangian data follows individual air parcels, and the data for each parcel over time is referred to as a "trajectory." Calculations which run on Eulerian output written to a file produce "offline trajectories." The accuracy of these calculations is limited by the spatial and temporal resolution of the written data. A common offline trajectory calculation tool is LAGRANTO (Sprenger and Wernli, 2015). Calculations which run on the native NWP model grid during a simulation produce "online trajectories" and are considered to be far more accurate.

Most of the analysis in this thesis is based on Lagrangian data obtained from online trajectory calculations. The advantages of using Lagrangian data is that one can track air as it moves with atmospheric winds as part of a phenomenon that one wants to study. In our case, we identify trajectories that move through a low-pressure system and ascend at least 600 hPa in no more than 48 hours as part of a WCB. These trajectories can then be investigated in detail to determine which microphysical processes occur at which pressure, temperature, or altitude.

The online trajectory module that was implemented in ICON for this thesis was developed by Miltenberger (2014) for the COSMO model, adapted for ICON by Miltenberger et al. (2020) and expanded by Oertel et al. (2023) to support grid-nesting. This module solves the trajectory equations $D\vec{x}/Dt = \vec{u}(\vec{x}, t)$ numerically using the Petterssen scheme (Petterssen, 1940) and the wind fields at each model time step. Trajectories are allowed to move from one domain to the other. Grid-scale variables, such as temperature, ice mass mixing ratios, or contributions to vapor from the turbulence scheme (turbulent "tendencies"), are interpolated onto trajectory positions from neighboring grid cells.

2.5 Overview of Results and Publications

This thesis consists of three studies, with the third study divided into three parts. I am the lead author of all the studies presented in this thesis. I analyzed the model output, interpreted the results, created the figures, wrote the text, and submitted and revised the papers in the review process. However, I am indebted to my co-authors for their significant contributions to conceptualizing the research questions, structuring the studies, interpreting the results, setting up and debugging ICON code, drafting the text, and ensuring its scientific accuracy. Author contributions are included in the publications and more detailed information is given below.

Each of the three studies addresses one of the three research questions presented in Section 2.2. These questions relate to the role of various factors in the transport of moisture into the upper troposphere/lower stratosphere (UTLS) by warm conveyor belts (WCBs). Therefore, the studies are described by which factor they examine.

The Role of Convection

The first study, presented in Chapter 3, appears as a publication in *Atmospheric Chemistry & Physics*, titled "The role of ascent timescales for warm conveyor belt (WCB) moisture transport into the upper troposphere and lower stratosphere (UTLS)". I am the lead author of this publication. I conducted the analysis, generated the figures, and wrote, submitted, and revised the manuscript. Annette Miltenberger set up, ran, and debugged the ICON simulation, guided the research, and supported the analysis of the results and drafting of the text.

In this study, we examine a WCB case study modeled by a convection permitting ICON simulation. Using Lagrangian online trajectory data, we analyze the microphysical processes that govern moisture transport and determine how these processes change when examining fast-ascending (convective) and slowly-ascending WCB trajectories. We find that the moisture content at the end of a WCB ascent shows a strong dependency on the ascent timescale.

Fast trajectories glaciate late and contain many supercooled liquid water droplets that freeze homogeneously, producing small ice crystals that remain suspended until the end of the ascent, producing cirrus clouds. In contrast, slowly ascending trajectories glaciate sooner, producing larger ice particles and rain that sediment out. This leaves a smaller hydrometeor content at the end of the ascent and produces fewer cirrus clouds. However, as a first-order approximation, the vapor content of air parcels at the end of their ascent is in equilibrium, i.e., it is controlled by temperature and pressure. Therefore, since slowly ascending trajectories end their ascent at warmer temperatures and pressures than fast-ascending trajectories, the vapor content at the end of the ascent is higher for slowly ascending trajectories.

The Role of Scale

The second study, which is divided into three parts presented in Chapters 4, 5 and 6, is in preparation to be submitted to *Atmospheric Chemistry & Physics*. The publications are titled "Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS) - Part I: Lagrangian Perspective" (Schwenk and Miltenberger, 2025a), "Part II: Eulerian Perspective" (Schwenk and Miltenberger, 2025b), and "Part III: Comparison with Observations" (Schwenk and Miltenberger, 2025c). I am the lead author of these studies. I conducted the analysis, developed the methods, generated the figures, and wrote the manuscript. I also set up and ran the RTTOV simulations. Annette Miltenberger set up and ran the ICON simulations, guided the research, and supported the analysis of the results and drafting of the text. Vanessa Santos provided the MSG satellite data.

In this three-part study, we examine the same WCB case as in the first study, but incorporate an additional ICON simulation run at a lower resolution that requires the parameterization of convection. In Part I, we use Lagrangian online trajectory data to compare the microphysical processes that govern the transport of moisture and the hydrometeor populations during the ascent between the two simulations. We find that, in the higher-resolution simulation, trajectories ascend faster. This results in a shift toward more frozen precipitation, as well as differences in the size and species distribution of hydrometeors during ascent. When trajectories are in the UTLS as part of the WCB outflow, the low-resolution simulation shows higher specific and relative humidities, as well as lower ice number mixing ratios, than the high-resolution simulation.

In Part II, we supplement Part I by conducting an Eulerian analysis of the two simulations. This is done because the Lagrangian data is not domain-filling, so important features of the WCBs may be missing, which could introduce biases into the results of Part I. Additionally, some variables, such as outgoing longwave radiation, cannot be analyzed from trajectory data. In this study, we present a new method for generating Eulerian WCB masks using trajectory data. Using these masks, we find that the WCB outflow is less humid in the higher-resolution simulation because higher ice number mixing ratios lead to faster supersaturation depletion. This, in turn, leads to more outgoing longwave radiation at the top of the atmosphere in the higher-resolution simulation. These results suggest that the microphysical calculations in the convection parameterization scheme are too crude, and result in too much vapor and too few hydrometeors being transported into the UTLS. This could lead to biases when determining the radiative forcing of WCBs in climate models.

The interpretation in Part II, which states that the convection parameterization transports "too much" vapor and "too few" hydrometeors, is based on the idea that the higher-resolution, convection-permitting simulation more accurately represents reality. To challenge this assumption, we present Part III, in which we develop and implement a new method to compare the model output from the two simulations to geostationary satellite data. In this method, ICON simulation data is transformed into synthetic satellite data via a radiative transfer model. Then, real and synthetic satellite products related to upper-level moisture are analyzed using object tracking and fractal analysis. The results suggest that patterns produced in higher-resolution simulations are more consistent with observations, supporting the idea that higher-resolution simulations are superior. However, both simulations equally fail to represent the advancing cold front in the southern regions of the WCB case, which shows that even the results from higher-resolution simulations cannot be considered completely representative of reality.

The Role of Parameter Perturbations

The third study, presented in Chapter 7, appears as a publication in *Atmospheric Chemistry & Physics* titled "Microphysical Parameter Choices Modulate Ice Content and Relative Humidity in the Outflow of a Warm Conveyor Belt" (Schwenk et al., 2025). I am the lead author of this publication. Annika Oertel set up and conducted the simulations for the PPE and provided the pre-selected Lagrangian WCB trajectory data. Annette Miltenberger and Annika Oertel developed the concept of this study. I conducted the analysis, developed the methods, generated the figures, and wrote, submitted, and revised the manuscript. Annika Oertel provided expertise, supported writing the text. Annette Miltenberger supported the analysis of the results and drafting of the text.

In this study, we examine a different WCB case than in the previous studies. This case was modeled by 70 convection permitting ICON simulations as part of a perturbed parameter ensemble set up by Oertel et al. (2025). In this ensemble, different parameters, constants, and calculations important for the representation of WCBs were perturbed within physically plausible ranges. We examine how these perturbations influence the modeled moisture content in the outflow of the WCB using Lagrangian online trajectories. We find that overall, the pressure, temperature, and specific humidity in the WCB outflow are not sensitive to parameter perturbations. However, for fast ascending trajectories, modifications to the saturation adjustment scheme, which allow for the production of supersaturated conditions over water when vertical velocities are high, are able to influence these three variables.

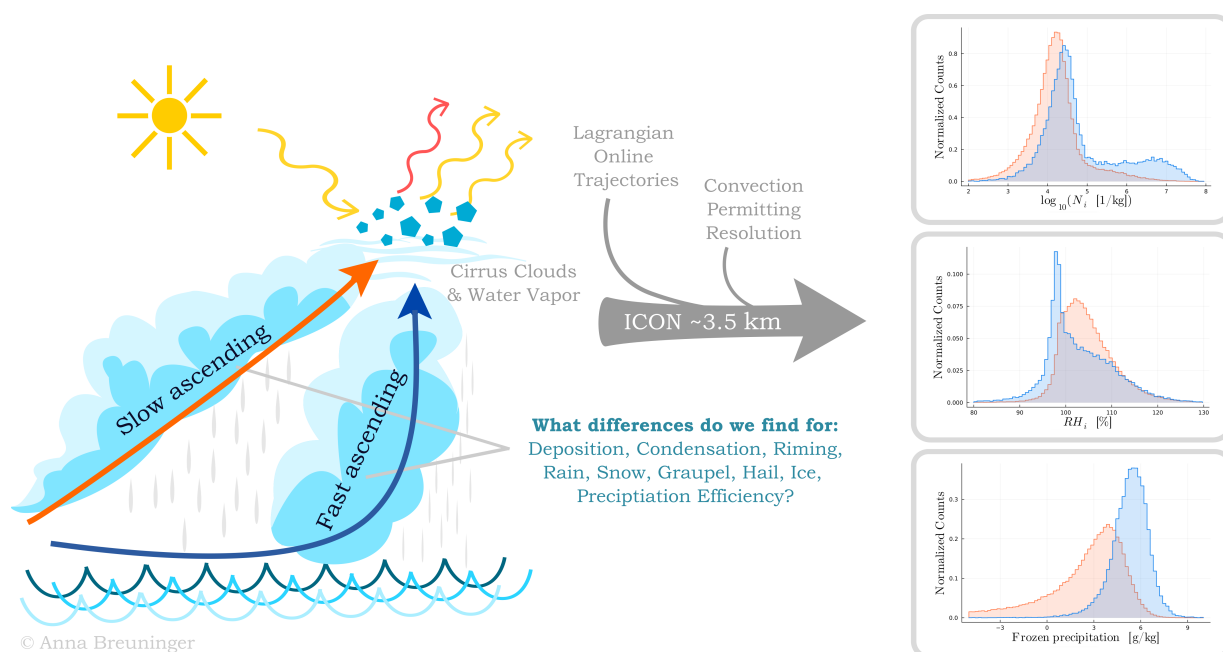
Changes to the sea surface temperature have almost no effect on the WCB outflow and do not change the pressure or temperature of glaciation. The ice content and relative humidity over ice in the WCB outflow are highly sensitive to the capacitance of ice and to processes influenced by INP concentrations. CCN concentrations control the number and size of hydrometeors during the WCB ascent but have little effect on the outflow. These results highlight important sensitivities in the model representation of WCBs and may inform the selection of parameters to perturb in future studies.

Chapter 3

The Role of Convection

Research Question I

Which microphysical processes control WCB moisture transport into the UTLS, and how do these processes differ between regions of convective and non-convective activity?



This Chapter has been published in *Atmospheric Chemistry & Physics*. For author contributions, see Section 2.5.

How to cite: Schwenk, C. and Miltenberger, A.: The role of ascent timescales for warm conveyor belt (WCB) moisture transport into the upper troposphere and lower stratosphere (UTLS), *Atmospheric Chemistry and Physics*, 24, 14 073–14 099, <https://doi.org/10.5194/acp-24-14073-2024>, 2024.c



The role of ascent timescales for warm conveyor belt (WCB) moisture transport into the upper troposphere and lower stratosphere (UTLS)

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Abstract. Warm conveyor belts (WCBs) are coherent ascending airstreams in extratropical cyclones. They are a major source of moisture for the extratropical upper troposphere and lower stratosphere (UTLS), where moisture acts as a potent greenhouse gas and WCB-associated cirrus clouds contribute to cloud radiative forcing. However, the processes controlling WCB moisture transport and cloud properties are poorly characterised. Furthermore, recent studies have revealed (embedded) convection as a ubiquitous feature of WCBs, highlighting the importance of understanding their updraught and microphysical structure. We present a Lagrangian investigation of WCB moisture transport for a case from the WISE (Wave-driven ISentropic Exchange) campaign based on a convection-permitting simulation. Lagrangian non-dimensional metrics of the moisture budget suggest that the ascent timescale (τ_{600}) strongly controls the end-of-ascent total moisture content, which is largest for slowly ascending trajectories ($\tau_{600} \geq 20$ h, 30 % of all WCB trajectories). This is due to relatively warm end-of-ascent temperatures and the strong temperature control on transported water vapour. Deviations from equilibrium water vapour condensate partitioning are largest for slow trajectories due to faster glaciation and lower ice crystal numbers. A local moisture transport minimum at intermediate τ_{600} results from a shift towards a riming-dominated precipitation formation pathway and decreasing outflow temperatures with decreasing τ_{600} . The fastest trajectories ($\tau_{600} \leq 5$ h, 5 % of all WCB trajectories) transport the largest condensate mass to the UTLS due to less efficient condensate loss and produce the longest-lived outflow cirrus clouds. Models that parameterise convection may under-represent these processes, potentially impacting weather forecasts and climate predictions.

1 Introduction

Warm conveyor belts (WCBs) are regions of large-scale, coherent, ascending airflow in the vicinity of extratropical cyclones (ETCs), which produce the elongated cloud band typically associated with ETCs. WCBs usually transport moist boundary layer air poleward and into the upper troposphere over a span of 2 d, during which a host of complex physical processes occur that make WCBs important across planetary scales for Earth's weather, climate and radiative budget (Madonna et al., 2014).

As a common weather phenomenon, WCBs occur frequently and are major drivers of mid-latitude weather, influ-

encing large-scale weather conditions for weeks after their dissipation (Joos et al., 2023; Madonna et al., 2014; Rodwell et al., 2018). The ascending air masses undergo diabatic heating, which can produce potential vorticity (PV) anomalies. These anomalies affect cyclone strength and lifetime (Binder et al., 2016; Rossa et al., 2000), modify the upper-level wave guide and jet stream (Pickl et al., 2023; Grams et al., 2011; Joos and Wernli, 2011; Wernli, 1997), and have been linked to atmospheric blocking (Pfahl et al., 2015; Wandel et al., 2024). Additionally, WCBs contribute to over 80 % of the total precipitation in Northern Hemisphere storm tracks (Pfahl et al., 2014; Eckhardt et al., 2004). WCBs are therefore crucial for predicting extreme weather events such as heat waves

and storms (Oertel et al., 2022; Flaounas et al., 2017), and the incorrect representation of WCBs has been identified as a key factor in amplifying forecast uncertainties (Pickl et al., 2023; Berman and Torn, 2019).

As a large-scale climatological phenomenon, WCBs significantly influence Earth's climate. When the warm, planetary boundary layer (PBL) air ascends into the upper troposphere and lower stratosphere (UTLS), a multitude of warm-phase, mixed-phase and cold-phase microphysical processes occur, producing various cloud types. Each of these cloud types has distinct microphysical properties and affects Earth's radiative budget through cloud radiative forcing (CRF) differently (Joos, 2019). Determining the net contribution to CRF is complex and has been investigated in several studies (Joos, 2019; Gehring et al., 2020; Stewart et al., 1998). The long-lived cirrus cloud shield that accompanies a WCB in the later stages of its life cycle is optically thin and has low cloud-top temperatures, resulting in positive CRF (Spichtinger et al., 2005; Binder et al., 2020; Joos, 2019). Conversely, the warm-phase, low-altitude clouds that form in the early stages of a WCB's life cycle at lower latitudes are optically thick and have high cloud-top temperatures, resulting in negative CRF (Joos, 2019). Mixed-phase clouds, with varying optical thickness and temperature, have an uncertain CRF sign. Therefore, the net CRF of WCBs and their impact on Earth's radiative budget are sensitive to the microphysical representation of clouds produced in WCBs (Joos, 2019).

In addition to CRF, WCBs influence Earth's radiative budget because they transport water vapour into the UTLS. This happens even though the moisture content of ascending air in a WCB is greatly reduced by the formation of precipitation (Madonna et al., 2014; Sprenger and Wernli, 2003). Water vapour is the most dominant greenhouse gas, and studies have shown that UTLS water vapour, in particular, is the most significant positive feedback factor in climate change (Li et al., 2024; Held and Soden, 2000). Even minor changes in UTLS specific humidity can have a substantial impact on the greenhouse effect (Wang et al., 2001; Hansen et al., 1984). WCBs have generally been associated with significant troposphere–stratosphere exchange (Gettelman et al., 2011), but so far, no studies have quantified the amount of vapour transported into the UTLS by WCBs. Only Zahn et al. (2014), who investigated the origins of measured UTLS water vapour using backwards trajectories, determined that WCBs are one of four dominant transport pathways for water into the UTLS. It also remains unclear how much moisture during the WCB ascent is converted into precipitation, which microphysical processes are responsible for this conversion and to what extent each process contributes.

Given the importance of WCBs, it is essential that weather and climate models accurately represent them. However, it remains unclear what uncertainties are introduced by the parameterisation of microphysical processes when modelling WCBs. These parameterisations are known to introduce significant uncertainties when modelling storms and convective

systems (Barthlott et al., 2022; Khain et al., 2015; Dearden et al., 2016; Hieronymus et al., 2022), and sensitivity experiments have shown that the choice of microphysical parameterisation scheme impacts the evolution of WCBs (Mazoyer et al., 2021, 2023). However, the uncertainties introduced by these parameterisations regarding WCB vapour transport into the UTLS and cloud formation, for example, have not been quantified. Consequently, the uncertainty of the CRF induced by WCBs remains largely unknown. Oertel et al. (2023) investigated which microphysical processes produce diabatic heating in a WCB and how these differ for quickly and slowly ascending WCB air parcels. However, it has not yet been studied which microphysical processes modify vapour and hydrometeor content in ascending WCB air, meaning that the uncertainties introduced for the transport of vapour are also unknown.

Another crucial aspect of WCBs that remains poorly understood is the role of deep and embedded convection for the transport of vapour and the production of clouds. Traditionally, WCBs are viewed as consisting of a coherent, slowly ascending and slantwise-ascending airstream. However, recent studies using observations and high-resolution simulations suggest that convection is a common phenomenon in WCBs (Binder et al., 2016; Oertel et al., 2019, 2020, 2021; Rasp et al., 2016). This has several important implications: (i) WCBs may produce extreme precipitation in regions of deep convection (Flaounas et al., 2017), (ii) deep convection could be a significant source of stratospheric ice clouds (Zou et al., 2021) and has been shown to transport more tracers into the upper atmosphere than large-scale advection (Purvis et al., 2003), and (iii) forecast errors tend to grow more rapidly in areas of convective activity (Rasp et al., 2016). Consequently, studies have examined the differences between convective and non-convective WCB air parcels. For example, Oertel et al. (2020) found that hydrometeor compositions are different for trajectories with differing ascent times, while Oertel et al. (2023) show that the microphysical processes contributing to adiabatic heating also differ across ascent timescales. These findings suggest that different microphysical processes govern the evolution of WCB air parcels during their ascent, depending on their convective characteristics. However, the extent to which convective regions within a WCB differ from areas of slantwise ascent with respect to water vapour and hydrometeor transport to the UTLS remains unclear.

Recent studies indicate that high-resolution simulations, which explicitly resolve convection, produce different outcomes compared to models that parameterise convection. Specifically, simulations suggest that Lagrangian WCB air parcels ascend more rapidly, reach higher altitudes and experience enhanced latent heating when resolution is increased (Choudhary and Voigt, 2022). The presence of rapid-ascent regions, as confirmed by observations (e.g. Blanchard et al., 2020), raises questions about how accurately commonly used coarse-resolution models with parameterised convection cap-

ture the effects of these convective elements regarding WCB dynamics, water transport and outflow cirrus properties. Potential biases in the simulated water vapour transport and cirrus properties have implications for projections of future UTLS moisture and thermodynamic structure. Consequently, to accurately investigate the role of convection in the transport of moisture by WCBs, it is essential to use high-resolution, convection-permitting simulations that can capture these processes in detail.

This paper aims to address these knowledge gaps by (i) investigating the microphysical processes that control WCB moisture transport into the UTLS and (ii) quantifying how these processes differ between regions of convective and non-convective activity. To this end, we conduct a case study of a North Atlantic WCB using a convection-permitting Icosahedral Nonhydrostatic (ICON) simulation (using the two-moment microphysics scheme from Seifert and Beheng, 2005) and investigate both Eulerian and Lagrangian output fields.

The paper is structured as follows. First, we describe our ICON model setup and the online trajectory module it utilises. Next, we present our novel Lagrangian methods for investigating microphysical processes during WCB ascent of air parcels. This section introduces and derives Lagrangian formulations for common diagnostics, such as the drying ratio and precipitation efficiency. We then present our case study and describe the WCB in question. In our results, we first examine ascent diagnostics to quantify the prevalence of convection in the WCB and then focus on moisture transport. We analyse several diagnostics (i) at the beginning of the ascent, (ii) at the end and in the hours following the ascent, and (iii) during the ascent to provide a comprehensive picture of the moisture transported by the WCB. Finally, we summarise our findings and conclude with the implications of this study for future research on WCBs.

2 Methods

In this chapter, we first describe the ICON model setup used for our simulations. We then outline our WCB trajectory selection algorithm and explain our method of normalising the ascent-time axis around the WCB ascent of a trajectory. Finally, we give a detailed overview of the variables used in our investigations. In particular, we derive Lagrangian formulations for the drying ratio, precipitation efficiency and condensation ratio, among others.

2.1 ICON model setup

Our case study considers a WCB that dominated weather over the northern Atlantic Ocean on 23 September 2017. To simulate the evolution of the atmosphere for this case study we used the Icosahedral Nonhydrostatic (ICON) modelling framework (version 2.6.2; Zängl et al., 2014). Our simulation was initialised with the operational ICON global analy-

sis at 00:00 UTC on 20 September 2017, a couple of hours before the cyclone formed. The simulation ran for 96 h until 00:00 UTC on 24 September 2017, at which time the WCB dissipated over northern Europe.

In addition to a computation domain that spans the entire globe, ICON can embed regions of higher spatial resolution (called “nests”) within the global grid. ICON includes an implementation of two-way nesting, meaning that simulations on lower-resolution domains couple to the next highest-resolution simulation by providing lateral boundary conditions and are in turn nudged towards the solutions of the higher resolution (Zängl et al., 2022). Our model ran with two nested domains. Computations on the global domain were conducted with a 120 s time step on a R03B07 grid (effective grid spacing of approximately 13 km) and on the nested domains with 60 and 30 s time steps on R03B08 (~ 6.5 km) and R03B09 (~ 3.3 km) grids, respectively. The nests were chosen such that they cover the main region of WCB ascent, which was visually determined before the model run using cloud cover and sea-level pressure data from ERA5. Figure 1 shows the nested setup with a snapshot of ERA5 reanalysis data (Hersbach et al., 2020) over the main WCB ascent region for 10:00 UTC on 23 September 2017. We can see that the main part of the WCB falls into the nested domains.

We chose the spatial resolutions of the highest-resolution nested domain such that it permits convection. On the global domain, convection is parameterised using the Tiedtke–Bechtold convection scheme (Tiedtke, 1989; Bechtold et al., 2008), whereas on the nested domains only shallow convection is parameterised. On all domains non-orographic gravity wave drag, sub-grid scale orographic drag (Lott and Miller, 1997; Orr et al., 2010) and turbulence are parameterised using the standard ICON schemes. Radiation is treated using the Rapid Radiative Transfer Model. Cloud microphysical processes, which are central to this paper, are calculated using the two-moment microphysics scheme by Seifert and Beheng (2005). This scheme represents hydrometeor mass mixing ratios q_x and the corresponding number concentrations n_x for six hydrometeor species: cloud droplets (c), rain drops (r), ice crystals (i), snow flakes (s), graupel grains (g) and hailstones (h). The model setup we have described here allows for a detailed representation of the region of interest, i.e. the WCB ascent and initial outflow, while offering a consistent representation of the larger-scale surroundings of the WCB.

Online trajectories

Lagrangian data can be computed online (during a model simulation, using wind fields computed at each time step) or offline (after a model simulation, using Eulerian output files). Online trajectories offer accurate solutions to the trajectory equation for high-resolution models and detailed perspective at high temporal resolution on e.g. microphysical-

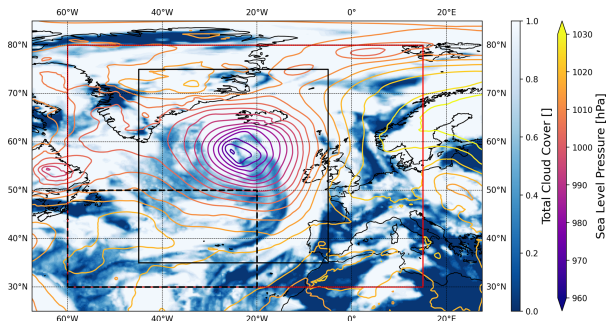


Figure 1. Snapshot of the atmosphere at 10:00 UTC on 23 September 2017 using ERA5 data. The nested domain boundaries are shown for the first nest (red) and the second nest (black). The starting area of the online trajectories is also outlined (dashed black line). Through the sea-level pressure contours and the total cloud cover one can clearly see the prominent WCB cloud band and low-pressure system.

process rates (Oertel et al., 2023; Miltenberger et al., 2020, 2016, 2013). We therefore used an online trajectory module implemented in ICON by Miltenberger et al. (2020) and extended by Oertel et al. (2023) to support domain nesting. We started online trajectories every 5 h during the 96 h simulation, and data were written to a file every 30 min. Starting positions were chosen at random from within the starting region (shown using the dashed black lines in Fig. 1) and from six vertical model levels (ranging between ca. 1000 and 800 hPa). We chose the starting region, which spans across the two nested domains, on the basis of prior offline trajectories that were calculated using ERA5 reanalysis data. This ensured that we obtained plenty of trajectories representing the WCB airstream.

2.2 Selection of warm conveyor belt trajectories

Our selection algorithm identified WCB trajectories conditional to them fulfilling two criteria. Firstly, a WCB trajectory must ascend at least 600 hPa in at most 48 h. This is a widely used criterion/limit imposed by previous studies (e.g. Madonna et al., 2014; Oertel et al., 2023; Rasp et al., 2016). Secondly, all WCB trajectories must be within two visually determined longitude–latitude regions at two certain times. These regions were determined on the basis of Eulerian cloud cover and sea-level pressure data from our simulation. This ensured that the trajectories were not part of a mesoscale convective system (MCS) unrelated to our WCB or any other cyclone.

2.3 Ascent timescales and characterisation of convective behaviour

In this section we introduce the metrics with which we determine the ascent timescale and convective behaviour of

WCB trajectories. The simplest and most widely used ascent timescale is the fastest time in which a trajectory ascends 600 hPa, called τ_{600} (Rasp et al., 2016; Oertel et al., 2023). We use this timescale to differentiate convective (small τ_{600}) from slantwise-ascending (large τ_{600}) WCB trajectories. We are however interested in the entire WCB ascent, not just the section where the ascent is fastest. We therefore also look at the time for which the ascent velocity remains above 8 hPa h^{-1} before and after the τ_{600} time segment and call this time τ_{WCB} . We use τ_{WCB} to define the period during which a trajectory is ascending as part of the WCB; before/after this time it is defined to be in the inflow/outflow.

To get a broader picture of the overall convective activity, we also consider additional variables. Similarly to τ_{600} , we also look at the 300 and 400 hPa ascent times (τ_{300} , τ_{400}) to quantify the convective behaviour for less deep ascent (i.e. embedded mid-level convection). We also look at the maximum 2 h pressure difference ($\max(\Delta p_{2\text{h}})$) and the minimum 10 h pressure difference ($\min(\Delta p_{10\text{h}})$) during the WCB ascent to assess the “regularity” with which trajectories rise. The last metric of course only makes sense for trajectories that take more than 10 h to ascend.

The metrics introduced here allow for a deeper insight into a trajectory’s ascent behaviour. Consider for instance the three most important metrics: τ_{600} , τ_{WCB} and $\max(\Delta p_{2\text{h}})$, visualised by a drawing in Fig. 2. In this example, a WCB trajectory is ascending from the boundary layer at around 950 to 350 hPa over a time of approximately 28 h. The ascent velocity remains above 8 hPa h^{-1} (τ_{WCB}) for 14 h and is coloured yellow. The minimum time taken for a 600 hPa ascent along the trajectory’s entire path (τ_{600}) is 8 h and is coloured red. The maximum pressure difference experienced by the trajectory in any 2 h time span ($\max(\Delta p_{2\text{h}})$) is 350 hPa and is coloured cyan. This trajectory would be described as ascending in a purely slantwise fashion if only τ_{WCB} is considered, but $\max(\Delta p_{2\text{h}})$ shows that there is a short period of convective ascent. It is therefore important to consider multiple metrics to correctly classify and understand the ascent behaviour of a trajectory.

2.4 Normalised ascent timescale

We want to be able to compare the physical processes of WCB trajectories with different ascent timescales at similar stages of their ascent. This becomes difficult when looking at data along a τ_{WCB} time axis, as fast trajectories will complete their ascent and reach the UTLS at a time when slow trajectories may still be in the lower troposphere. We therefore normalise the τ_{WCB} ascent time for all trajectories and introduce the normalised ascent time $\tilde{t}$. The time at which trajectories begin to ascend at a velocity of at least 8 hPa h^{-1} (t_0 , defined as the beginning of τ_{WCB} ; see Sect. 2.3) is set to 0 ($\tilde{t} = 0$), and the time at which they stop ascending at a velocity of at least 8 hPa h^{-1} ($t_{\text{end}} = t_0 + \tau_{\text{WCB}}$) is set to 1 ($\tilde{t} = 1$). Now all normalised ascent times are bound by 0 and 1, and

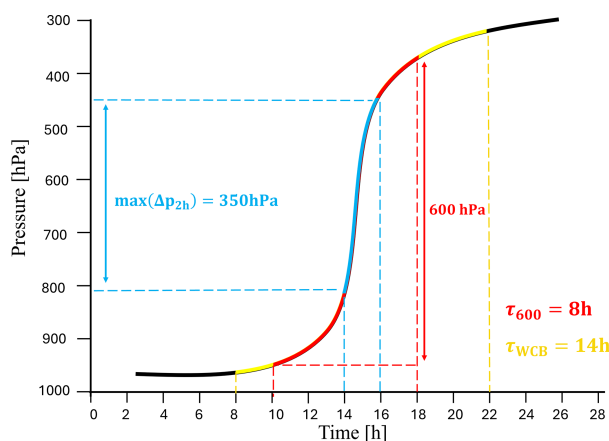


Figure 2. Visualisation of the three most important ascent metrics, τ_{600} (red), τ_{WCB} (yellow) and $\max(\Delta p_{2h})$ (cyan), which help to give a full picture of a trajectory's ascent behaviour.

any two trajectories at a given normalised time are likely to be at similar stages of their ascent. This approach allows for an investigation and comparison of physical processes along a coherent and ascent-oriented time axis.

2.5 Time-integrated rates

We consider time-integrated (as opposed to instantaneous) microphysical-process rates. This is done so that we can determine the accumulated effects of a microphysical process along a trajectory during its WCB ascent. ICON computes instantaneous microphysical-process rates $\psi_{\text{inst}}(t)$ with units of $[\psi_{\text{inst}}(t)] = \text{kg kg}^{-1} \text{s}^{-1}$ (see Supplement). We use these to define the time-integrated rates that are calculated by the online trajectory module using the spatially interpolated instantaneous rates at each model physics time step:

$$\psi(t) = \int_0^t \psi_{\text{inst}}(t') dt', [\psi] = \text{kg kg}^{-1}. \quad (1)$$

To focus on the WCB ascent only, i.e. ignoring anything that happened before, the time-integrated rates are set to 0 (in post-processing) at the start of the WCB ascent, i.e. $\psi(\tilde{t} = 0) = 0$.

2.6 Moisture budget

To understand the physical mechanisms that control the final moisture content in the outflow of a WCB, we will now define and derive variables that describe the specifics of removing/adding moisture from/to an air parcel in a Lagrangian framework.

2.6.1 Lagrangian drying ratio

In the most general sense, the efficiency with which moisture is removed from a parcel can be described by the drying ratio (DR) (e.g. Miltenberger, 2014). It quantifies how much moisture initially contained in the Lagrangian parcel is removed by the end of the ascent and is given by

$$\text{DR} = \frac{Q_{\text{tot}}(0) - Q_{\text{tot}}(1)}{Q_{\text{tot}}(0)} \quad (\text{note: } Q_{\text{tot}}(1) = Q_{\text{tot}}(0) \cdot (1 - \text{DR})). \quad (2)$$

Here $Q_{\text{tot}}(\tilde{t})$ is the total moisture content at the beginning ($\tilde{t} = 0$) and at the end ($\tilde{t} = 1$) of the WCB ascent, so the numerator is equal to the change in total moisture during the ascent. $Q_{\text{tot}}(\tilde{t})$ is the sum of the specific humidity (q_v) and all hydrometeor mass mixing ratios (q_c, q_r, q_i, q_s, q_g and q_h ; see Appendix B). This definition of DR is a Lagrangian one because we consider the differences in moisture along the trajectories.

DR is non-zero if moisture is removed from a trajectory by the end of the ascent ($\text{DR} = 1$ if all moisture is removed). The mechanisms for moisture removal in our simulation are (i) the turbulence parameterisation, (ii) the convection parameterisation, (iii) precipitation and (iv) numerical uncertainties/interpolation errors. We can therefore rewrite Eq. (2) as follows:

$$\text{DR} = \frac{Q_{\text{trc}}(1) + P(1)}{Q_{\text{tot}}(0)}. \quad (3)$$

In this equation, $Q_{\text{trc}}(1)$ is the sum of the moisture removed by the turbulence and convection parameterisations and the numerical uncertainties by the end of the ascent (for details see Appendix B). We refer to this as the moisture loss due to the “mixing” processes. $P(1)$ is the time-integrated net precipitation rate at the end of the ascent, which accounts for the moisture lost by the gravitational removal of hydrometeors from the parcel (see Appendix B). Note that this is not surface precipitation but the net flux of hydrometeors leaving the parcel.

We can conclude that the numerator terms in Eq. (3) include all the mechanisms in our simulation that can remove moisture from a trajectory. Their sum is therefore equal to the change in total moisture during the ascent and Eq. (3) is equal to Eq. (2).

2.6.2 Lagrangian microphysical and mixing drying ratio

More detailed insight into the processes that determine DR can be gained by splitting DR into terms that describe different moisture removal mechanisms. First we define the mixing drying ratio,

$$\text{DR}_{\text{mix}} = \frac{Q_{\text{trc}}(1)}{Q_{\text{tot}}(0)}, \quad (4)$$

which describes what fraction of initial moisture is lost due to mixing processes. All remaining moisture ($Q_{\text{tot}}(0) - Q_{\text{tr}}(1)$) either remains in the parcel or is removed by precipitation, which brings us to the definition of the microphysical drying ratio:

$$\text{DR}_{\text{mphys}} = \frac{P(1)}{Q_{\text{tot}}(0) - Q_{\text{tr}}(1)}. \quad (5)$$

DR_{mphys} is the fraction of this remaining moisture that is removed by the precipitation of hydrometeors. Its name comes from the fact that moisture loss due to precipitation depends on the microphysical processes that form and grow hydrometeors.

For completeness, note that the denominators of the two drying ratios we have defined are not the same. Our choice of denominators ensures that DR_{mphys} only describes moisture removal by precipitation *after* the removal by mixing processes is taken into account. DR_{mphys} is not 1 if all the moisture has been removed by precipitation alone but if all the moisture that has not been removed by turbulence/convection has been removed by precipitation. This choice reflects the focus of this paper on microphysical processes. Regardless of the choice of denominator, DR can be recovered from DR_{mix} and DR_{mphys} as follows:

$$\text{DR} = \text{DR}_{\text{mix}} + \text{DR}_{\text{mphys}} - \Theta, \quad \text{with:} \quad (6)$$

$$\Theta = \text{DR}_{\text{mix}} \cdot \text{DR}_{\text{mphys}}.$$

The interaction term Θ results from the fact that both moisture removal by mixing and precipitation occur simultaneously, i.e. any moisture removed by turbulence/convection is not available for precipitation. We now shift our focus to the processes that govern DR_{mphys} .

2.6.3 Lagrangian precipitation efficiency (PE) and condensation ratio (CR)

For water vapour to be removed from an air parcel through precipitation, it must first be converted into cloud condensate. This conversion can happen either by forming new hydrometeors (nucleation) or by the growth of existing hydrometeors through deposition and condensation. These hydrometeors may already be present in the parcel, or they may enter it through sedimentation or mixing via turbulent processes. This conversion can be quantified by the condensation ratio (CR) (e.g. Barstad et al., 2007; Miltenberger, 2014). After the conversion, condensate may leave the parcel as precipitation. This process is quantified by the precipitation efficiency (PE) (e.g. Miltenberger, 2014; Dacre et al., 2023).

For the Lagrangian CR we obtain the following expression:

$$\text{CR} := \frac{C_{\text{hy}}(1) + E_{\text{v}}(1)}{\text{VAP}(1)}. \quad (7)$$

A detailed definition of the above terms can be found in Appendix B4, but in short, $C_{\text{hy}}(\tilde{t})$ quantifies the conversion of

vapour into hydrometeors, $E_{\text{v}}(\tilde{t})$ quantifies the reverse (conversion of hydrometeors into vapour) and $\text{VAP}(\tilde{t})$ is equal to the initial vapour content minus the vapour lost to parameterisations other than cloud microphysics, all at the normalised ascent time $\tilde{t}$. Therefore, this equation states that CR is equal to the net hydrometeor growth by microphysical processes ($C_{\text{hy}}(1) + E_{\text{v}}(1)$) divided by the net initial water vapour content ($\text{VAP}(1)$). It thus gives the fraction of vapour that is initially present or brought in through turbulence, convection or numerical residuals, which is subsequently converted into hydrometeors. Using VAP , instead of $Q_{\text{tot}}(\tilde{t} = 0)$ as in the definition of DR, ensures that CR is bounded by 0 and 1.

For PE we obtain

$$\text{PE} := \frac{P(1)}{C_{\text{hy}}(1) + E_{\text{v}}(1) + \text{HYD}(1)}. \quad (8)$$

This formulation states that PE is equal to the total net precipitation out of the parcel ($P(1)$) divided by the sum of net hydrometeor growth and net initial hydrometeor content ($\text{HYD}(1)$, initial hydrometeor content minus the hydrometeors lost to parameterisations other than cloud microphysics by the end of the ascent; see Appendix B4). In other words, PE tells us how many hydrometeors that (i) were formed during the ascent, (ii) were already present at the beginning, or (iii) were carried in by the turbulence or convection scheme (or numerical residual) are precipitated out of the parcel by the end of the ascent. The incorporation of (iii) ensures that PE is bounded by 0 and 1.

Finally, we get DR_{mphys} from PE and CR as follows:

$$\text{DR}_{\text{mphys}} = \text{PE}(\text{CR} + \epsilon) \frac{\text{VAP}(1)}{\text{VAP}(1) + \text{HYD}(1)}, \quad (9)$$

using

$$\epsilon := \frac{\text{HYD}(1)}{\text{VAP}(1)}. \quad (10)$$

We point out that

$$\frac{\text{VAP}(1)}{\text{VAP}(1) + \text{HYD}(1)} = \frac{P(1)}{Q_{\text{tot}}(0) - Q_{\text{tr}}(1)} \quad (11)$$

is equal to 1 if all initial moisture is lost exclusively to precipitation. For small values of $\text{HYD}(1)$ the fraction on the right-hand side of Eq. (9) goes to 1 and ϵ goes to 0, meaning that we can approximate

$$\text{DR}_{\text{mphys}} \approx \text{PE} \cdot \text{CR} \quad \text{for } |\text{HYD}(1)| \ll 1. \quad (12)$$

This equation is typically used in studies focussing on a regional moisture budget analysis from an Eulerian perspective (e.g. Barstad et al., 2007; Miltenberger, 2014).

In this section we have presented all variables using the normalised ascent timescale introduced in Sect. 2.4. However, the definitions are also valid on the real time axis as long as the start and end times for the integrated rates are set

accordingly. It is also possible to quantify at DR, PE, etc. at arbitrary times during the ascent by replacing $\tilde{t} = 1$ with a normalised time $\tilde{t} \in (0, 1]$ (leaving the initial conditions unchanged). We will now discuss the physical meaning of PE and CR in DR_{mphys} .

2.6.4 Interpretation of PE and CR's physical meaning in DR_{mphys}

The physical meaning of PE and CR in DR_{mphys} can be easily interpreted if we assume that the net initial hydrometeor content $HYD(1)$ is negligible. In this case DR_{mphys} approximates to $PE \cdot CR$. CR determines how much moisture is converted into hydrometeors, and PE determines how many hydrometeors are removed. If CR is large and PE is small, then the air parcel is converting moisture efficiently but removing it inefficiently. Conversely, if CR is small and PE is large, then the parcel is efficient at hydrometeor removal but has low hydrometeor production. DR is less than 1 if some moisture remains, 1 if all moisture has been removed and 0 if no moisture has been removed. The following theoretical edge cases illustrate this:

1. $PE = 1$, $CR = 1$ and $DR = 1$.
There is no remaining moisture. All the vapour has been converted, and every hydrometeor has precipitated.
2. $PE = 1$, $CR < 1$ and $DR < 1$.
Any remaining moisture is vapour, and all hydrometeors have precipitated.
3. $PE < 1$, $CR = 1$ and $DR < 1$.
Hydrometeors make up all remaining moisture, and there is no vapour.
4. $PE = 0$, $CR > 0$ and $DR = 0$.
Water vapour is converted into hydrometeors, but they stay in the parcel. Q_{tot} is unchanged.
5. $PE > 0$ and $CR = 0$ but $DR > 0$ in the case that $HYD(1) > 0$ is not negligible.
This case only makes sense if the net initial hydrometeor content is not neglected, meaning $\epsilon > 0$. No moisture is converted, and only the initial hydrometeors precipitate. Q_{tot} decreases, but Q_v is unchanged.

As the last case illustrates, including $HYD(1)$ ensures that initial hydrometeors and additional hydrometeor removal processes (e.g. turbulence) are accounted for. Our interpretation of the role of PE and CR therefore holds as long as $HYD(1)$ is much smaller than $VAP(1)$. In Appendix B5 we show that our Lagrangian definitions for PE, CR and DR_{mphys} can heuristically be compared to the Eulerian definitions of the same variables from previous studies.

3 Case study

Here, we present a detailed description of the WCB that serves as the focus of our case study. A strong cyclone with a pronounced WCB characterised the weather over the northern Atlantic from the afternoon of 22 September 2017 to the end of 23 September 2017. Measurements in air masses related to this system were carried out as part of the *HALO-WISE* (*High Altitude and long Range Aircraft–Wave-driven ISentropic Exchange*) campaign (Haynes and Palm, 2023). Note that it is an example of an almost purely open-ocean WCB.

The conditions for cyclogenesis began to develop early on 20 September 2017. Strong southeasterly winds, resulting from the combination of the wind fields around a high-pressure system off the coast of Newfoundland and a low-pressure system off the southern tip of Greenland drove a large and cold air mass from the Greenland and the Labrador Sea across the northern Atlantic (Fig. A1a). On the western side of the high-pressure system (located at 45°W , 45°N) warm air was transported northwards, leading to the formation of a 1500 km long front at 52°N (Fig. A2a). The converging winds associated with the high- and low-pressure systems acted frontogenically and strengthened the temperature gradient. This induced a baroclinic zone with strong horizontal wind shear and began lifting air masses at the point of convergence. At this time (ca. 04:00 UTC on 21 September 2017), the low-pressure system started forming.

At the same time, an upper-level trough/positive potential vorticity (PV) anomaly approached from the northwest and propagated to the southeast (Fig. A2b). It induced cyclonic rotation in the atmospheric column, which, together with PV generated at lower levels from latent heating, resulted in the development of a PV tower. This induced strong cyclonic flow in the lower levels of the atmosphere (Fig. A3a). As a result, the cyclone underwent explosive cyclogenesis, with central surface pressure dropping from 1002 hPa at 04:00 UTC on 21 September to 977 hPa at 04:00 UTC the following day. In the cloud field a comma-shaped cloud band typical of WCBs formed on 22 September 2017 and propagated northeast until its northern tip reached Iceland early on 23 September (Fig. 1). At this point the cold front (along which convective clusters were found) stretched 4000 km across the Atlantic and reached south of 40°N . On 24 September 2017 the cyclone dissipated over the Norwegian Sea. The upper-level trough propagated further downstream (Fig. A5b). On the surface the cyclone brought relatively mild air (ca. 15°C) all the way to Iceland and Svalbard and therefore had a large impact on central and northern European weather.

From our simulation of the case (see Sect. 2.2 for a description of the setup) we identified 393 070 WCB trajectories associated with the cyclone described above. The spatial and pressure evolution for every 10th trajectory throughout the entire simulation is shown in Fig. 3. The colouring shows that most trajectories experience a large change in pressure

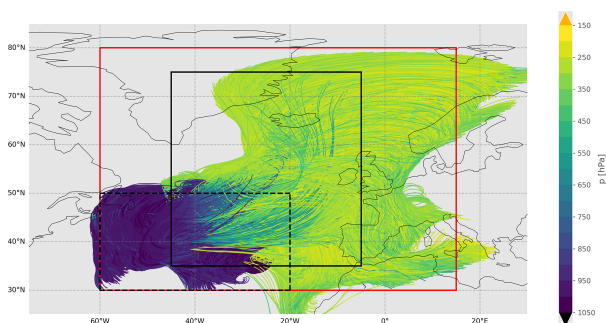


Figure 3. The geographical and pressure course of some selected WCB trajectories for the entire simulation time in relation to the first nest (red), second nest (black) and starting area (dashed black line).

(and hence their WCB ascent) within the highest-resolution domain. The average horizontal distance covered during the ascent is 2400 km. A few trajectories rise very early and remain in the southern region of the WCB. A majority of trajectories move along northwards during their ascent, in line with the comma-shaped cloud structure, and move eastwards after reaching the UTLS. Some then turn clockwise and move to the southeast, reaching as far south as Morocco, while others, completing their ascent further north, move across central or northern Europe. A large proportion end up as far north as Greenland.

4 Ascent statistics and convective behaviour

We characterise the ascent of WCB trajectories on the basis of their τ_{600} (Sect. 2.3). Figure 4a shows a histogram of τ_{600} and τ_{WCB} ascent times for all selected WCB trajectories. We see a unimodal distribution with a mean, median and mode for τ_{600} of 16.9, 14.5 and 11 h, respectively. There is large variability in ascent timescales with a steep decline for timescales shorter than 10 h and a tail towards longer ascent times. We define three ascent “regimes”: (i) “convective” trajectories with $\tau_{600} < 5$ h, representing 6 % of all WCB trajectories; (ii) “normal” trajectories with $18 \text{ h} \geq \tau_{600} \geq 6 \text{ h}$, representing 55 %; and (iii) “slow” trajectories with $\tau_{600} > 20 \text{ h}$, representing 30 % of all cases. However, τ_{600} ascent times do not form a complete picture of ascent behaviour, as they quantify only deep ascent. The 300 and 400 hPa ascent times (τ_{300} , τ_{400} , Fig. 4b) focus on smaller-scale embedded convective activity. Their distribution is not unimodal but instead bimodal with peaks at about 5 h (3 h) and 1.5 h (1 h) for τ_{400} (τ_{300}). Many trajectories therefore ascend rapidly for a large part of the total 600 hPa ascent but fail to maintain a high rate of ascent throughout the troposphere (an example for this is shown in Fig. C1b).

The widespread occurrence of rapid-ascent segments across the τ_{600} spectrum is confirmed by the distribution of

maximum pressure drops over 2 h (Fig. C1a): 38 % of slow trajectories experience periods where they rise about 200 hPa in 2 h, and 78 % of all trajectories have a 0.5 h period in which the average ascent velocity is larger than 120 hPa h^{-1} (gives 600 hPa in 5 h). Many trajectories also rise in “steps”, with two or more sections of rapid ascent followed by extended periods of almost no ascent (not shown); 30 % of slow trajectories experience a 10 h period in which they ascend less than 10 hPa ($\min(\Delta p_{10\text{h}}) < 10 \text{ hPa}$), and 10 % even experience a segment within τ_{WCB} in which they *descend* more than 40 hPa (not shown). Only 9 % of slow trajectories have $\max(\Delta p_{2\text{h}}) < 100 \text{ hPa}$, meaning that truly slantwise-ascending trajectories are in the minority and one cannot assume that a large τ_{600} value means that the entire ascent is slow and monotonous.

To ensure that the problematic cases (where τ_{600} is large but the trajectory is rising convectively for most of the ascent) are neglected, we remove slow trajectories that rise quickly for a large part of the ascent, i.e. $\max(\Delta p_{2\text{h}}) > 350 \text{ hPa}$, from the dataset for further analysis (removes about 5 % of slow trajectories). A similar condition has been imposed by Rasp et al. (2016). This additional condition is why the orange colour in Fig. 4a does not fill the histogram all the way to the top.

In summary, the WCB trajectories show a wide range of ascent timescales and almost all of them, even the slowly ascending ones, experience bursts of convective activity (Fig. C1). This shows that even for an open-ocean WCB, convection is a ubiquitous feature and that truly slantwise-ascending trajectories are a minority. We will now analyse the transport of moisture by the WCB.

5 Moisture transport analysis

WCB ascent connects the planetary boundary layer with the upper troposphere and thereby constitutes a source of moisture for the UTLS. The actual amount of water reaching the UTLS depends on the initial water content at the start of the ascent and the moisture loss processes during the ascent. For the selected case study, we present a detailed analysis of this moisture budget in the following.

5.1 Trajectory characteristics at the start of the WCB ascent

Convective trajectories have initial temperatures and pressures of approximately 20°C and 975 hPa, respectively, whereas slow trajectories have about 18°C and 965 hPa (Fig. 5a; recall that $\tilde{t} = 0$ is the start of the τ_{WCB} ascent and $\tilde{t} = 1$ is the end). Hence, on average faster-ascending trajectories begin their ascent at higher temperatures and pressures and have a higher initial total moisture content than slower-ascending trajectories. However, the spread of conditions for the start of slow trajectories is much larger than for the fast trajectories, as can be seen by the difference in

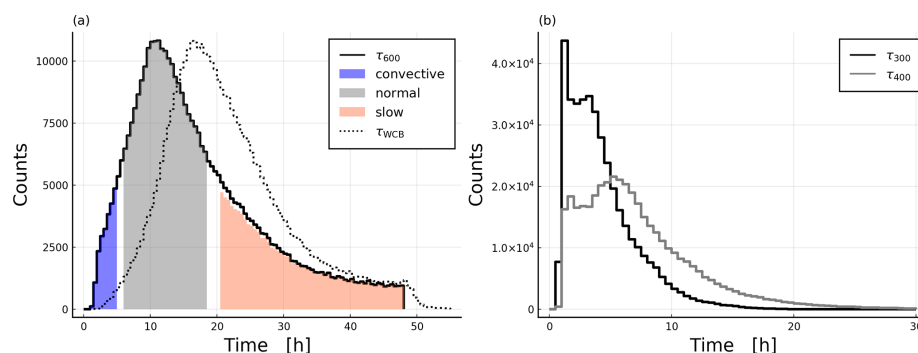


Figure 4. Ascent statistics that help us understand the convective behaviour of WCB trajectories. In (a) histograms for τ_{600} (solid line) and τ_{WCB} (dashed line) ascent times are shown, as well as the three categories into which they are grouped. In (b) the τ_{300} (black) and τ_{400} (grey) ascent times are shown. The distributions are different because many trajectories ascend a large portion convectively but not the full 600 hPa.

the mean and median values. This is consistent with the finding by Oertel et al. (2023) that convective trajectories originate in a more southerly part further away from the cyclone centre compared to slow trajectories. The initial total water content behaves similar to the temperature; i.e. it decreases with increasing ascent time (Fig. 6a). Consequently, the average $Q_{\text{tot}}(0)$ is 12.9 g kg^{-1} for convective trajectories and 11.3 g kg^{-1} for slow trajectories. The spread increases with ascent time. This means that we find a strong correlation for short ascent times with high vapour content and high temperatures. We suggest that this is because higher temperatures and vapour content mean that more buoyancy can be generated by latent heating, leading to faster ascent times (this is in line with findings from Schäfler and Harnisch, 2014). The initial hydrometeor content varies but is generally small ($< 0.4 \text{ g kg}^{-1}$, average of 0.08 g kg^{-1} for all trajectories, Fig. 6a) with a weak dependence on τ_{600} .

5.2 Trajectory characteristics at the end of the WCB ascent

5.2.1 Temperature and pressure

The pressure and temperature at the end of the ascent is lower for trajectories with smaller τ_{600} values compared to more slowly rising trajectories (Fig. 5b). This difference persists even many hours after completion of the ascent, with the minimum pressure and temperature that convective trajectories reach averaging 236 hPa and -56°C and 314 hPa throughout the entire simulation and -44°C for slow trajectories (not shown). This is partially due to the fact that faster trajectories ascend earlier and therefore further south (not shown), where the atmosphere is deeper. However, even when trajectories are rising at the same latitude, those with lower τ_{600} values will rise on average 90 hPa higher (not shown). The differences seen for final temperature and pressure are therefore largely due to differences in initial humid-

ity and temperature, where larger values favour latent heat release and stronger cross-isentropic transport.

5.2.2 Hydrometeor content

The total hydrometeor content at the end of the ascent ($Q_{\text{hy}}(1)$) is strongly influenced by ascent time and increases for a shorter τ_{600} (the maximum value of 0.17 g kg^{-1} is reached for $\tau_{600} = 1.5 \text{ h}$, Fig. 6b). Therefore, on average, convective trajectories transport far more hydrometeors (mass and number) into the UTLS than slow trajectories. This is made evident by the fact that even though convective trajectories make up only 5.8 % of all trajectories, they account for 14.3 % of $Q_{\text{hy}}(1)$. The mean $Q_{\text{hy}}(1)$ remains below 0.04 g kg^{-1} for $\tau_{600} > 10 \text{ h}$, where the spread is low. Ice is the dominant hydrometeor species at the end of the ascent for all trajectories, constituting 99 % of $Q_{\text{hy}}(1)$ (the remaining 1 % is predominantly snow). No liquid hydrometeors are found, although $T(1)$ is not below 235 K (the homogeneous freezing temperature) for all trajectories.

5.2.3 Vapour content and relative humidity over ice

Aside from condensate, WCB trajectories also transport water vapour into the UTLS. Specific humidity is strongly correlated with the saturation specific humidity over ice, which depends on temperature and pressure only (Fig. 7b), indicating that transport of specific humidity has strong thermodynamic constraints. This is also evident from the final vapour content $Q_v(1)$ (Fig. 6) showing the same dependence on τ_{600} as the final temperature (Fig. 5b). As a result, the majority of trajectories (65 %) have a relative humidity with respect to ice ($\text{RH}_i(1)$) of $(100 \pm 6) \%$ (Fig. 7a). However, while the thermodynamic conditions at the end of the ascent are the dominant factors in determining vapour content, a significant minority of trajectories deviate from 100 % saturation: 3.5 % of trajectories have $\text{RH}_i(1) < 94 \%$, and 31 % have $\text{RH}_i(1) > 106 \%$. Notably, more trajectories are super-

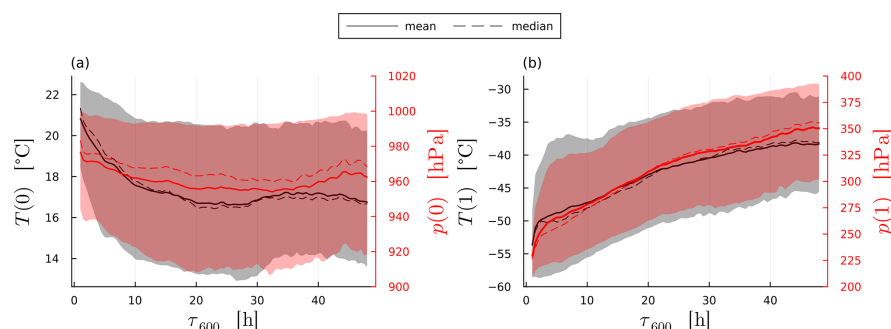


Figure 5. Mean (solid line) and median (dashed line) of the temperatures (black) and pressures (red) at the (a) beginning and (b) end of the WCB ascent. The 10th and 90th percentiles are shaded. Faster trajectories start their ascent at higher temperatures and pressures and conversely end their ascent at significantly lower temperatures and pressures. Note: the pressure axis is *not* flipped to make the correlation with temperature more obvious.

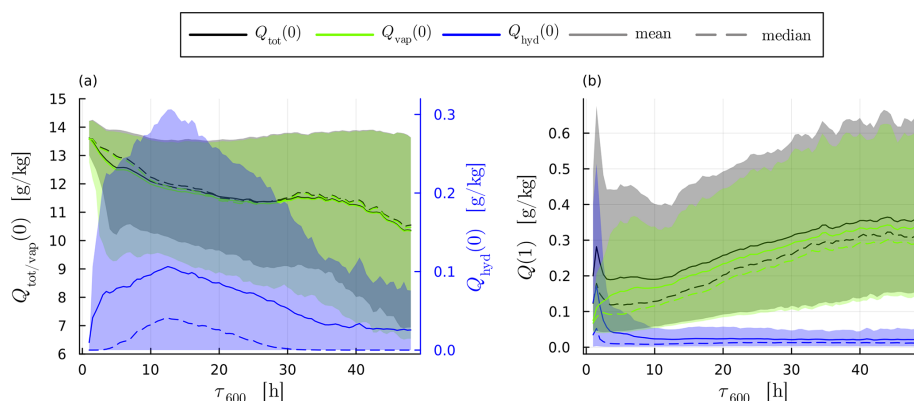


Figure 6. Mean (solid line) and median (dashed line) total moisture content (Q_{tot} , black) as well as the vapour (Q_{vap} , green) and hydrometeor (Q_{hyd} , blue) content at the (a) beginning and (b) end of the WCB ascent. The 10th and 90th percentiles are shaded. Note: the different y axis for $Q_{\text{hyd}}(0)$ in (a) is due to the difference in magnitude.

saturated than sub-saturated, with 70 % of all trajectories being supersaturated with respect to ice upon completing their ascent and entering the UTLS. This indicates that the WCB transports substantial amounts of ice-supersaturated air into the UTLS. The variability in $\text{RH}_i(1)$ likely stems from the different microphysical properties of the trajectories which control the relaxation of supersaturation to saturation values.

There are notable differences in $\text{RH}_i(1)$ between convective and slow trajectories. On average, convective trajectories have a slightly lower $\text{RH}_i(1)$ of 102.6 % compared to 104.7 % for slow trajectories. Although these averages are similar, the $\text{RH}_i(1)$ distributions differ significantly between the two groups. Convective trajectories exhibit a sharp peak at 97 % with a tail extending to higher values, while slow trajectories show a more bell-shaped distribution, peaking at 108 % (Fig. 7a). Consequently, only 56.5 % of convective trajectories are supersaturated, compared to 80 % of slow trajectories. This finding is particularly interesting given that convective trajectories contain much higher hydrometeor con-

centrations than slow trajectories at the end of the ascent. The lower ice number concentration in slow trajectories (Fig. 11) likely increases the supersaturation relaxation timescale and allows for the persistence of higher relative humidity.

We can summarise Sect. 5.2 by noting that convective trajectories reach higher into the atmosphere at the end of the ascent and have lower specific humidity than slow trajectories and higher hydrometeor contents than slow trajectories. The specific humidity at the end of the ascent is predominantly constrained by the outflow temperature. Substantial spread in the specific humidity at a given temperature is introduced by deviations away from saturation, which are more abundant in slow compared to convective trajectories. At the end of the ascent, convective trajectories contain, on average, 10 times more ice crystals (or 2 times more for the median) and roughly twice the ice mass compared to slow trajectories. Consequently, while both convective and slow trajectories transport a similar total water mass to the UTLS – with a minimum observed for intermediate ascent timescales – the

partitioning between water vapour and condensate, as well as the resulting ice cloud properties, depends significantly on the ascent timescale.

5.3 Moisture loss pathways

In this section we will compare the behaviour of moisture loss between convective and slow-ascending trajectories using the variables introduced in Sect. 2.6.

5.3.1 Total fractional moisture loss (DR)

DR indicates the percentage of moisture lost by the end of the ascent, and on average WCB trajectories lose more than 95 % of their moisture (Fig. 8a). DR decreases with increasing ascent time, meaning that faster trajectories lose a larger fraction of their initial moisture than slower trajectories. As we will discuss in the next paragraphs, this is mainly due to the lower temperature attained by fast-ascending trajectories, which results in low-saturation vapour pressure (see discussion on CR). The slight decrease in DR for the fastest-ascending trajectories is due to their elevated hydrometeor content (Fig. 6a; see discussion on PE). This shows that DR is predominantly controlled by thermodynamic conditions but can be modulated by the transport of hydrometeors.

5.3.2 Fractional moisture loss due to mixing processes (DR_{mix})

The fraction of the initial humidity that is removed by turbulent mixing during the ascent is quantified by DR_{mix} (Eq. 4). Overall, DR_{mix} increases with increasing ascent time (Fig. 8c), meaning that slower-ascending trajectories experience more fractional moisture loss due to processes other than precipitation than fast-ascending trajectories. The individual contributions to DR_{mix} (turbulence, convection parameterisation, numerical residuals) as functions of τ_{600} are shown in Fig. C4. The largest contribution to DR_{mix} comes from the turbulence parameterisation; the convection parameterisation and numerical residuals play a secondary role. This makes sense because a longer ascent time means more time spent in the planetary boundary layer and a longer time for surrounding air to be mixed into the air parcel through turbulence. Indeed, for all WCB trajectories the moisture loss due to mixing occurs in the first part of the ascent at pressures larger than about 700 hPa (Fig. C5a, b). A surprising trend is that for the longest ascent times DR_{mix} decreases slightly. We presume that this is because the slowest trajectories move along regions that experience smaller horizontal and vertical wind shear (which drive turbulence). Circumstantial evidence for this assumption is provided by temporally averaged wind shear amplitudes along the ascent (Fig. C2), which indicates that slower-ascending trajectories are more part of the large-scale coherent flow of the WCB than fast-ascending trajectories.

5.3.3 Fractional moisture loss due to precipitation ($P(1)/Q_{\text{tot}}(0)$)

The net precipitation $P(1)$ decreases with increasing ascent time (Fig. C3a), correlating with less available moisture in the slowly ascending trajectories (Fig. 6a). This suggests (see discussion on PE) that precipitation formation is very efficient and precipitation amounts are strongly controlled by thermodynamic constraints on condensate formation. Fast-ascending trajectories experience greater fractional moisture loss due to net precipitation ($P(1)/Q_{\text{tot}}(0)$) than slow-ascending trajectories (Fig. 8f). $P(1)/Q_{\text{tot}}(0)$ is also larger than DR_{mix} for all trajectories, showing that precipitation is the dominant mechanism for moisture loss of WCB trajectories, regardless of their ascent time. Note that $P(1)/Q_{\text{tot}}(0)$ is essentially the mirror image of DR_{mix} as both terms contain all possible moisture loss processes.

The dominant form of net precipitation for almost all trajectories is the warm phase (rain) (Fig. C3a). Only the fastest-ascending trajectories have equal amounts of frozen precipitation (ice, snow, graupel, hail) and rain. The net flux of frozen hydrometeors is negative for many slow-ascending trajectories and is largely a mirror image of the net rain flux. This suggests that slow trajectories convert frozen hydrometeors entering them from above (mainly graupel and snow, Fig. C3b) into rain by melting. Slow trajectories spend a large fraction of their ascent just below the melting layer, which allows for a substantial influx of frozen hydrometeors in melting conditions (see Sect. 5.4). For all trajectories with $\tau_{600} < 20$ h, graupel precipitation is the dominant precipitating hydrometeor and makes up 33 % of precipitation for convective trajectories (Fig. C3b). So for all trajectories but especially for fast-ascending trajectories, frozen hydrometeors play a larger role in removing moisture.

5.3.4 Efficiency of moisture loss through precipitation (DR_{mphys} , PE, CR)

Finally, we measure how condensate formation and the removal of condensate through sedimentation each contribute to the variability in net precipitation and outflow moisture content for different τ_{600} values. The fraction of inflow moisture that is not removed by turbulence and is converted to condensate is quantified by the Lagrangian condensation ratio CR. CR is on average 0.98 and decreases from about 0.99 for convective to about 0.97 for slowly rising trajectories (Fig. 8d). The large CR reflects the large vertical displacement of all trajectories and supports the strong cross-isentropic ascent of the WCB. Water vapour content at the end of the ascent is predominantly thermodynamically constrained (Sect. 5.2), and it is therefore not surprising that the dependency of CR on τ_{600} reflects the outflow temperature of the WCB trajectories.

The fraction of the condensate that is removed by gravitational settling can be quantified by the Lagrangian precip-

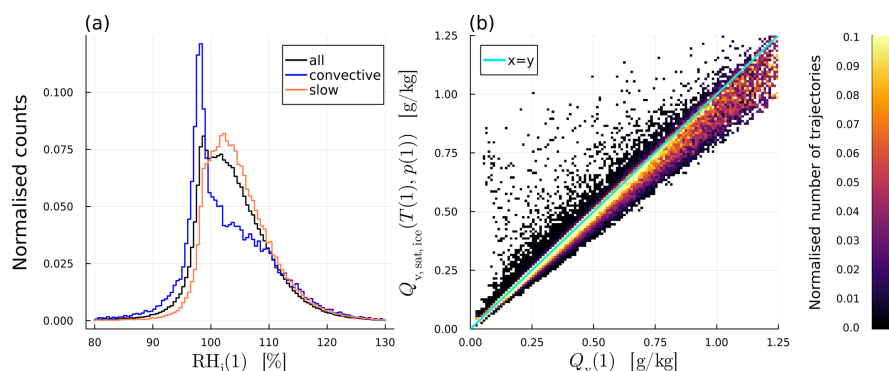


Figure 7. (a) Relative humidity over ice for all trajectories as well as the convective and slow trajectories, shown as a normalised histogram at the end of the ascent for all (black), convective (blue) and slow (orange) trajectories. (b) Two-dimensional histogram of calculated saturation specific humidity (over ice) given temperature and pressure at the end of the ascent (y axis), over the actual specific humidity (x axis).

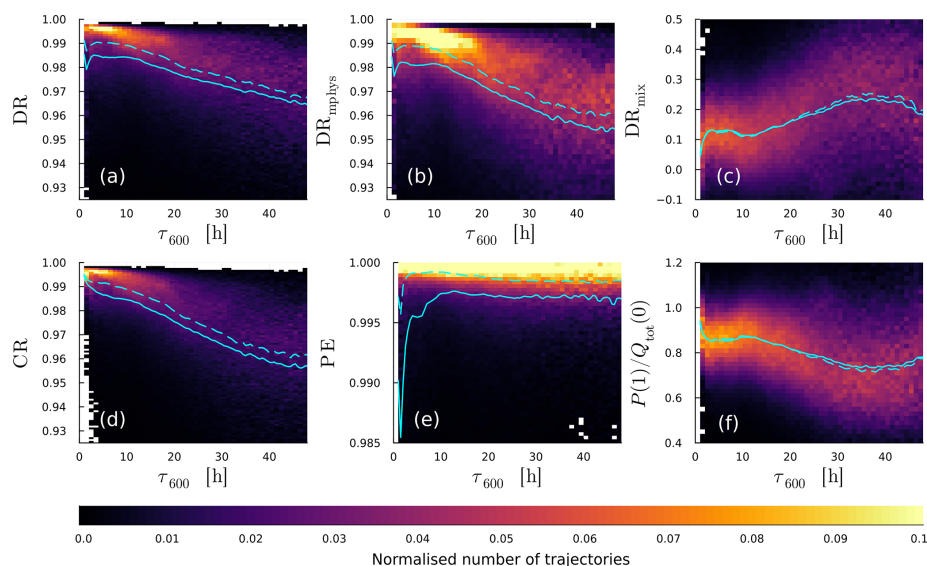


Figure 8. Two-dimensional normalised histograms showing (a) DR, (b) DR_{mphys} , (c) DR_{mix} , (d) CR, (e) PE and (f) $P(1)$ over τ_{600} . Mean (solid) and median (dashed) are plotted in cyan.

itation efficiency PE. PE is very close to 1 for all trajectories (Fig. 8e), indicating that WCB air parcels are extremely efficient at removing the hydrometeors they form/grow during the ascent. For the fastest ascents, PE decreases sharply, which is reflected in the large final hydrometeor content for convective trajectories. Because of the longevity of ice particles in convective trajectories, PE also remains larger for faster ascent times in the hours after the ascent (not shown).

Finally, PE and CR can be combined to a microphysical drying ratio DR_{mphys} (Eq. 9) analogous to DR_{mix} and the fractional moisture loss discussed earlier. In contrast to the fractional moisture loss, DR_{mphys} quantifies the loss of moisture due to cloud microphysical processes in the absence of turbulent mixing. DR_{mphys} is approximately equal to $PE \cdot CR$ as $HYD(1)$ and ϵ are much smaller than 1 (not

shown). Furthermore, since PE is essentially 1 for most trajectories, DR_{mphys} is almost equal to CR. Only for the fastest trajectories do we see a dip in DR_{mphys} that reflects the sharp decrease in PE.

In summary, the main pathway for moisture removal for all trajectories is precipitation with mixing processes playing a secondary role. Fractional moisture loss is $> 95\%$ for all trajectories and is mainly controlled by the vertical displacement that WCB parcels experience, i.e. temperature at the end of the ascent. This is reflected in CR, which behaves like final pressure and temperature, and PE, which is close to 1. The exception for PE is convective trajectories, where PE decreases sharply with decreasing ascent time. This shows that convective activity has a large impact on the efficiency with which moisture is removed by the end of the WCB ascent.

Note that the variability in PE, CR and DR_{mpyhs} with τ_{600} may appear small because all parcels are rising very strongly, but this variability gives rise to substantial variability in the moisture content in the UTLS (about a factor of 2 in $Q_{\text{tot}}(1)$, Sect. 5.2).

5.4 Trajectory characteristics during the ascent

Here, we present a detailed analysis of the moisture and thermodynamic evolution of rising WCB air parcels to explain the τ_{600} dependence of moisture budget terms. Specifically, we compare the evolutions of slow ascending and convective ascending parcels along the normalised ascent-time axis introduced in Sect. 2.4. Our focus is on the microphysical removal of moisture and key moisture budget terms, namely CR and PE.

5.4.1 Processes contributing to condensation efficiency (CR)

The condensation ratio at the end of the ascent CR(1) is higher for fast-ascending trajectories than slow-ascending trajectories (Sect. 5.3) and is largely controlled by the temperature. The evolution of CR (Fig. 9a) and temperature (Fig. 9b) during the ascent reveal a similar picture, with CR increasing much faster for convective trajectories than for slow ones, mimicking the faster decrease in temperature for fast trajectories.

In Sect. 5.3 we also found distinct differences in the relative humidity distribution for slantwise-ascending and convective ascending parcels. To clarify the physical processes driving these differences, we present the time evolution of relative humidity over water RH_w and ice RH_i , the liquid fraction (i.e. the ratio of liquid condensate mass to total condensate mass), and the ice crystal number concentration (Fig. 9c–f; note that the x axis is now temperature, which allows us to focus on deviations away from thermodynamic equilibrium). During the ascent but at temperatures above 0°C , the median RH_w increases as WCB parcels approach the cloud base and reaches close to 100 % at about 12°C for both sets of trajectories. It remains close to 100 % for a substantial part of the ascent, reflecting the control of the saturation adjustment scheme on the specific humidity in the model. Note that small deviations away from 100 % are likely due to uncertainties in the interpolation of specific humidity, temperature and pressure to the trajectory position in the on-line trajectory module.

After reaching temperatures below -6°C , the median RH_w decreases rapidly for slow trajectories, while for convective trajectories, conditions close to water saturation persist until temperatures of about -17°C . This difference is due to the more rapid glaciation of slowly rising trajectories compared to convective ones, as is also evident from the differences in the median liquid fraction below -10°C (Fig. 9e). Smaller vertical velocities for slow-ascending tra-

jectories (Fig. C5c) mean both that the Wegener–Bergeron–Findeisen process is more efficient and that there is more time for depositional growth (Fig. C6a–c).

A total of 90 % of both sets of trajectories are fully glaciated before reaching the homogeneous freezing temperature for cloud droplets (-38°C), which is also reflected in the fact that RH_i values close to saturation are being reached already at temperatures of about -30°C . In the fully glaciated part of the ascent (temperatures between -30 and -40°C), the median RH_i is smaller for convective than for slow trajectories despite larger vertical velocities. This is likely because the much higher ice crystal number concentrations of convective trajectories in this temperature range allow for a more efficient conversion of vapour into ice (Fig. 9f). The difference in ice number concentrations can be attributed to two factors: (i) within this temperature range, the Hande et al. (2015) parameterisation has a strong temperature dependence of ice-nucleating particles (INPs) for immersion freezing. Since slow-ascending trajectories glaciate at warmer temperatures, immersion freezing produces fewer ice crystals when compared to convective trajectories. (ii) Convective trajectories have less time available for the sedimentation of ice crystals and the aggregation to snow.

In summary, the dependence of CR and the $RH_i(\bar{\tau} = 1)$ distribution on the ascent timescale τ_{600} can be explained by the stronger vertical displacement and less rapid glaciation in convective trajectories. The use of the saturation adjustment scheme, along with the specifics of the ice nucleation parameterisation, may affect the quantitative results. We are currently conducting a follow-up study to investigate the impact of these factors.

5.4.2 Processes contributing to precipitation efficiency (PE)

As discussed in Sect. 5.3, the precipitation efficiency PE also increases with decreasing τ_{600} , with a weak indication of reduced PE for the fastest ascents ($\tau_{600} \leq 2$ h). The temporal evolution of PE (Fig. 10a) and of relevant microphysical properties (Fig. 10b–f) during the ascent reveal the underlying physical processes that lead to this difference. The evolution of PE is much more similar between slow and convective trajectories than the evolution of CR and is not directly tied to the evolution of temperature and pressure. This further validates the hypothesis that CR and PE are useful metrics for understanding how the (thermo)dynamic and microphysical conditions control precipitation formation (see also the discussion in e.g. Miltenberger et al., 2016; Miltenberger, 2014).

In the first quarter of the ascent, corresponding to pressures roughly below 800 hPa (Fig. C5b), PE increases very quickly to 0.9. At this time warm-rain processes and accretion dominate the microphysics (Fig. 9b), while parcels are located below the main WCB cloud, where they are increasingly af-

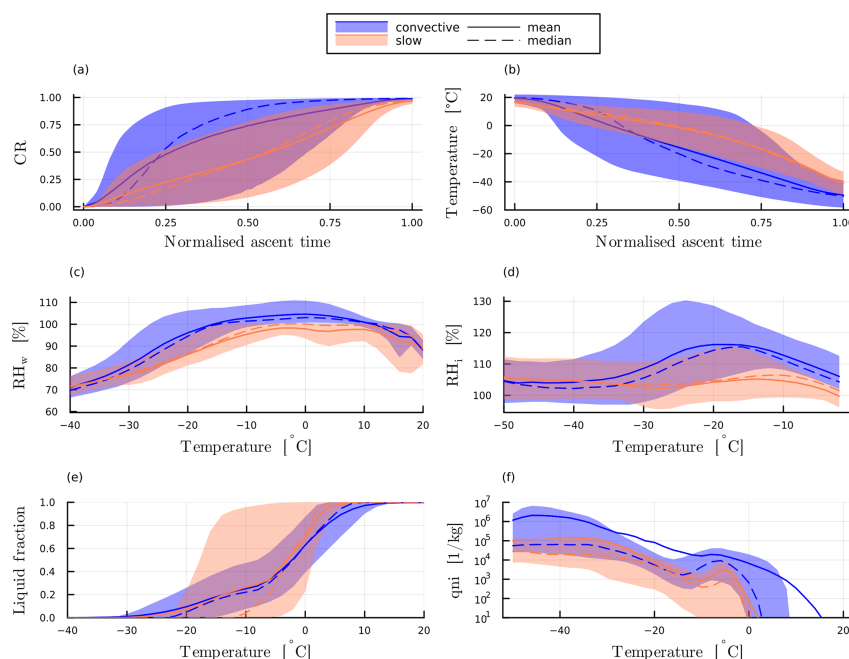


Figure 9. (a) Condensation ratio CR and (b) temperature over normalised ascent time. (c) Relative humidity over water RH_w , (d) relative humidity over ice RH_i , (e) liquid fraction and (f) ice crystal number concentration over parcel temperature during the WCB ascent. In all panels the mean (solid) and median (dashed) and the 10th and 90th percentiles (shaded) are plotted in blue for convective and orange for slow trajectories.

ected by hydrometeor influx from above (not shown). Consistently, the hydrometeor mass mixing ratio strongly increases in the first quarter of the ascent for both slow and convective trajectories (Fig. 10c), suggesting that condensate loss is less efficient than condensate production. At the end of this first phase of the ascent both sets of trajectories have a similar PE.

After $\tilde{t} = 0.25$, PE slowly increases for both sets of trajectories, but by $\tilde{t} = 1$ it is slightly larger for convective trajectories. This shows that for PE the microphysical evolution at later stages of the ascent is decisive. During this period, the hydrometeor mass mixing ratio steadily decreases for convective trajectories, whereas it remains nearly constant or even slightly increases for slow trajectories (Fig. 10b). This suggests that with a similar condensate formation rate, precipitation efficiency rises more rapidly in convective trajectories than in slow ones. Notably, the median frozen precipitation shows no net loss of condensate mass for slow trajectories until $\tilde{t} = 0.75$, while it rapidly increases beyond $\tilde{t} = 0.4$ for convective trajectories (Fig. 10c). This indicates that in slow trajectories, falling frozen precipitation does not collect much additional hydrometeors while passing through the considered parcels. Conversely, frozen precipitation in convective trajectories gathers significant additional condensate mass, amplifying sedimentation flux across the air parcel. Partitioning of frozen condensate reveals that convective trajectories contain a much larger proportion of grau-

pel, while snow formation dominates in slow trajectories (Fig. 10e and f). Larger graupel mass mixing ratios are consistent with larger vertical velocities (Fig. C6d), a larger liquid fraction (Fig. 9e) and a more vertical than slantwise ascent (not shown) of convective trajectories. They also suggest efficient riming growth (Fig. C6c), which explains positive net frozen precipitation. Towards the end of the ascent and after full glaciation, i.e. $\tilde{t} \geq 0.75$, slow trajectories also have a positive net frozen precipitation, which is predominantly snow (not shown).

In summary, the dependence of PE on τ_{600} is due to different mixed-phase precipitation formation pathways controlled by vertical velocities and the availability of supercooled liquid. Riming growth dominates precipitation formation in convective trajectories, while aggregation and deposition are more important for hydrometeor growth in slow trajectories. This is consistent with findings in earlier studies on precipitation formation (Oertel et al., 2023). These distinct pathways, along with variations in mass growth rates and particle fall velocities, result in a lower PE for slow trajectories despite the longer time available for precipitation formation and sedimentation. Although PE appears reduced for very short τ_{600} ascent times, we refrain from further analysis due to the small number of trajectories in this category and the coarse time resolution of our trajectory (output) data.

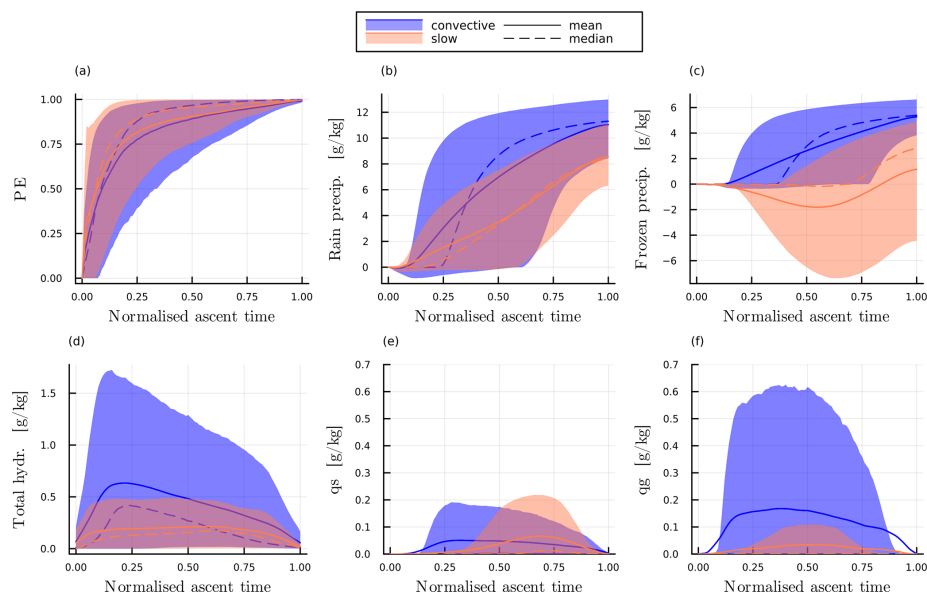


Figure 10. (a) Precipitation efficiency PE; (b) net rain precipitation; (c) net frozen precipitation; and the (d) total hydrometeor, (e) snow and (f) graupel mass mixing ratio over the normalised ascent time. In all panels the mean (solid) and median (dashed) and the 10th and 90th percentiles (shaded) are plotted in blue for convective and orange for slow trajectories. Note: positive net precipitation signifies a loss of condensate from the trajectories from gravitational settling.

5.5 Trajectory characteristics after the ascent

Since virtually all hydrometeors at the end of the ascent are ice, the differences in hydrometeor contents in the hours after the ascent between convective and slow trajectories are also seen when looking only at ice particle mass and number concentrations (Fig. 11). The ice number concentration shows a particularly large difference, with convective trajectories containing up to 15 times more ice particles than slow trajectories directly at the end of the ascent. This difference remains large up to 24 h after the ascent, after which convective trajectories contain (on average) roughly 10^4 kg^{-1} ice particles and slow trajectories contain roughly $5 \times 10^3 \text{ kg}^{-1}$ ice particles (Fig. 11a). Ice mass mixing ratios for both groups are similar roughly 5 h after the end of the ascent (Fig. 11b). The initial strong decrease in number concentration and particle mass is due to sedimentation and aggregation of (large) ice crystals. This is reflected in the geometric radius of ice particles (Fig. 11c), which decreases with time. There is a small increase/stagnation of ice particle radius after 5 h, which is due to the production of new ice particles through homogeneous and heterogeneous freezing (not shown). Overall, the geometric radius of ice particles is smaller for convective trajectories than for slow trajectories. Convective trajectories therefore appear to produce clouds with longer lifetimes and greater optical thickness (more and smaller ice particles) than slow trajectories. These clouds also have lower temperatures, since convective trajectories ascend to lower pressures and temperatures (Fig. 5b). A key

takeaway from this finding is that the radiative properties of clouds formed by air parcels in the WCB outflow depend on their ascent pathways. The significance of these differing radiative properties is further influenced by the extent to which WCB outflow clouds are overlaid by in situ cirrus clouds and the optical thickness of these cirrus layers.

In the hours after the ascent, RH_i decreases faster for slow trajectories than for convective trajectories (Fig. 11d). For both groups the 90th percentile remains above 100 % even after 25 h, but the mean and median for convective trajectories are much larger. This is in part because in the hours after the ascent, convective trajectories keep ascending, whereas slow trajectories descend approximately 20 hPa after 20 h (not shown). Convective trajectories however also produce more ice particles that remain in the air parcel for longer, and they counteract the formation of subsaturation by sublimating in case the relative humidity goes below 100 %.

After the ascent, we can summarise the behaviour of trajectories by noting that convective trajectories remain supersaturated for longer periods (up to more than a day) and retain a higher ice number and mass concentrations much longer than slow trajectories.

6 Conclusions and discussion

In this paper, we investigate moisture transport into the UTLS by a WCB that occurred on 23 September 2017. Our analysis is based on numerical simulations using the ICON model in a double-nested, convective-permitting setup

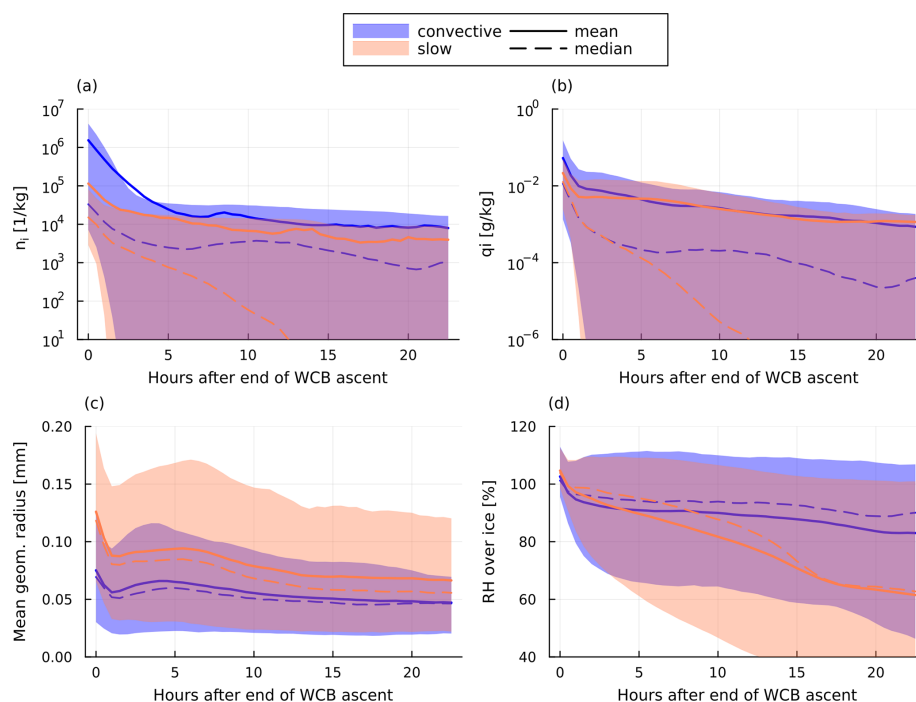


Figure 11. The (a) ice number concentration, (b) ice mass concentration, (c) mean geometric radius of ice particles and (d) relative humidity over ice in the hours after the end of the WCB ascent. The mean (solid) and median (dashed) and the 10th and 90th percentiles (shaded) are plotted in blue for convective and orange for slow trajectories.

and utilises high-resolution air mass trajectory data from the ICON online trajectory module.

After selecting trajectories that ascend at least 600 hPa within 48 h (our definition of a WCB trajectory), we first investigated their ascent behaviour. Various ascent-time metrics suggest an abundance of fast, convective ascent in the WCB: 78 % of all trajectories experience a 0.5 h period in which the average ascent velocity is larger than 120 hPa h^{-1} . The fast ascent for most trajectories does not extend over the full 600 hPa ascent, consistent with recent observations of embedded convection in the WCB (Blanchard et al., 2020). A small fraction of WCB trajectories ($\sim 5\%$) completes the entire WCB ascent in few hours. Hence, (embedded) convection is confirmed as an ubiquitous feature also in the WCB investigated here. The distribution of 600 hPa (τ_{600}) is similar in structure to earlier investigations of Rasp et al. (2016) and Oertel et al. (2021, 2023), although unsurprisingly case-to-case variations in the fraction of very fast-ascending trajectories emerge.

Next, we analysed how moisture content and thermodynamic conditions change at the beginning, during and at the end of the WCB ascent and how these changes vary across different ascent timescales. To highlight the impact of different ascent timescales we contrast properties from trajectories with ascent timescales of less than 5 h (“convective” trajectories) and those with ascent timescales more than 20 h

(“slow” trajectories). The convective trajectories begin their ascent at higher temperatures and pressures than slow trajectories and therefore have larger specific humidity values at the start of the ascent. Only small variations in the initial hydrometeor content are found. By the end of the ascent, all WCB trajectories have lost more than 95 % of their initial moisture. However, there is substantial variability in the outflow moisture content: total moisture varies by roughly a factor of 10 across all trajectories (~ 0.05 to $\sim 0.6 \text{ g kg}^{-1}$) and about a factor of 3 if the median values per ascent timescale are considered (~ 0.1 to $\sim 0.3 \text{ g kg}^{-1}$). Our analysis of moisture loss pathways suggest that 80 % to 90 % of moisture is lost by precipitation formation (predominantly rain). Turbulent mixing is more prominent for slow trajectories but still only contributes up to 20 % of moisture loss and mainly occurs at pressures larger than 700 hPa.

Variability in end-of-ascent moisture content primarily reflects differences in specific humidity, with slow trajectories showing higher values than convective ones, opposite to their initial conditions. This is largely due to colder outflow temperatures for convective trajectories and the strong correlation between specific humidity and temperature. While deviations from the saturation specific humidity are of secondary importance, they may still influence the evolution of outflow cirrus clouds. The fraction of supersaturated trajectories is larger for slow (80 %) than for convective trajectories

(56.5 %). It is shown that the final partitioning between water vapour and condensate is strongly influenced by the temperature at which air parcels fully glaciate, as well as the number of ice crystals present at temperatures below approximately -20°C . Slow trajectories glaciate earlier than fast trajectories and contain fewer ice crystals.

Hydrometeor content at end of the ascent is generally small. The median hydrometeor content slightly decreases with decreasing ascent timescales but strongly increases for trajectories with ascent timescales less than about 3 h (the mean increases for ascent timescales less than about 10 h). The evolution of microphysical properties and precipitation loss during the WCB ascent suggest that the increasing efficiency of condensate removal towards shorter timescales is driven by a transition of the precipitation formation pathway from deposition–aggregation-dominated to riming-dominated. The strong decrease in removal efficiency for the shortest ascent timescales may be driven by a decreasing ratio of the ascent timescale to the precipitation formation timescale. However, the output frequency of trajectory data is too small to allow for an in-depth analysis of the microphysical evolution of these trajectories.

Finally, we investigate the evolution of WCB outflow after the end of the ascent. While hydrometeors contribute little to the total moisture transport to the UTLS (except for an ascent timescale less than 10 h), they contribute to the WCB-associated cirrus clouds and therefore are important for cloud radiative heating. At the end of the ascent, convective trajectories have roughly twice the ice crystal number concentration of slow trajectories, although both have similar ice crystal mass. This difference in ice crystal concentration affects the mean geometric radius of the crystals, which in turn influences the cloud's lifetime, along with variations in the outflow position and the associated vertical motion. Specifically, the median ice mass mixing ratio in convective outflow cirrus clouds remains above 0.1 mg kg^{-1} for about 15 h, whereas in cirrus clouds formed from slow trajectories, this median value drops below 0.1 mg kg^{-1} after about 5 h.

The physical processes governing the moisture transport into the UTLS are captured by non-dimensional metrics of the Lagrangian moisture budget and provide a concise overview thereof:

- DR_{mix} generally increases with τ_{600} highlighting stronger moisture loss through turbulent processes.
- CR increases with decreasing τ_{600} reflecting the larger fraction of initial humidity converted to condensate for fast-ascending trajectories. This tendency is mainly due to larger vertical displacement and colder outflow temperatures with a secondary contribution of a smaller fraction of supersaturated trajectories for fast-ascending trajectories.
- PE slightly increases with decreasing τ_{600} due to a shift in the dominant precipitation formation pathway

towards riming-dominated. For very short τ_{600} ascent times, PE decreases substantially, resulting in a large hydrometeor mass mixing ratio in the WCB outflow.

The findings presented in this paper are broadly consistent with the limited number of earlier studies on the cloud and updraught structure of WCB clouds (Rasp et al., 2016; Oertel et al., 2019, 2021, 2023; Binder et al., 2020; Blanchard et al., 2020, 2021). Most of these studies focus on the ascent characteristics and mid-level latent heating distribution. Both modelling and observational case studies suggest significant variability in the vertical velocities with ubiquitous rapid-ascent segments. The modelling studies of Oertel et al. (2019, 2021, 2023) suggest different latent heating structures and microphysical pathways for precipitation formation for different local vertical velocity. On a climatological scale, the analysis of CloudSat and CALIPSO (Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations) data by Binder et al. (2020) supports the ubiquitous presence of rapid-ascent segments with different microphysical characteristics, i.e. reflectivity structure. However, we are not aware of an earlier study quantifying the impact of the mesoscale variability in ascent characteristics on the moisture import into the UTLS and WCB outflow cirrus clouds. While the consistency in mid-level cloud characteristics with earlier studies suggests model results presented here are physically plausible, the ICON model and in particular the representation of microphysical processes therein have some important limitations: first, the ICON model uses a saturation adjustment scheme, which enforces water-saturated conditions in mixed- and warm-phase cloud regions. Secondly, the parameterisation of ice formation (primary and secondary) still involves large uncertainties, which may influence the glaciation of WCB clouds. Therefore, the verification of our key results with observational data is strongly warranted and is planned for a follow-up study. However, an obstacle to this is the sparsity of high-quality (in-cloud) humidity data in the altitude range between 300 and 100 hPa.

Despite the pending verification with observational data, we think our findings are important, as they suggest that lower-resolution models with parameterised convection likely do not adequately capture the mesoscale variability in moisture transport and cirrus properties. Specifically, they may underestimate the intensity of convection in WCBs and therefore incorrectly model the transport of hydrometeors and moisture to the UTLS. This could have implications for studies that aim to determine the role of WCBs in Earth's climate. For instance, Joos (2019) conducted a WCB climatology using the ERA-Interim dataset (which parameterises convection) and found that WCBs usually have a cooling effect when located further south in the early stages of their development and a heating effect further north later on. The balance between the cooling and heating effect over the cyclone life cycle may be affected by the preferential occurrence of convective trajectories in the southern part of the WCB out-

flow and the associated long-lived cirrus clouds with comparably small effective radii. Additionally, the possible mis- or under-representation of microphysical processes during the WCB ascent could impact studies that investigate the role of WCBs in the formation of precipitation extremes (Catto et al., 2015) and forecast inaccuracies (Grams et al., 2011; Grams and Archambault, 2016; Martínez-Alvarado et al., 2015). This study demonstrates that the mesoscale updraught structure directly influences the moisture distribution within the WCB outflow in the UTLS. This suggests a pathway for error growth and climatological impacts of mesoscale structures beyond their effects on potential vorticity. Accurately capturing mesoscale variability is therefore essential for quantifying the impact of WCBs on extratropical weather and climate.

The results presented in this paper are strictly only valid for the analysed WCB case, which we have selected to represent a typical, non-extreme event of an open-ocean WCB. While we believe our findings provide valuable insights into WCB moisture transport and may be applicable to similar WCB cases, the extent to which they apply to the larger WCB population remains to be addressed in future studies. Furthermore, some of the findings summarised above may depend on the particular parameterisation setup used in the analysed simulation, such as the specific microphysics scheme and the absence of a deep-convection parameterisation. The influence of these model-specific factors on our key findings is currently being assessed through model sensitivity experiments and comparisons with observational data.

Appendix A: Case study plots

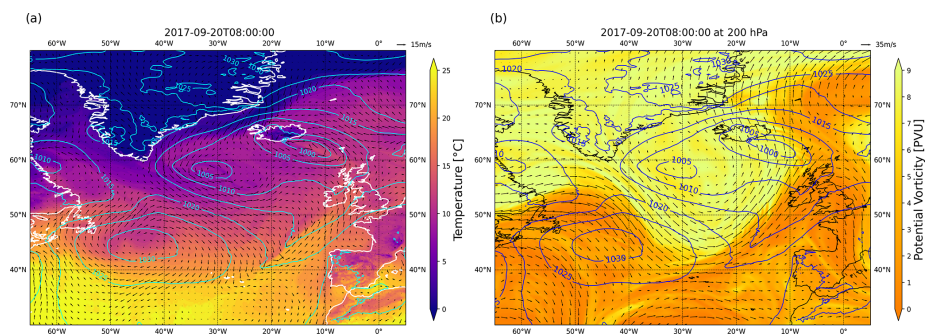


Figure A1. Snapshot of the northern Atlantic at 08:00 UTC on 20 September 2017. **(a)** The 2 m temperature (heatmap), sea-level pressure (contours) and 10 m wind field (arrows). **(b)** The upper-level PV (heatmap) and wind field (arrows) at 200 hPa and sea-level pressure (contours). The combining winds from the high-pressure system (-45° W, 45° N) and the low-pressure system (-30° W, 58° N) blow a cold air mass over the northern Atlantic.

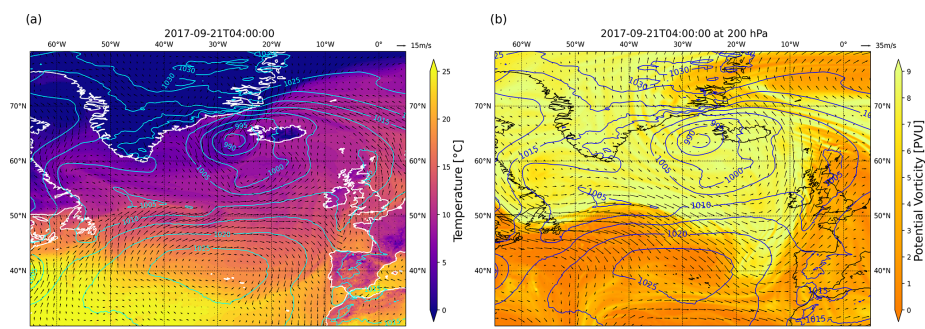


Figure A2. Snapshot of the northern Atlantic at 04:00 UTC on 21 September 2017. **(a)** The 2 m temperature (heatmap), sea-level pressure (contours) and 10 m wind field (arrows). **(b)** The upper-level PV (heatmap) and wind field (arrows) at 200 hPa and sea-level pressure (contours). The low-pressure system starts forming at -45° W, 53° N, where the winds converge. At the same time, an upper-level PV ridge (-55° W, 55° N) approaches from the west.

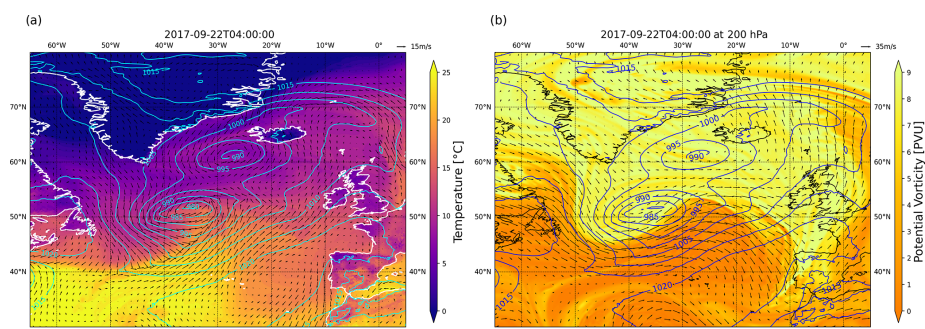


Figure A3. Snapshot of the northern Atlantic at 04:00 UTC on 22 September 2017. **(a)** The 2 m temperature (heatmap), sea-level pressure (contours) and 10 m wind field (arrows). **(b)** The upper-level PV (heatmap) and wind field (arrows) at 200 hPa and sea-level pressure (contours). The cyclone has undergone explosive cyclogenesis due to the upper-level PV anomaly inducing cyclonic rotation in the atmospheric column.

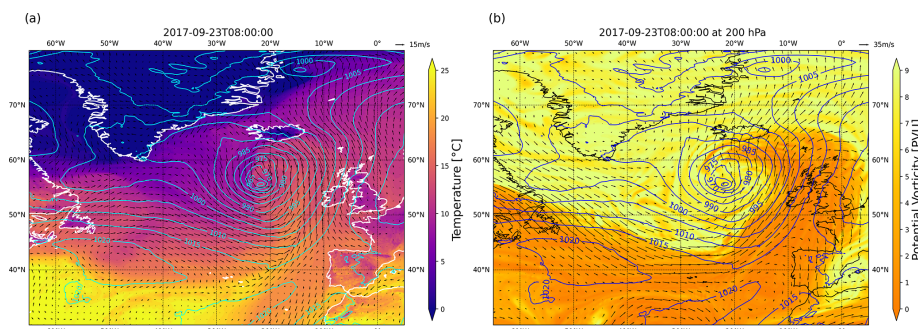


Figure A4. Snapshot of the northern Atlantic at 08:00 UTC on 23 September 2017. **(a)** The 2 m temperature (heatmap), sea-level pressure (contours) and 10 m wind field (arrows). **(b)** The upper-level PV (heatmap) and wind field (arrows) at 200 hPa and sea-level pressure (contours). The cyclone has assumed the typical WCB comma shape and is transporting low PV into the upper atmosphere. The northern tip of the WCB reaches Iceland.

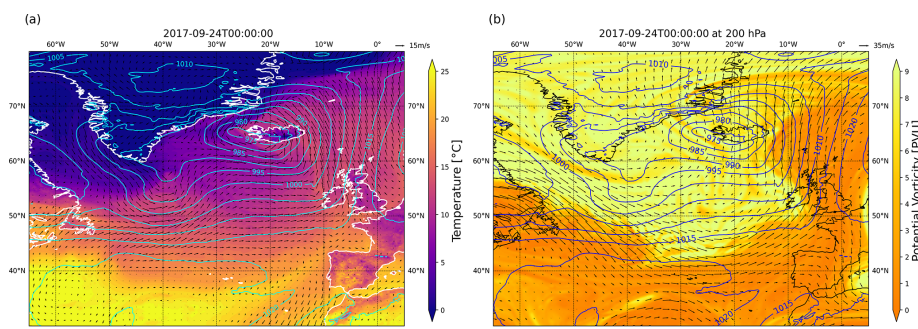


Figure A5. Snapshot of the northern Atlantic at 00:00 UTC on 24 September 2017. **(a)** The 2 m temperature (heatmap), sea-level pressure (contours) and 10 m wind field (arrows). **(b)** The upper-level PV (heatmap) and wind field (arrows) at 200 hPa and sea-level pressure (contours). The cyclone begins to dissipate over northern Europe and has created an upper-level trough.

Appendix B: Moisture budget terms: definitions and derivations

In this section, we define, describe and/or derive variables that we use in this paper to formulate Lagrangian versions of DR, PE and CR (Sect. 2.6.2, 2.6.3). Finally, we compare our Lagrangian versions with Eulerian versions from previous studies and show how they can be heuristically related.

B1 Total moisture content

The total moisture content is defined as follows

$$Q_{\text{tot}}(t) := \sum_x q_x(t), x \in [v, c, r, i, s, g, h]. \quad (\text{B1})$$

B2 Moisture removal by mixing processes

The total moisture removed by mixing processes is defined as a function of normalised ascent time $\tilde{t}$ as follows:

$$Q_{\text{tr}}(\tilde{t}) := Q_{\text{turb}}(\tilde{t}) + Q_{\text{conv}}(\tilde{t}) + \mathfrak{R}(\tilde{t}). \quad (\text{B2})$$

$Q_{\text{turb}}(\tilde{t})$ is the accumulated moisture removed from the parcel by the turbulent diffusion scheme and is defined as

$$Q_{\text{turb}}(\tilde{t}) := \sum_x q_{\text{turb}}(\tilde{t}), x \in [v, c, i]. \quad (\text{B3})$$

$Q_{\text{conv}}(\tilde{t})$ is the accumulated moisture removed from the parcel by the convection scheme, which is only of consequence when a trajectory is outside either of the nested domains. It is defined as

$$Q_{\text{conv}}(\tilde{t}) := \sum_x q_{\text{conv}}(\tilde{t}), x \in [v, c, r, i, s]. \quad (\text{B4})$$

For detailed explanation of q_{turb} and q_{conv} , see Supplement. Finally, $\mathfrak{R}(\tilde{t})$ is the residual term which accounts for

the accumulated moisture lost due to numerical and interpolation errors. It is defined as

$$\mathfrak{R}(\tilde{t}) := Q_{\text{tot}}(0) - Q_{\text{tot}}(\tilde{t}) + Q_{\text{tc}}(\tilde{t}) - P(\tilde{t}). \quad (\text{B5})$$

This term is necessary because although the online trajectory scheme provides highly accurate data computed at each time step using the Eulerian wind fields from ICON, numerical errors cannot be completely avoided. In each time step, the trajectory equation ($\frac{D\mathbf{x}}{Dt} = \mathbf{u}(\mathbf{x}, t)$) is solved to calculate the new trajectory position (Miltenberger et al., 2013). The wind field implied by this solution may differ slightly from that given by ICON, leading to the first possible source of uncertainty. Variables from the ICON Eulerian grid are then linearly interpolated to the new trajectory position, providing the second source of uncertainty. One of the consequences of this is that the moisture budget, calculated by summing the total water content and all possible moisture removal/addition mechanisms, is not closed:

$$\begin{aligned} Q_{\text{budget}}(\tilde{t}) := & \underbrace{\sum_{x=v,c,r,i,s,g,h} qx(\tilde{t})}_{=: Q_{\text{tot}}(\tilde{t})} \\ & - \underbrace{\sum_{x=v,c,i} qxturb(\tilde{t}) - \sum_{x=v,c,r,i,s} qxconv(\tilde{t})}_{=: Q_{\text{tc}}(\tilde{t})} \\ & + P(\tilde{t}) \neq \text{const}. \end{aligned} \quad (\text{B6})$$

The numerical residual term $\mathfrak{R}(\tilde{t})$ is defined such that it closes the moisture budget. We can split up the residual into the water vapour and the hydrometeor residual ($\mathfrak{R}_v(\tilde{t})$ and $\mathfrak{R}_{\text{hy}}(\tilde{t})$, respectively):

$$\mathfrak{R}(\tilde{t}) = \mathfrak{R}_v(\tilde{t}) + \mathfrak{R}_{\text{hy}}(\tilde{t}). \quad (\text{B7})$$

These terms are defined as

$$\mathfrak{R}_v(\tilde{t}) := Q_v(0) - Q_v(\tilde{t}) - Q_v(\tilde{t}) - \Delta\mathcal{H}(\tilde{t}) + Q_{v,\text{tc}}(\tilde{t}) \quad (\text{B8})$$

and

$$\mathfrak{R}_{\text{hy}}(\tilde{t}) := Q_{\text{hy}}(0) - Q_{\text{hy}}(\tilde{t}) + \Delta\mathcal{H}(\tilde{t}) + Q_{\text{hy},\text{tc}}(\tilde{t}) - P(\tilde{t}), \quad (\text{B9})$$

where we use $Q_{v,\text{tc}}(\tilde{t})$, $Q_{\text{hy},\text{tc}}(\tilde{t})$ and $P(\tilde{t})$ from Sect. 2.6 and define $\Delta\mathcal{H}(\tilde{t})$ as the sum over all microphysical processes that change the vapour/hydrometeor content (written now without time dependence):

$$\begin{aligned} \Delta\mathcal{H} := & qihh + qcnuc + \text{cond} + \text{evap} + qxdep \\ & + \text{satad_II} + \text{revap} + \text{fevap}. \end{aligned} \quad (\text{B10})$$

These processes are described in the Supplement.

B3 Lagrangian net precipitation rate

The Lagrangian net precipitation rate is defined as follows:

$$P(\tilde{t}) := \sum_x [qx_{\text{in}}(\tilde{t}) - qx_{\text{out}}(\tilde{t})], \quad x \in [r, i, s, g, h]. \quad (\text{B11})$$

The variables $qx_{\text{in}}(\tilde{t})$ and $qx_{\text{out}}(\tilde{t})$, in the Lagrangian framework of online trajectories, are the time-integrated rates of hydrometeors entering the parcel from above (in) and leaving it at its lower edge (out) at the ascent time $\tilde{t}$. Note that this means that we are not talking about surface precipitation but precipitation out of the Lagrangian air parcel. We refer to $qx_{\text{in/out}}(\tilde{t})$ as *fluxes*. Calculating precipitation in this way ensures that only precipitation formed (or grown) *within* the parcel is taken into account. A net precipitation rate of 0 therefore means that any precipitation entering the parcel from above does not collect (and thus remove) any additional water and is equal to the precipitation leaving the parcel at its lower edge.

B4 Additional microphysical variables for derivation of PE and CR

B4.1 Lagrangian hydrometeor growth term (C_{hy})

The hydrometeor growth term is the sum of all microphysical processes that convert moisture into hydrometeors. It is given by

$$C_{\text{hy}}(\tilde{t}) = \text{cond}(\tilde{t}) + qcnuc(\tilde{t}) + qihh(\tilde{t}) + \text{depo}(\tilde{t}), \quad (\text{B12})$$

where we have the sum of condensation (cond), cloud condensation nuclei (CCN) nucleation (qcnuc), homogeneous and heterogeneous ice nucleation (qihh), and deposition (depo) (see Supplement).

B4.2 Lagrangian vapour growth term (E_v)

The Lagrangian vapour growth term is the sum over all microphysical processes that increase the vapour content by evaporating ice or sublimating water. It is given by

$$E_v(\tilde{t}) = \text{evap}(\tilde{t}) + \text{revap}(\tilde{t}) + \text{fevap}(\tilde{t}) + \text{subl}(\tilde{t}), \quad (\text{B13})$$

where we have the evaporation of cloud drops (evap); evaporation of rain (revap); sublimation of snow, graupel and hail (fevap); and the sublimation of ice (subl) (see Supplement).

B4.3 Hydrometeor/water vapour tendency ($Q_{\text{hy},\text{tc}}(\tilde{t})$ and $Q_{v,\text{tc}}(\tilde{t})$)

The hydrometeor/water vapour tendency gives the accumulated hydrometeor/water vapour mass transported *out* of the air parcel during the ascent through either the turbulence or the convection scheme. For hydrometeors and vapour, respectively, it is given by

$$\begin{aligned} Q_{\text{hy},\text{tc}}(\tilde{t}) &= qcturc(\tilde{t}) + \sum_{x=c,r,i,s} qxconv(\tilde{t}), \\ Q_{v,\text{tc}}(\tilde{t}) &= qvturc(\tilde{t}) + qvconv(\tilde{t}). \end{aligned} \quad (\text{B14})$$

Here we use the *corrected turbulence tendencies* $qcturc$ and $qvturc$ that account for the strong instantaneous compensation between process rates from the turbulence scheme and

the second call to the saturation adjustment in ICON (see Supplement).

B4.4 Net initial hydrometeor/vapour content (HYD and VAP)

Using these variables (and the numerical hydrometeor/water vapour residuals $\mathfrak{R}_{\text{hy}}(\tilde{t})$ and $\mathfrak{R}_{\text{v}}(\tilde{t})$), we formulate the net initial hydrometeor content,

$$\text{HYD}(\tilde{t}) := Q_{\text{hy}}(0) - Q_{\text{hy,tc}}(\tilde{t}) - \mathfrak{R}_{\text{hy}}(\tilde{t})$$

(with $Q_{\text{hy}}(\tilde{t}) := \sum_{x=\text{c,r,i,s,g,h}} qx(\tilde{t})$), (B15)

and the net initial vapour content,

$$\text{VAP}(\tilde{t}) := Q_{\text{v}}(0) - Q_{\text{v,tc}}(\tilde{t}) - \mathfrak{R}_{\text{v}}(\tilde{t})$$

(note : $\text{HYD}(\tilde{t}) + \text{VAP}(\tilde{t}) = Q_{\text{tot}}(0) - Q_{\text{tc}}(\tilde{t})$). (B16)

These terms are almost equal to the initial hydrometeor/vapour content but take into account additional hydrometeors/vapour transported in or out of the parcels by the turbulence or convection scheme (or numerical residuals). We create these terms because we follow the following logic: hydrometeors that precipitate out of a parcel but were brought in by turbulence, for example, should not be counted as precipitation formed within the parcel. Using only the initial hydrometeor/vapour contents would mean that we could, theoretically, see more precipitation “formation” than we have initial total water, which would be confusing.

B5 Comparison of PE, CR and DR to Eulerian variables

We can compare our Lagrangian definitions of PE, CR and DR_{mphys} to the Eulerian definitions from Miltenberger (2014) by making simplifications. If we let $Q_{\text{tc}}(1)$ and $Q_{\text{hy}}(0)$ go to 0 (i.e. we neglect the initial hydrometeor content, all numerical residuals, and all transport by the turbulence and convection scheme), we get

$$\lim_{Q_{\text{tc}}(1), Q_{\text{hy}}(0) \rightarrow 0} \text{DR}_{\text{mphys}} = \frac{P(1)}{C_{\text{hy}}(1) + E_{\text{v}}(1) + \text{HYD}(1)^0} \cdot \frac{C_{\text{hy}}(1) + E_{\text{v}}(1) + \text{HYD}(1)^0}{\text{VAP}(1)^0} \cdot \frac{\text{VAP}(1)^0}{Q_{\text{tot}}(0) - Q_{\text{tc}}(1)^0} \quad (\text{B17})$$

which reduces to

$$\begin{aligned} \text{DR}_{\text{mphys}}^{\text{Eul}} &= \underbrace{\frac{P(1)}{C_{\text{hy}}(1) + E_{\text{v}}(1)}}_{\text{PE}_{\text{Eul}}} \cdot \underbrace{\frac{C_{\text{hy}}(1) + E_{\text{v}}(1)}{Q_{\text{tot}}(0)}}_{\text{CR}_{\text{Eul}}} \\ &= \text{PE}_{\text{Eul}} \cdot \text{CR}_{\text{Eul}}. \end{aligned} \quad (\text{B18})$$

These definitions are similar to those from Miltenberger (2014). The simplifications we have may also result in $\text{DR}_{\text{mix}} = 0$, meaning that $\text{DR} = \text{DR}_{\text{mphys}}^{\text{Eul}}$. We have thus shown that our Lagrangian definitions of DR, PE and CR can be understood as generalisations of those in the Eulerian framework (or alternatively, the Eulerian definitions from Miltenberger, 2014, are special cases that neglect initial hydrometeor contents, numerical residuals, and transport by turbulence and convection schemes).

Appendix C: General plots

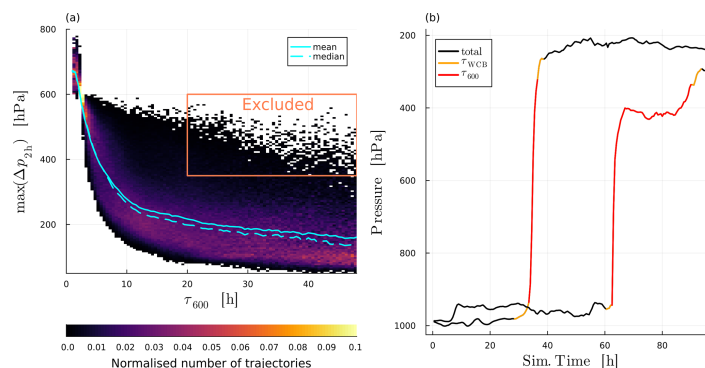


Figure C1. (a) The pressure course for two example trajectories, one convective and one slow, showing that some trajectories ascend convectively but not for the entire 600 hPa. (b) The maximum pressure velocity over 2 h for each trajectory ($\max(\Delta p_{2h})$) as a two-dimensional normalised trajectory over τ_{600} . Trajectories that fall into the orange square are excluded from the analysis.

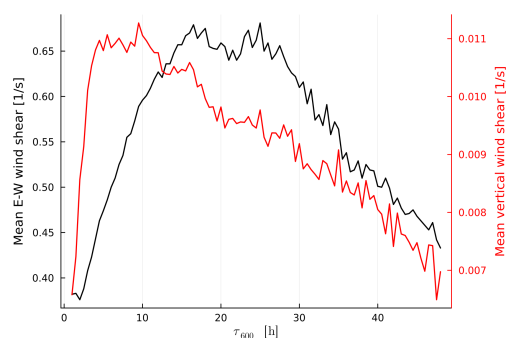


Figure C2. Mean horizontal (black, left y axis) and vertical (red, right y axis) wind shear for all data points along the WCB ascent for trajectories over τ_{600} .

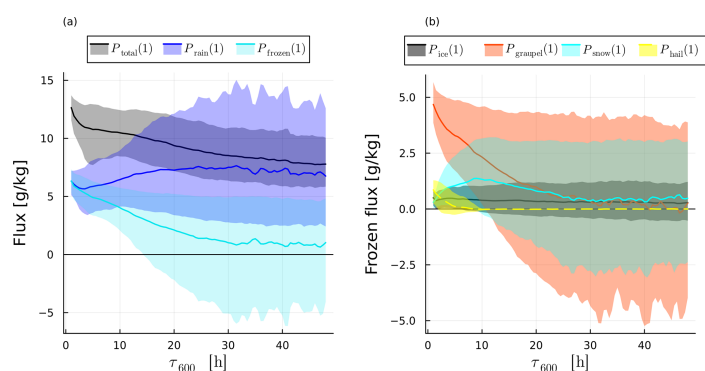


Figure C3. (a) The median total (black), warm-phase (blue) and cold-phase (cyan) precipitation flux by the end of the ascent, with the 10th and 90th percentiles shaded. (b) The individual contributions to the cold-phase flux. Ice (black), graupel (red), snow (cyan) and hail (dashed yellow line; note: zero everywhere except for $\tau_{600} < 10$ h).

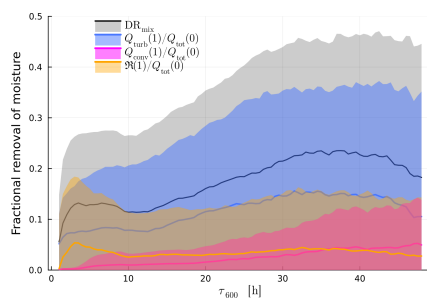


Figure C4. Mean DR_{mix} (black) with individual contributions from turbulence (blue), convection (red) and numerical residual (yellow). The 10th and 90th percentiles shaded.

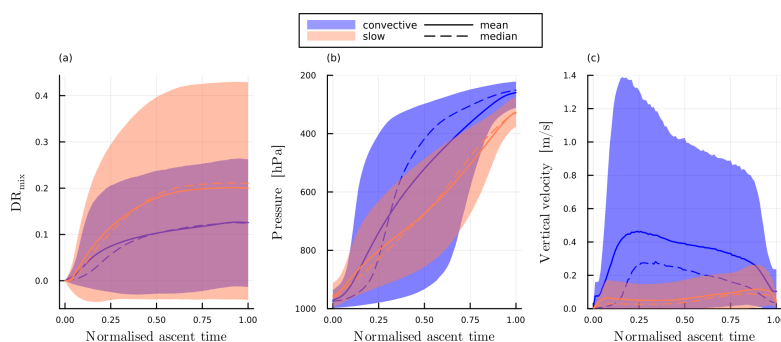


Figure C5. (a) DR_{mix} , (b) pressure and (c) vertical velocity over normalised ascent time. In all panels the mean (solid) and median (dashed) and the 10th and 90th percentiles (shaded) are plotted in blue for convective and orange for slow trajectories.

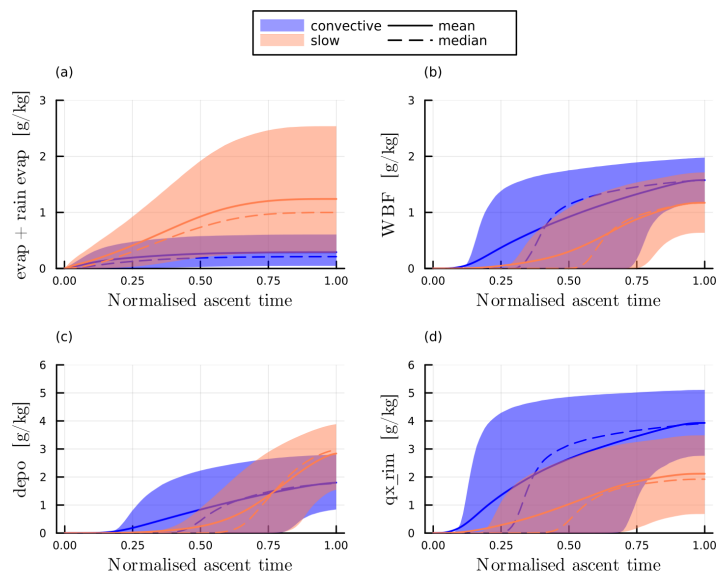


Figure C6. (a) Time-integrated evaporation, (b) Wegener–Bergeron–Findeisen, (c) deposition and (d) riming rate over normalised ascent time. In all panels the mean (solid) and median (dashed) and the 10th and 90th percentiles (shaded) are plotted in blue for convective and orange for slow trajectories.

Code and data availability. The ICON source code is distributed under an institutional licence issued by the German Weather Service (DWD). The model output of the ICON simulation is available from the authors upon request. The code and data necessary to recreate the results shown here are available at <https://doi.org/10.5281/zenodo.13913564> (Schwenk and Miltenberger, 2024).

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Author contributions. CS and AM designed the experiment and conducted the numerical simulations. AM implemented the online diagnostics, and CS wrote the post-processing code. CS and AM worked jointly on the interpretation of the results. CS drafted the final manuscript with contributions from AM.

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The role of ascent timescales for warm conveyor belt (WCB) moisture transport into the upper troposphere and lower stratosphere (UTLS)

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S1 ICON calculations and the two-moment microphysics scheme

In this section we describe which calculations are performed by the two-moment microphysics scheme (Seifert, 2008) in ICON. The operations/calculations are performed in the order in which they are listed. Variables in bold are of particular interest for us.

- Update of tracer fields and slow-physics tendencies
At each fast-physics time step, ICON first updates the tracer fields and then fetches the slow-physics tendencies from the previous time step. The turbulent transfer coefficients for the atmosphere-surface interface are then computed.
- First saturation adjustment (**satad_I**)
In this step all sub- or supersaturation generated in the current time step is removed by evaporating or condensing cloud droplets. This establishes thermodynamic equilibrium between water vapour and liquid water in a process that is technically instantaneous, i.e. it is assumed to take place on a time scale much smaller than the fast-physics time step. The latent heating resulting from **satad_I** is used to update the temperature and exner pressure.
- Turbulent diffusion (**qxtrub** for $x = v, c, i$)
The turbulence scheme Doms et al. (2021) (Chapter 3) developed by Matthias Raschendorfer calculates the turbulent diffusion in the atmospheric column for vapour, cloud droplets and ice.

ICON then continues with the two-moment cloud microphysics scheme that does the following calculations (in order):

- CCN activation (**qcnuc**)
The activation of cloud condensating nuclei (CCN) is computed through a parameterisation adapted and modified from Hande et al. (2016).
- Homogeneous and heterogeneous ice nucleation (**qihh**)
The parameterisation for homogeneous freezing is based on Kärcher et al. (2006) and Kärcher and Lohmann (2002) and the parameterisation for heterogeneous freezing is modified from Hande et al. (2015) to also include immersion and deposition freezing.
- Homogeneous freezing of cloud droplets (**qihom**)
The parameterisation for the homogeneous freezing of cloud droplets is taken from Jeffery and Austin (1997). This scheme exponentially increases the freezing rate with decreasing temperature and freezes all droplets immediately if the temperature drops below -50°C .
- Depositional growth of all ice particles (**qxdep**)
In this step, both vapour deposition and ice sublimation are calculated individually for each frozen hydrometeor species using a relaxation timescale approach based on Morrison et al. (2005).
- Collisional growth of frozen hydrometeors (not written to file)
ICON then calculates the particle-particle and self-collections of all frozen hydrometeor species as well as the wet growth and conversion of graupel to hail.
- Riming of frozen hydrometeors with cloud droplets and rain drops (**qxrim**)
For ice and snow there are two separate schemes for calculating riming with both cloud and rain drops. Hail and graupel are treated together, with two schemes for cloud-drop and rain-drop riming. The sum of these processes is taken. These calculations are based on Seifert and Beheng (2005).
- Freezing of rain and conversion to ice/graupel/hail (**rfrez**)
The calculated heterogeneous freezing rates of rain drops, based on Bigg (1953), increase with raindrop radius and decreasing temperature, with all raindrops instantly freezing below -40°C .

- Melting of frozen hydrometeors to cloud droplets (**qcmelt**)
If the temperature rises above 0° C, all ice particles are immediately melted to form cloud drops.
- Melting of frozen hydrometeors to rain (**qrmelt**)
Snow, graupel and hail melting are treated separately and the melting rate is explicitly calculated. If the temperature rises above 10° C, all snow particles immediately melt into rain.
- Evaporation of snow, graupel and hail (**fevap**)
The liquid forming on the surface of melting hydrometeors can evaporate. This calculation is similar to rain evaporation at 0° C and is based on Seifert and Beheng (2005).
- Raindrop growth (not written to file)
Here the auto-conversion and accretion (Beheng, 1994), the self-collection (Seifert and Beheng, 2001) and the break-up of rain (Seifert, 2008) are calculated.
- Evaporation of rain (**revap**)
If the air is sub-saturated, this final computation calculates the evaporation of raindrops according to Seifert (2008).
- Size limits for all hydrometeors are enforced.

End of two-moment cloud microphysics scheme. ICON then resumes with:

- Sedimentation (**qxin** and **qxout** for $x = r, i, s, g, h$)
Different computations are implemented to calculate the sedimentation of (i) rain, (ii) snow and ice, and (iii) graupel and hail according to Blahak (2020).
- Second saturation adjustment (**satad_II**)
Any sub- or supersaturation resulting from the microphysical calculations is dealt with by a second saturation adjustment using the same procedure as the first.
- Slow physics: convective tendencies (**qxconv** for $x = v, c, r, i, s$)
While the previous steps are completed at each fast-physics time step, the subsequent slow-physics processes are completed at reduced time intervals. We are interested in the Tiedke-Bechthold convection scheme, which is switched on for the global domain and produces mass-flux tendencies for vapour, cloud droplets, rain drops, ice particles and snow (Tiedtke, 1989; Bechtold et al., 2008). Other slow-physics processes not relevant to the scope of this work include radiation (calculated on a coarser resolution grid) and non-orographic and orographic gravity wave drag.

S2 Diagnostics and corrections

Note all diagnostics are computed on the instantaneous rates interpolated to the trajectory position at each model physics timestep.

S2.1 Computation of evaporation and condensation rates from first call to saturation adjustment

Condensation and evaporation rates are diagnosed from the tendency from the first call to the saturation adjustment scheme according to its sign:

```

1: if satad_I > 0:
2:     cond = cond + satad_I
3: elseif satad_I < 0:
4:     evap = evap + satad_I

```

S2.2 Computation of deposition rate, mass transfer rate by the Wegener-Bergeron-Findeisen processes, and “corrected” turbulence tendencies

In the ICON model deposition onto and sublimation of frozen particles, saturation adjustment, and the turbulence parameterisation determine the partitioning of water vapor and condensate. Saturation adjustment is called at the beginning of the microphysics timestep, then tendencies due to the deposition and turbulence parameterisations are computed followed by a second call to saturation adjustment. Hence, the redistribution of water mass from liquid to frozen hydrometeors by the Wegener-Bergeron-Findeisen (WBF) process effectively manifests as depositional growth of hydrometeors followed by evaporation in the second call of the saturation adjustment scheme. While having implications for latent heating, the WBF process does not result in condensate loss or gain and therefore its contributions to evaporation and deposition should be excluded from the moisture budget diagnostics. How this is done is documented below.

A second correction to the raw model tendencies was deemed necessary for a meaningful analysis of the moisture budget: In the ICON there is a strong compensation between tendencies to water vapor and cloud mass mixing ratio from the turbulence scheme and the second call to the saturation adjustment. Despite the root cause of this feature not clear yet, this compensation happens within the same timestep and therefore is thought to have no or limited physical meaning. We used a “correction” of the turbulence tendencies removing instantaneous compensation between turbulence and saturation adjustment tendencies as outlined below.

```
1: if |qxdep| < 1e-15 (i.e equal to zero):
2:     % There is no depositional growth and hence no WBF. We just need to account for the compensation between tendencies
3:     % from turbulence and second saturation adjustment.
4:     qvturc = qvturb - satad_II
5:     qcturc = qcturb + satad_II
6:
7: elseif qxdep > 1e-15:
8:     % There is depositional growth. WBF, deposition, and "corrected" turbulence tendencies need to be computed.
9:
10:    if qcturb + satad_II > 0:
11:        % There is net condensation. This should not happen based on microphysics only (deposition should push relative
12:        % humidity to below water saturation). Therefore all deposition is counted as mass transfer from water vapor
13:        % to condensate and the net cloud condensate rate (sum of original turbulence and second saturation adjustment
14:        % rate) is attributed to the turbulence scheme. Note this may overestimate the turbulence contribution, if the
15:        % tendencies are partially down to WBF.
16:        qvturc = qvturb - satad_II
17:        qcturc = qcturb + satad_II
18:        depo = qxdep
19:
20:    elseif qcturb + satad_II <= 0:
21:        % There is net evaporation, WBF is redistributing condensate mass. Note this may overestimate the role of WBF,
22:        % if net evaporation is down to turbulent mixing only.
23:        if qcturb - satad_II > qxdep:
24:            % There is more evaporation of cloud condensate than deposition on ice, i.e. turbulent processes and WBF
25:            % influence the gas-condensate partitioning. We assign the full deposition tendency to WBF mass transfer
26:            % and the rest of the net cloud condensate tendency to turbulence.
27:            qvturc = qvturb - satad_II - qxdep
28:            qcturc = qcturb + satad_II + qxdep
29:            wbf = qxdep
30:        elseif qcturb - satad_II <= qxdep:
31:            % There is less evaporation of cloud condensate than deposition on ice, i.e. there is more depositional growth
32:            % than can be explained by WBF. We assign all evaporation of cloud condensate to the WBF mass transfer
33:            % and the rest of the deposition tendency to actual mass transfer between the gas phase and condensate.
```

```
34:          % There is no tendency from turbulence of cloud condensate, only for water vapor.
35:          qvturc = qvturb + qcturb
36:          depo = qxdep + satad_II + qcturb
37:          wbf = -satad_II - qcturb
38:
39: elseif qxdep < -1e-15:
40:     % Frozen hydrometeors are sublimating, hence WBF is not active. There is no physical reason for the cloud microphysics
41:     % to produce super-saturated conditions. The tendency from deposition parameterisation is assigned to the sublimation
42:     % rate. The net rate from turbulence and second saturation adjustment is assigned to the "corrected" turbulence
43:     % tendency.
44:     qvturc = qvturb - satad_II
45:     qcturc = qcturb + satad_II
46:     subl = qxdep
```

Note that the re-assignment of rates leaves the net tendency of condensate and water vapor at the end of the physics timestep unchanged.

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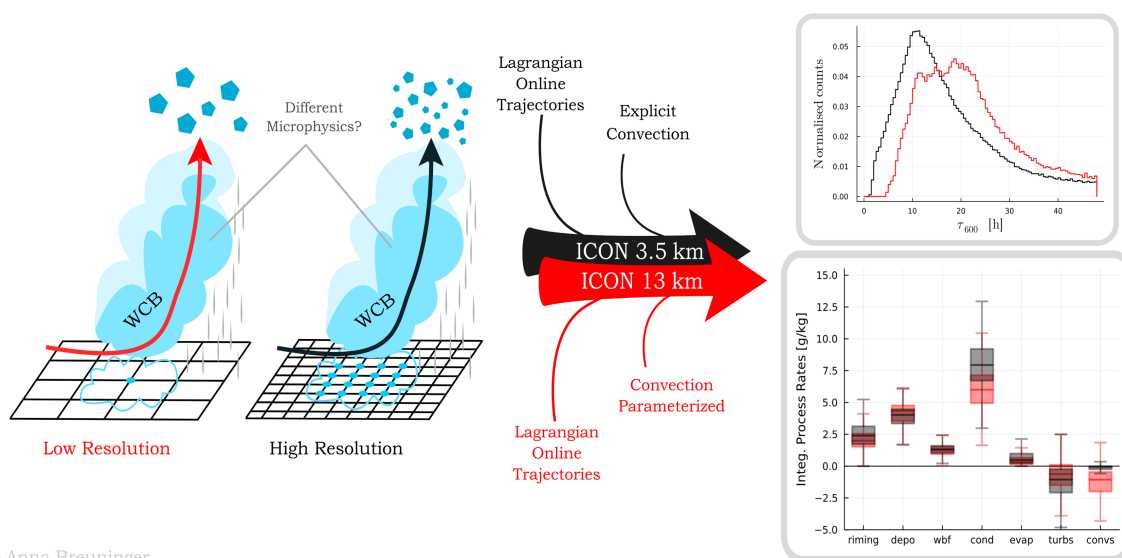
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Chapter 4

The Role of Scale - Part I: Lagrangian Perspective

Research Question IIa

How do the modeled moisture transport, the governing microphysical processes, and the radiative effects of WCBs differ when transitioning from a convection-parameterizing simulation to a convection-permitting one?



This Chapter is in preparation to be submitted to *Atmospheric Chemistry & Physics* under the title "Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS)— Part I: Lagrangian Perspective" (Schwenk and Miltenberger, 2025a). For author contributions, see Section 2.5.

Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS)— Part I: Lagrangian Perspective

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Abstract. Warm conveyor belts (WCBs) are key features of extratropical cyclones that transport water vapour and hydrometeors into the upper troposphere and lower stratosphere (UTLS), where they influence Earth’s radiative balance. Previous studies have shown that horizontal grid spacing strongly affects WCB characteristics, including ascent speed, diabatic heating, and end-of-ascent pressure, especially when transitioning from convection-parameterising to convection-permitting resolutions. It has also been shown that convectively ascending air parcels undergo distinct microphysical processes and have different properties in the outflow compared to those that ascend more slowly. In this first part of a three-part study, we use Lagrangian online trajectories to investigate how grid spacing impacts microphysical processes and the transport of moisture and hydrometeors into the UTLS by a WCB. We analyse two ICOSahedral Nonhydrostatic (ICON) simulations — one convection-parameterising (~ 13 km) and one convection-permitting (~ 3.5 km) — of a WCB case that occurred over the north Atlantic on 23 September 2017. We hypothesize that key differences arise from the higher vertical velocities in the high-resolution simulation. We find that across all ascent timescales, the convection-permitting simulation produces more rapid ascent. The WCB outflow in the convection-permitting simulations is also drier, with lower specific and relative humidity, which we attribute to higher ice number concentrations that induce shorter relaxation timescales for supersaturation over ice. Additionally, the high-resolution simulation shows more pronounced frozen-phase processes (e.g., riming), stronger frozen precipitation, and notable differences in hydrometeor mass mixing ratios, number concentrations, and radii — indicating that horizontal resolution significantly influences modelled WCB cloud composition. These findings suggest that weather and climate models using convection-parameterising resolutions may misrepresent WCB cloud properties and outflow humidity, potentially affecting the representation of Earth’s radiative budget.

1 Introduction

Water vapour is the most dominant greenhouse gas (Schneider et al., 2010), and in the upper troposphere / lower stratosphere (UTLS), variations in its concentration drive one of the most significant positive climate feedback mechanisms (Li et al., 2024; Held and Soden, 2000; Dessler et al., 2013, e.g.). Even minor changes in UTLS water vapour can substantially impact Earth’s radiative budget (Wang et al., 2001; Hansen et al., 1984) and influence both mean and regional circulation patterns (Charlesworth et al., 2023; Ploeger et al., 2024, e.g.). However, even though water vapour measurements in the UTLS have increased over the past 20 years (Zahn et al., 2014; Jeffery et al., 2022; Hurst et al., 2011; Tilmes et al., 2010; Konjari et al., 2024), it remains one of the most poorly characterised regions of the atmosphere, mainly due to challenging measurement conditions (Jeffery et al., 2022).

Nevertheless, in recent years multiple studies have tried to address this gap and to quantify the moisture content of the UTLS and associated transport pathways.

Several modelling studies (Ueyama et al., 2018, 2023; Dauhut et al., 2018; Hassim and Lane, 2010) and experimental studies (Corti et al., 2008; Danielsen, 1993; Lee et al., 2019; Gordon et al., 2024) have shown that deep convection in the tropics and the sublimation of detrained ice crystals are a major source of UTLS moisture. Homeyer et al. (2024) and Homeyer et al. (2014) specifically addressed the UTLS over the continental United States and confirmed that in this (limited) sample of the extratropics, overshooting convection remains an important contributor to UTLS moisture. Future warming of the tropopause is expected to increase UTLS moisture levels (Gettelman et al., 2010), while enhanced sublimation of ice stemming from convective activity is also projected (Dessler et al., 2016). Historically, transport directly into the tropopause region has increased from 2011 to 2020 compared to the 1980s (Jeske and Tost, 2025), but it is unclear how this will change in the future. Convection is however not the only transport pathway for UTLS moisture.

In addition to convection, warm conveyor belts (WCBs) were also identified as a major source of moisture in the extratropical UTLS (Zahn et al., 2014; Guo and Miltenberger, 2025). WCBs are large-scale ascending air streams that develop in extratropical cyclones (ETCs). They produce the elongated cloud bands typically observed in these systems, and over the course of roughly two days, WCBs transport moist boundary-layer air poleward and upward into the upper troposphere. During this time, the ascending air parcels experience complex microphysical and dynamical processes that lead to the formation of clouds and precipitation, affect upper-level wave propagation, and introduce both water vapor and hydrometeors into the UTLS (Madonna et al., 2014). Consequently, WCBs shape mid-latitude weather patterns, guide storm evolution, and influence Earth's radiative balance (Madonna et al., 2014; Rodwell et al., 2018; Joos, 2019). The large amounts of precipitation formed by WCBs often create hazardous conditions (Pfahl et al., 2014), with WCBs estimated to account for over 80% of total precipitation in Northern Hemispheric storm tracks (Pfahl et al., 2014; Eckhardt et al., 2004). Therefore, for weather forecasts it is important that models accurately represent WCBs, since their incorrect representation has been identified as a key factor in increasing forecast uncertainty (Pickl et al., 2023; Berman and Torn, 2019; Rodwell et al., 2018; Grams et al., 2018).

However, correctly representing WCBs is also important for climate forecasts (Joos, 2019). In the early stages of their lifetime, when WCBs are positioned in the southern extra-tropics, the ascending air produces low-level clouds that have a predominantly cooling effect. As the WCB moves northward and the air rises further, high-level frozen clouds have either a warming or a cooling effect, depending on solar radiation and cirrus optical thickness (Krämer et al., 2020; Joos, 2019). Mixed-phase clouds, present in the transition phase, contribute an uncertain net radiative effect. In the out-flow region, WCBs are often associated with ice-supersaturated air masses (Spichtinger et al., 2005) and a cirrus cloud shield extending up to 3 km in depth (Binder et al., 2020), which usually results in a warming effect. A recent climatological study from Guo and Miltenberger (2025), which analysed ERA5 reanalysis data from 2010 to 2019, determined that, in December, January, and February, grid-scale advection accounts for up to 13.8% of the total moisture transport by WCB into the upper troposphere in the main ascent regions over the North Atlantic and Pacific ocean basins. However, it remains difficult to precisely quantify the amount of water vapour WCBs transport into the UTLS — and how this amount will evolve in a warming climate. As a result, the overall radiative impact of WCBs remains difficult to quantify, and how the balance between cooling and warming effects may shift under future climate conditions remains an open question (Joos, 2019).

Several factors make the accurate modelling of WCBs challenging. As ascending air undergoes warm-phase, mixed-phase, and cold-phase microphysical processes (Binder et al., 2020; Forbes and Clark, 2003; Gehring et al., 2020), the resulting clouds span a wide range of temperatures and pressures and exist in different phases. This variability adds significant complexity to their representation in models (Oertel et al., 2023; Schwenk and Miltenberger, 2024; Hieronymus et al., 2025). Therefore, it is unsurprising that the parameterisation of the microphysical processes in WCBs are a major source of numerical weather prediction (NWP) model uncertainty (Morrison et al., 2020; van Lier-Walqui et al., 2012; Posselt and Vukicevic, 2010). For instance, sensitivity experiments have shown that the choice of microphysical parameterisation schemes significantly influences WCB ascent characteristics (Mazoyer et al., 2021, 2023). Additionally, the diabatic processes occurring during WCB ascent are highly sensitive to the specific cloud microphysical parameters used within a given scheme (Hieronymus et al., 2022; Neuhauser et al., 2023; Forbes and Clark, 2003; Oertel et al., 2025). Schwenk et al. (2025) also show that changes in microphysical parameter choices in NWP models (such as the capacitance of ice and snow) within physically plausible ranges significantly alter the microphysical composition and ice content in the outflow of a WCB. These studies demonstrate the complexity of microphysical processes in WCBs and the substantial challenges they pose for accurate model representation.

Another factor that adds complexity and has recently gained attention is the role of embedded convection within WCBs (Rasp et al., 2016; Oertel et al., 2020; Schwenk and Miltenberger, 2024). Oertel et al. (2020) found that convective regions in WCBs differ from classical synoptic-scale slantwise ascending regions by producing enhanced surface precipitation, more intense cloud diabatic heating, and distinct imprints on the PV structure of the WCB outflow. Additionally, they observed that convective air parcels do not ascend continuously but experience intermittent periods of strong ascent and even descent. This finding was supported by Schwenk and Miltenberger (2024), who further showed that convective air parcels transport substantially more ice into the UTLS than their slower-ascending counterparts and experience different microphysical processes during the ascent. They also identified differences in relative humidity distributions between fast- and slow-ascending air parcels, despite vapour conditions being primarily controlled by temperature. These studies, based on high-resolution convection-resolving simulations, raise an important question: Can low-resolution, convection-parameterising simulations adequately represent WCBs, particularly their role in moisture transport into the UTLS?

A number of studies have found that increasing model resolution (especially when moving from a resolution that requires convective parameterisation to one that explicitly resolves some fraction of convective activity) changes and improves NWP model results. Senf et al. (2020) showed that refining the grid spacing up to 2.5 km significantly improved simulated top-of-atmosphere cloud-radiative effects over the North Atlantic. Vergara-Temprado et al. (2020) found that fine-resolution models (exploring ranges from 50 km down to 2.2 km) with deep convection treated explicitly led to notable improvements in precipitation patterns and the diurnal cycle in annual simulations of the European climate. Homeyer (2015) determined that when changing the vertical and horizontal resolutions within the convection permitting regime from 3 km to 0.33 km for simulations of extratropical convection, the depth of overshooting and cross-tropopause transport increased with horizontal and decreased with vertical model resolution.

Focusing on WCBs, Choudhary and Voigt (2022) found that as grid spacing decreased gradually from 80 km to 2.5 km, more WCB air parcels ascended faster and higher into the atmosphere. They also observed an increase in diabatic heating at all pressure levels, primarily driven by cloud-associated

phase changes in water. These results suggest that the representation of microphysical processes during WCB ascent are resolution-dependent, potentially reflecting the influence of convective air parcels. This is supported by the fact that the widely used Tiedtke-Bechtold convection parameterization scheme only provides a basic representation of cloud microphysics (Tiedtke, 1989; Bechtold et al., 2008). Additionally, Schwenk and Miltenberger (2024) showed that convectively ascending WCB air parcels undergo distinct microphysical processes than slowly ascending ones.

In spite of these various studies, so far, no study has examined how WCB moisture transport changes when transitioning from convection-parameterising to convection-permitting resolutions — a gap this paper aims to address. Therefore, our research question is: How does increasing model resolution from convection-parameterising (13.6 km effective grid spacing) to convection-permitting (3.2 km effective grid spacing) affect the transport of moisture and hydrometeors into the UTLS by a WCB? Given the results from Schwenk and Miltenberger (2024), we hypothesize that many of the differences observed between the simulations are primarily driven by changes in vertical velocity associated with higher model resolution.

This paper primarily focuses on comparing the representation of WCB processes in convection-parametrised and convection-permitting models, and analysing the resulting differences along WCB trajectories. Note that in the Lagrangian perspective, the differences are necessarily only identified in air parcels undergoing WCB ascent based on grid-scale wind. Therefore, the impact of the convection parameterisation is only seen as an additional tendency in the Lagrangian budget if it affects WCB parcels. To address this issue and focus more on the overall outflow properties, we present an Eulerian analysis of the same simulations in the second part of the paper. Despite the limitations of the Lagrangian perspective, we shall see that it offers important physical process-oriented insight into the differences between the simulations, which cannot be gained by Eulerian analysis only.

This paper is structured as follows: First, we will describe the WCB case, the model setup, the online trajectories, and the Lagrangian diagnostics performed. This part is similar to Schwenk and Miltenberger (2024), since this is a follow-up study. We then show the results of the Lagrangian analysis, where we compare the trajectories for the global and the nested simulation. Since the Lagrangian data does not represent all relevant variables and does not necessarily fill the entire WCB domain, potentially leading to biased results, an Eulerian analysis is undertaken in Part-II (Schwenk and Miltenberger, 2025a) of this study. This study largely reinforces the findings from this paper with the addition of investigating differences in radiation.

2 Methods

In this study we examine two simulations of a WCB case study, run at convection permitting and convection parameterising resolutions. This work builds on the study by Schwenk and Miltenberger (2024), since we examine the same case study and in the case of the convection permitting simulation, the same simulation output. The model setup for the convection parameterising simulation is very similar. Therefore, we will provide only a brief overview of the WCB case, model setup, and Lagrangian data analysis, and direct the reader to Schwenk and Miltenberger (2024) for further details.

2.1 WCB Case, Model Setup and Online Trajectories

This study investigates the same WCB case as Schwenk and Miltenberger (2024). From the afternoon of 22 September 2017 to the end of 23 September 2017, a strong cyclone with a characteristic WCB

band dominated the weather over the northern Atlantic. The initial low pressure system formed off the coast of Newfoundland (approximately 52°N and 45°W) at 04:00 UTC on 21 September 2017, and underwent explosive cyclogenesis to form an extratropical cyclone over the central northern Atlantic 24 hours later at 04:00 UTC on 22 September 2017. On the morning of 23 September 2017, the northern edge of the cyclone reached Iceland and the eastern edge Ireland. By the evening it dissipated over northern Europe (for figures see Schwenk and Miltenberger (2024)).

We simulated this event using the Icosahedral Nonhydrostatic (ICON) model (v2.6.2; Zängl et al. (2014)). Two simulations were initialised with the operational ICON global analysis at 00:00 UTC on 20 September 2017, running for 96 hours until 00:00 UTC on 24 September 2017, when the WCB dissipated over northern Europe. The first simulation, from now on referred to as the "nested" simulation, has a setup which includes a global domain with two nested high-resolution grids, covering the main WCB ascent region (Fig. 1 b, Fig. A4 a). The global domain uses a R03B07 grid (~13 km effective grid spacing), while the nested domains employ R03B08 (~6.5 km) and R03B09 (~3.3 km) grids and are linked to each other and the global domain by two-way coupling (Zängl et al., 2022). In the higher-resolution domains only shallow convection is parametrised, whereas in the global domain also deep convection is parametrised by the Tiedtke–Bechtold scheme (Tiedtke, 1989; Bechtold et al., 2008). Other sub-grid processes, including turbulence, orographic drag, and radiation, follow standard ICON schemes. Cloud microphysics are simulated with the two-moment scheme by Seifert and Beheng (2005), representing six hydrometeor species (cloud droplets, rain, ice, snow, graupel, and hail). The second simulation, from now on referred to as the "global" simulation, has the same model set-up and only differs in that it only has one global domain on a R03B07 grid (~13 km resolution; Fig. 1 a) and therefore parameterises deep convection using the Tiedtke–Bechtold scheme everywhere.

To calculate the Lagrangian data we used the online trajectory module in ICON, developed by Miltenberger et al. (2020) and extended by Oertel et al. (2023) for domain nesting. In both simulations, trajectories were initiated every 5 hours throughout the 96-hour simulation, with output recorded every 30 minutes. Starting positions were randomly selected within a designated region across six vertical levels (~1000–800 hPa). This region was determined using prior offline ERA5-based trajectories to ensure robust WCB representation (Fig. 1).

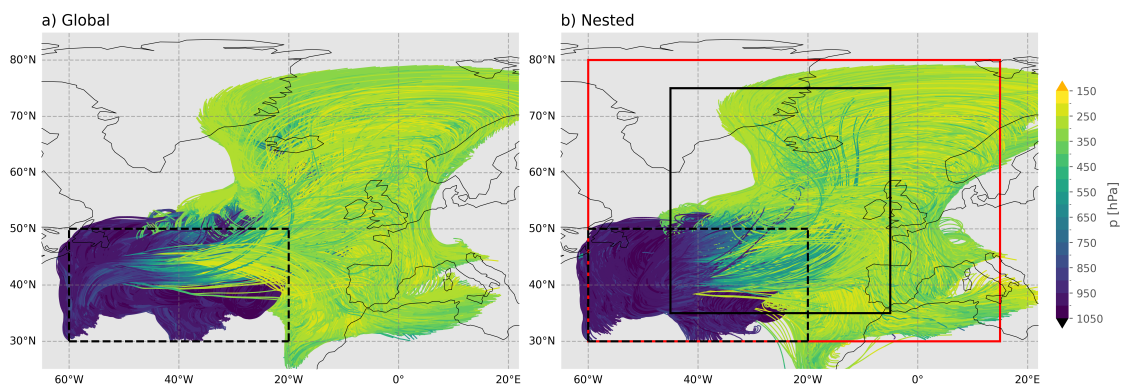


Figure 1. Simulation setup as well as pressure, latitude and longitude course of selected WCB trajectories for every fifth trajectory in the global simulation in a), and every tenth trajectory in the nested simulation in b). The online trajectory starting area is marked by the dashed lines, nested domain 2 (red) and 3 (black) boundaries are indicated in b) by the solid lines.

2.2 WCB trajectory selection, ascent timescales, time-integrated process rates

Our selection algorithm identifies WCB trajectories based on two criteria. First, a WCB trajectory must ascend at least 600 hPa within 48 hours (a common definition of WCB ascent, see Madonna et al. (2014); Oertel et al. (2023)), and second, the trajectories must lie within two specified longitude-latitude regions at two distinct times, determined using Eulerian cloud cover and sea-level pressure data from our simulation. This ensured the trajectories were part of the WCB and not from unrelated mesoscale convective systems (MCSs) or other cyclones. This selection algorithm resulted in 113 993 WCB trajectories (from in $15.5 \cdot 10^6$ trajectories overall) for the global simulation, and 393 070 (from in $91.8 \cdot 10^6$ trajectories overall) in the nested simulation.

For the characterization of trajectory ascent time we use the common ascent timescale τ_{600} (τ_{300}) defined as the fastest time in which a trajectory ascends 600 (300) hPa (Rasp et al., 2016; Oertel et al., 2023; Schwenk and Miltenberger, 2024). To define the beginning and end of the ascent, we use τ_{WCB} , which is defined by extending the τ_{600} -period backwards and forwards as long as the ascent velocity remains above 8 hPa h^{-1} (Schwenk and Miltenberger, 2024; Guo and Miltenberger, 2025). During τ_{WCB} , a trajectory is defined as ascending within the WCB, before (after) this time it is defined as being part of the WCB inflow (outflow). To quantify the convection experienced per trajectory (during the ascent), we use the maximum pressure difference experienced by the trajectory in any 2 h time span ($\max(\Delta p_{2\text{h}})$).

As in Schwenk and Miltenberger (2024), microphysical-process rates (such as deposition or riming) are set to zero at the start of τ_{WCB} and accumulated until the end of the ascent. The instantaneous process rates from ICON have the units $\text{kg kg}^{-1} \text{s}^{-1}$; we additionally have "integrated" process rates with the units kg kg^{-1} along air mass trajectories (see Schwenk and Miltenberger (2024) for details).

2.3 Precipitation Efficiency (PE), Condensation Ratio (CR)

Two useful metrics to quantify the removal of moisture from an air parcel are the condensation ratio (CR) and the precipitation efficiency (PE) (Barstad et al., 2007; Miltenberger, 2014; Dacre et al., 2023). To remove water vapour by microphysical processes (as opposed to turbulence), it must be converted to hydrometeors that precipitate. The efficiency with which the conversion takes place is quantified by CR, and the efficiency with which the hydrometeors precipitate by PE. In this paper, we use the Lagrangian definitions of CR and PE from Schwenk and Miltenberger (2024) that take into account all relevant microphysical processes, as well as moisture lost or gained by the turbulence parameterisation, the convection parameterisation and numerical residuals. In formulation by Schwenk and Miltenberger (2024), CR represents the fraction of water vapour that is present at the start of the ascent, or is brought in by turbulence, convection, or numerical residuals, and that is converted into hydrometeors by the end of the ascent. PE measures the fraction of hydrometeors that either formed during the ascent, were present at the start, or were carried in by turbulence, convection, or numerical residuals, and that are precipitated by the end of the ascent. Both PE and CR are bounded by 0 and 1. For the full definitions and derivations see Schwenk and Miltenberger (2024).

3 Differences in properties of WCB trajectories from convection-permitting and -parameterising simulations

In this section, we use the online trajectory data to analyse and compare the ascent characteristics and moisture transport into the UTLS by the WCB in both the global and nested simulations. Since this

study is a follow-up to Schwenk and Miltenberger (2024), the results for the convection-permitting simulation and their physical interpretation are largely identical to Schwenk and Miltenberger (2024).

3.1 Ascent characteristics and evolution over the WCB lifetime

When comparing the ascent timescales τ_{600} and τ_{300} , the nested and the global simulation are very different (Fig. 2). The distributions of τ_{600} -values are of a similar shape in both simulations (Fig. 2 a), with a single peak, a tail to larger values, and a sharp decline for smaller values. However, in the global simulation, peak is more flat, and the distribution is shifted to larger values, indicating a slower ascent overall. This is reflected in the mean (median) τ_{600} , which is 17.0 h (14.5 h) and 21.4 h (20.0 h) for the nested and the global simulation, respectively. Because of this shift, the minimum τ_{600} -value in either distribution is 4.5 h for the global simulation, compared to 0.5 h for the nested simulation. This means that the global simulation has virtually no convective trajectories when using the definition from Schwenk and Miltenberger (2024), which considers a trajectory to be dominated by convective motion if $\tau_{600} < 5$ h.

The τ_{300} distribution is also shifted to larger values for the global simulation (Fig. 2 b). Both distributions hint at bi-modality, with a primary peak at smaller τ_{300} and a secondary peak at larger values. However, the primary peak is much more pronounced in the nested simulation than in the global simulation. This indicates that in the global simulation, much fewer trajectories experience the short convective bursts identified by Schwenk and Miltenberger (2024) for the nested simulation. This is further supported by the $\max(\Delta p_{2h})$ -values (Fig. A7 a), showing that in the global simulation, virtually no trajectory ascends more than 350 hPa in 2 hours, whereas in the nested simulation plenty of trajectories ascend even more than 600 hPa in the same time (Fig. A7 a). We also find that in every τ_{600} -bin, the distribution of $\max(\Delta p_{2h})$ is shifted to smaller values for the global simulation, indicating a generally smaller abundance of convective behaviour irrespective of the total time needed for a 600 hPa ascent. Finally, when binning vertical velocities experienced by trajectories during the main ascent time-frame by pressure (Fig. A7 b), we find that they are the most different between 500 and 900 hPa, which is consistent with observational studies locating convective activity in the lower to mid-troposphere (Blanchard et al., 2021). Above 400 hPa, vertical velocities in the nested simulation are only slightly larger.

These results suggest that in the global simulation, where convection is parameterised, WCB trajectories experience significantly less convective motion, and depending on the definition, may experience no convection at all. This is not surprising, as the convection parameterization scheme does not result in higher vertical velocities in the ICON model's wind fields, which the online trajectory model relies on to calculate the paths of ascending air parcels. In other words, in a convection parameterising simulation, the trajectories do not "sense" convection in terms of vertical velocities.

Nevertheless, when we compare the instantaneous Δp_{2h} -values for each trajectory at a certain simulation run-time (Fig. 3 a) we see that both simulations exhibit large-scale ascending behaviour at similar times in the synoptic evolution of the associated cyclone. In our analysis, we identify three time-frames (TF) given the Δp_{2h} -values (Fig. 3 a) and geographic positions of trajectories (Figs. A3, A4 and A5) which each exhibit different large-scale ascending behavior. During TF1 (from 12:00 UTC 2017-09-20 to 10:00 UTC 2017-09-21; Fig. A3) trajectories are mainly located in the southern north Atlantic, are close together and are just beginning the large-scale ascent. However, most of this initial ascent is located outside of domain 3, where we want to focus our analysis (differences between the convection-parameterising and -permitting simulation will be most prominent in domain 3). In TF1, the mean (median) Δp_{2h} for both simulations does not exceed -10 hPa (-1 hPa).

In TF2 (00:00 UTC 2017-09-22 to 06:00 UTC 2017-09-23; Fig. A4) we see the strongest large-scale ascent that mostly occurs within domain 3, which is why we will focus most of the subsequent analysis on TF2. The mean (median) Δp_{2h} reaches values of approximately -32 hPa (-10 hPa) for the nested and -22 hPa (-12 hPa) for the global simulation. During TF2, the ascent is more vigorous in the nested simulation, with the largest 95th percentiles for Δp_{2h} of -150 hPa being more than 50 hPa larger (i.e. more negative) in the nested simulation than in the global simulation (Fig. 3 a). Histograms of pressure at the beginning and end of TF2 indicated that most trajectories complete their ascent within TF2 (Fig. 3 b). TF3 (10:00 UTC 2017-09-23 to 20:00 UTC 2017-09-23; Fig. A5) notes the time towards the end of the WCB lifetime, where most of the trajectories have completed their ascent, reached the UTLS and spread to the North and South and have propagated often times to regions outside of domain 3 (Fig. A5 a). TF3 is interesting because by this time most trajectories are part of the WCB outflow and they mostly remain in the UTLS at pressures below 400 hPa (Fig. A5 b). TF3 therefore allows us to investigate the outflow properties of the WCB.

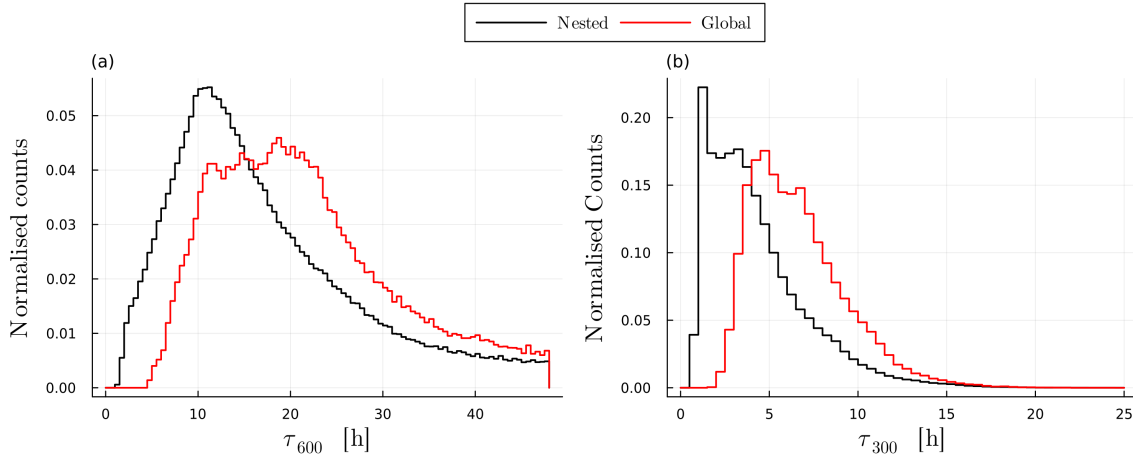


Figure 2. Histograms of τ_{600} (a) and τ_{300} (b) for the WCB trajectories in the global simulation (red) and nested simulation (black).

Considering diagnostics separately for the three TFs allow us to focus on distinct areas of interest: i) the southern early stages of the WCB, where low- to mid-level liquid clouds dominate (not a focus in this study); ii) the period of strongest ascent, where a full range of mixed-phase clouds exists, and where we expect significant differences in microphysical processes and vertical hydrometeor distribution due to variations resolved ascent behaviour and the representation of the unresolved processes in the convection parameterisation (main focus in this study); and iii) the late stage, when most of the WCB trajectories have reached and remain in the UTLS as part of the outflow, allowing us to assess whether there are differences in hydrometeor and vapour content that remain even after the trajectories have left the highest-resolution domain.

Although our primary focus is on the ascent of WCB trajectories, we briefly point out interesting aspects of the descending motion of trajectories. In the nested simulation, Schwenk and Miltenberger (2024) found that most trajectories also experience periods of descent, a finding that is consistent with our analysis of the global simulation. However, the descending motion in the global simulation occurs at a different time and in a different location compared to the nested simulation. In the nested simulation, the strongest descents start from approximately 600 hPa and are spread out from 21.09.2017 and 23.09.2017 (Fig. A6 b). In contrast, the global simulation shows a bundled and concentrated period of descent starting at approximately 350 hPa and 12:00 on 22.09.2017 (Fig. A6 a). Some few

trajectories in the nested simulation also experience intense descent of $\Delta p_{2h} > 400$ hPa, whereas in the global simulation the maximum is 107 hPa. This indicates that not only the ascending but also the descending flow in the atmosphere changes when changing from a convection-parameterising to a convection-permitting resolution.

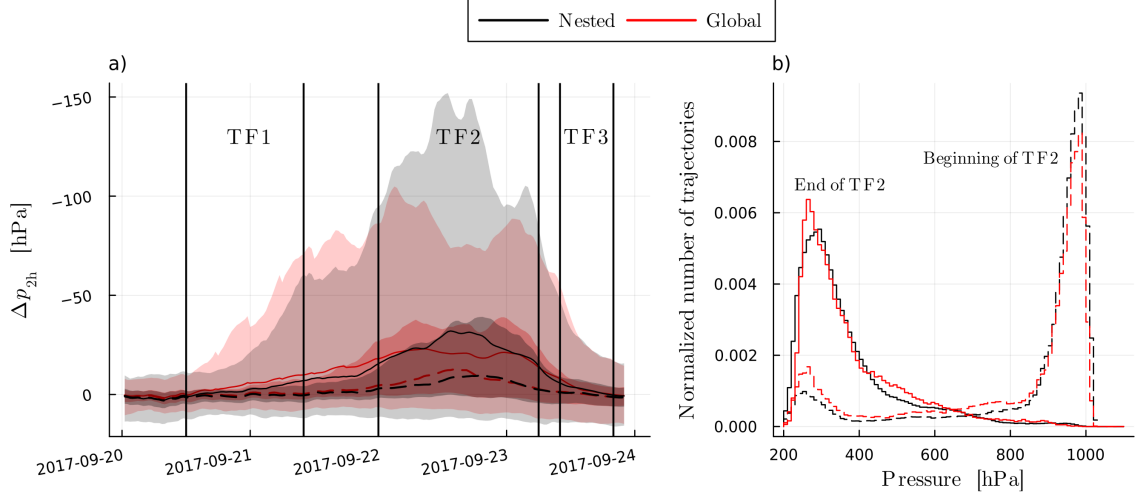


Figure 3. a) Δp_{2h} -values per trajectory over simulation run-time for the global (red) and the nested (black) simulation. Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation. Vertical lines indicate the three time-frames (TF) that exhibit different ascent-behavior. b) shows histograms for the pressure of trajectories in the nested and global simulation at the beginning (dashed histograms) and end (solid-line histograms) of TF2.

3.2 Conditions at the end of the ascent

In this section we summarise the properties of trajectories at the end of their WCB ascent (after τ_{WCB}) in the nested and the global simulation, and discuss the causes thereof.

Pressure, Temperature

In both simulations, trajectories end their ascent at pressures between 200 hPa and 400 hPa (Fig. 4 a), and at temperatures between -65°C and -20°C (Fig. 4 b). The pressures at the end of the ascent are very similar (Fig. 4 a), but at the end of the whole simulation, trajectories in the nested simulation are located on average at 319 hPa, compared to 328 hPa for the global simulation, indicating that they remain in higher regions of the atmosphere for longer. The average end-of-ascent temperature is -44.6°C in the nested and -44.0°C in the global simulation, and at the end of the simulation this differences grows to -41.4°C and -39.6°C (Fig. 4 b).

At the end of the simulation, trajectories in the nested simulation are on average 24 h old (age defined as hours passed since end of ascent), compared to 25 h for the global simulation, which could in part explain the difference in pressure and temperature at this time. Because of the small shifts in the temperature and pressure distributions we will often compare variables between simulations at similar temperatures and pressures to eliminate biases (besides allowing for a more physics oriented analysis). Finally, the geographical distribution of trajectories at the beginning and end of the ascent, as well as at the end of the simulation, is similar in both simulations. The only notable difference is that in the nested simulation, WCB trajectories are found slightly further north and south at the end of their ascent (Fig. A1).

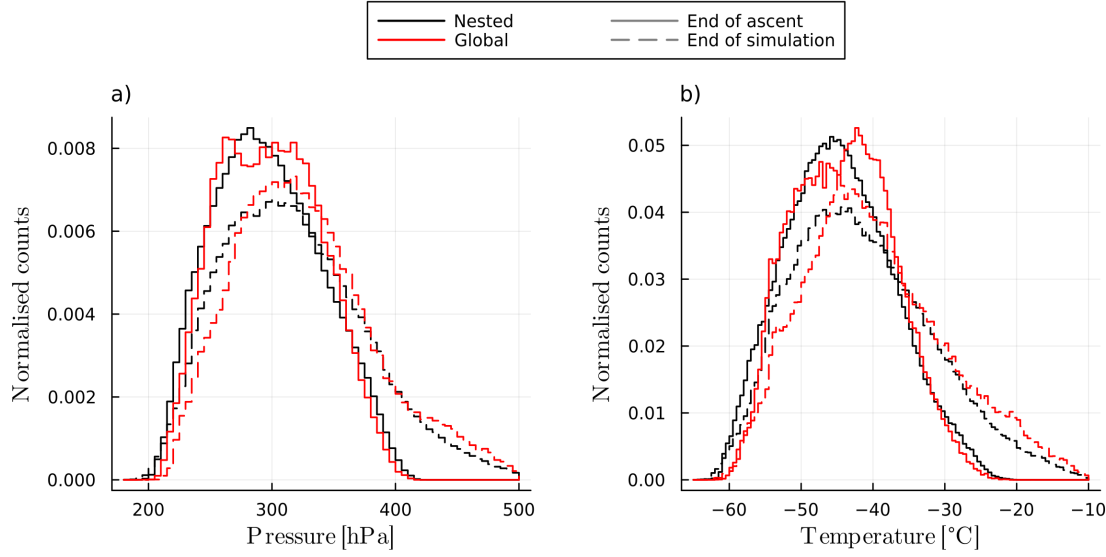


Figure 4. Normalised histograms of pressure (a) and temperature (b) at the end of the ascent (solid line histogram) as well as at the end of the simulations (dotted line histograms) for the nested (black) and global (red) simulations.

Vapour content and relative humidity over ice

The logarithm of the specific humidity ($\log_{10}(q_v)$) at the end of the ascent in the nested simulation follows a Gaussian distribution, whereas in the global simulation the distribution is skewed slightly towards larger values (Fig. 5 a). This results in a larger mean q_v value for the global simulation of 0.216 g/kg compared to 0.212 g/kg in the nested simulation. In both cases, q_v at the end of the ascent is primarily constrained by the temperature and pressure at the end of the ascent (Fig. A8), meaning that the higher q_v in the global simulation is primarily due to higher temperatures and lower pressures at the end of the ascent (Fig. 4 b). However, when looking at q_v values across pressure bins (Fig. 5 b), we find that between 225 hPa and 250 hPa, the mean, median and lower percentiles for q_v are consistently higher in the global simulation. This difference is also reflected when binning q_v by temperature (between -60°C and -50°C) but is harder to identify due to the much stronger dependence of q_v on T (and is therefore not shown).

We therefore also investigate the relative humidity over ice (RH_i) at the end of the ascent. Here we find that, on average, RH_i is slightly larger in the global simulation, with an average of 103.9 % compared to 103.4 % in the nested simulation (Fig. 5 c). This difference in mean values is pronounced when binned by pressure at approximately 250 hPa (Fig. 5 d). However, differences in the shapes of the RH_i distributions are much more striking than the differences in mean values. In the nested simulation, RH_i values have a sharp peak just below 100 % which is mostly absent in the global simulation (Fig. 5 c). In Schwenk and Miltenberger (2024), this peak was attributed to convectively ascending trajectories that have a higher condensation ratio and thus remove supersaturation more efficiently. Since convective trajectories are largely absent in the global simulation (Fig. 2), RH_i assumes a more Gaussian distribution, with only a faint hint of a second peak below 100 %. These results clearly show that, at the end of the ascent and at pressures below 300 hPa, the vertical vapour distribution between the simulations differs greatly, and hint at the fact that this difference might be due to the lack of convectively ascending trajectories in the global simulation.

In order to determine a physical mechanism why some trajectories in the global simulation are more or less humid than their counterparts at the same pressure in the nested simulation when they

finish their ascent, we adopt the same approach as Schwenk et al. (2025); Khvorostyanov (1995); Khvorostyanov and Sassen (1998); Khvorostyanov et al. (2001) and examine the relaxation timescale for supersaturation over ice ($\tau_{\text{sat,ice}}$). For an air parcel with a vertical velocity of zero it is given by:

$$\tau_{\text{sat,ice}} = \frac{1}{4\pi \cdot \text{CAP} \cdot N_i \cdot r_i \cdot D_v} \quad (\text{with } [N_i] = \text{m}^{-3}), \quad (1)$$

with D_v the water vapour diffusion coefficient (taken as a constant value of $3 \cdot 10^{-5} \text{ m}^2 \text{ s}^{-1}$) and CAP the capacitance of ice (0.5 for a spherical particle). This equation states that the time-scale for the removal of supersaturation is inversely proportional to the number concentration of ice (N_i) times the radius of ice (r_i), and is only valid when vertical velocities are small (below $\sim 0.1 \text{ m s}^{-1}$). The vertical velocities at the end of the ascent are $(0.006 \pm 0.080) \text{ m s}^{-1}$ in the nested, and $(0.010 \pm 0.032) \text{ m s}^{-1}$ in the global simulation. Therefore, we can safely use Equation 1 for our analysis when looking at the end of the ascent.

As in Schwenk et al. (2025), we find that $\tau_{\text{sat,ice}}$ plays a strong role in determining RH_i at the end of the ascent and that this dependence is the same in both the global and the nested simulation (Fig. A9). Because $\tau_{\text{sat,ice}}$ -values span many orders of magnitude, we look at the distribution of $\log_{10}(\tau_{\text{sat,ice}})$ -values at the end of the ascent and find that their shape is overall similar, but shifted to larger values for the global simulation (Fig. 5 e). The mean $\log_{10}(\tau_{\text{sat,ice}})$ in the global simulation is 3.56 ($\sim 7000 \text{ s}$ for $\tau_{\text{sat,ice}}$) compared to 3.44 ($\sim 5000 \text{ s}$ for $\tau_{\text{sat,ice}}$) in the nested simulation. When binned by pressure, $\log_{10}(\tau_{\text{sat,ice}})$ is larger in the global simulation from 225 hPa to 350 hPa corresponding to the pressure range where RH_i is also either larger or similar to the values in the nested simulation. Below 225 hPa, $\log_{10}(\tau_{\text{sat,ice}})$ in the global simulation is smaller, just as RH_i . Therefore, $\tau_{\text{sat,ice}}$ seems to be a suitable diagnostic to explain difference in RH_i at tend of the ascent and in particular suggest that difference in the hydrometeor population between both simulations are causing the variations in the RH_i distribution. Differences in the hydrometeor population between global and nested simulation are explored in the next paragraph.

Hydrometeor Content

In both simulations, almost all trajectories are fully glaciated by the end of the ascent, with the only remaining hydrometeor species being ice (not shown). However, the ice mass and number concentrations (q_i and N_i) at the end of the ascent are very different between the simulations. q_i is on average 0.025 g/kg for the nested simulation, which is almost twice as high as the average of 0.015 g/kg for the global simulation (Fig. 6 b). The distributions for $\log_{10}(q_i)$ assume a mostly Gaussian shape in both simulations, but are more narrow and shifted to smaller values for the global simulation (Fig. 6 c). N_i is on average 4 times higher in the nested simulation ($\sim 4 \cdot 10^5 \text{ 1/kg}$) than in the global ($\sim 10^5 \text{ 1/kg}$) (Fig. 6 a). Both $\log_{10}(N_i)$ -distributions have a tail towards large N_i and are similarly shaped, but again the distribution from the global simulation is shifted to smaller values. The distribution for r_i is shifted to larger values in the global simulation (Fig. 6 e), and r_i values are clearly larger on average at pressures below 300 hPa (Fig. 6 f). This indicates that in the global simulation, the overall ice content at the end of the ascent is smaller in mass and number, while ice crystal radii are larger.

When binned by pressure, q_i and N_i at the end of the ascent are smaller in the global simulation in the same pressure regions (250-225 hPa) where q_v and RH_i are larger in the nested simulation or roughly the same both simulations (Fig. 6 b and d). Below 225 hPa, this behaviour is inverted, with q_i and N_i being larger in the global simulation. This inversion suggests that, when changing from a convection-parameterising to a convection-permitting resolution, the amount and vertical distribu-

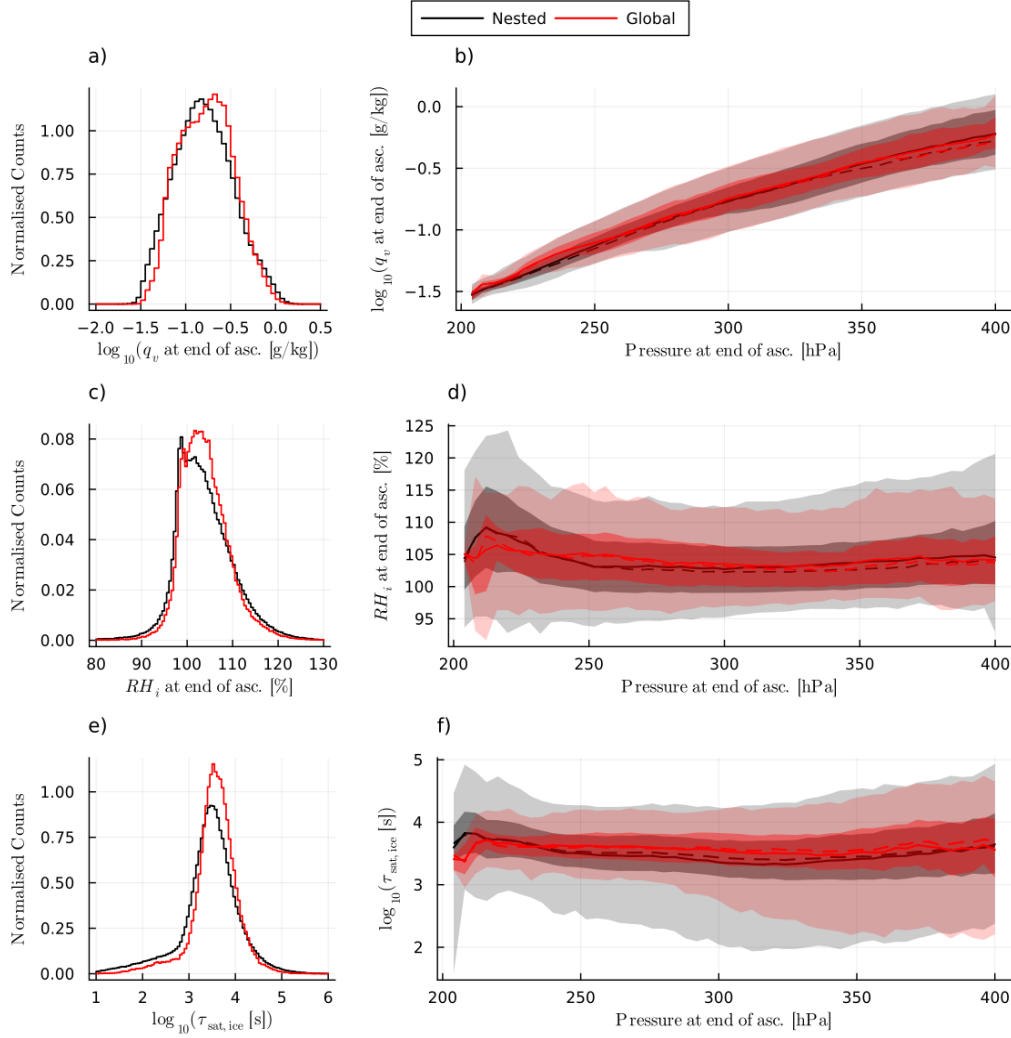


Figure 5. Normalised histograms of $\log_{10}(q_v)$ (a), RH_i (d) and $\log_{10}(\tau_{\text{sat,ice}})$ (f) at the end of the WCB ascent. The corresponding distributions binned by pressure are shown in b) and c) for $\log_{10}(q_v)$, in e) for RH_i and g) for $\log_{10}(\tau_{\text{sat,ice}})$. Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

tion of ice transported to the UTLS change. Above 300 hPa, the ice crystal radius (r_i) in the global simulation is similar to the nested simulation (Fig. 6 g), and below this pressure it is consistently larger in the global compared to the nested simulation (Fig. 6 f). Nevertheless, when averaged over all altitudes in the outflow, r_i is larger in the global simulation (~ 0.11 mm on average) than in the nested simulation (~ 0.10 mm on average), reflecting the higher N_i and q_i in the nested simulation. The distribution shapes of r_i are similar, with the slight difference that in the nested simulation, there is a hint of a secondary peak at very small radii (Fig. 6 e).

In the introduction, we hypothesized that the differences we see between simulations for variables such as ice crystal content and size at the end of the ascent will likely be due to larger vertical velocities in the nested simulation. We have already shown that overall, the ascent timescales are much smaller in the nested simulation (Fig. 2). Even for trajectories with similar τ_{600} , the $\max(\Delta p_{2h})$ -values are much larger than in the nested simulation (Fig. A7 a). The intense convective motion for $\tau_{600} < 12$, where $\max(\Delta p_{2h})$ increasing exponentially for decreasing τ_{600} , is missing entirely in the global simulation.

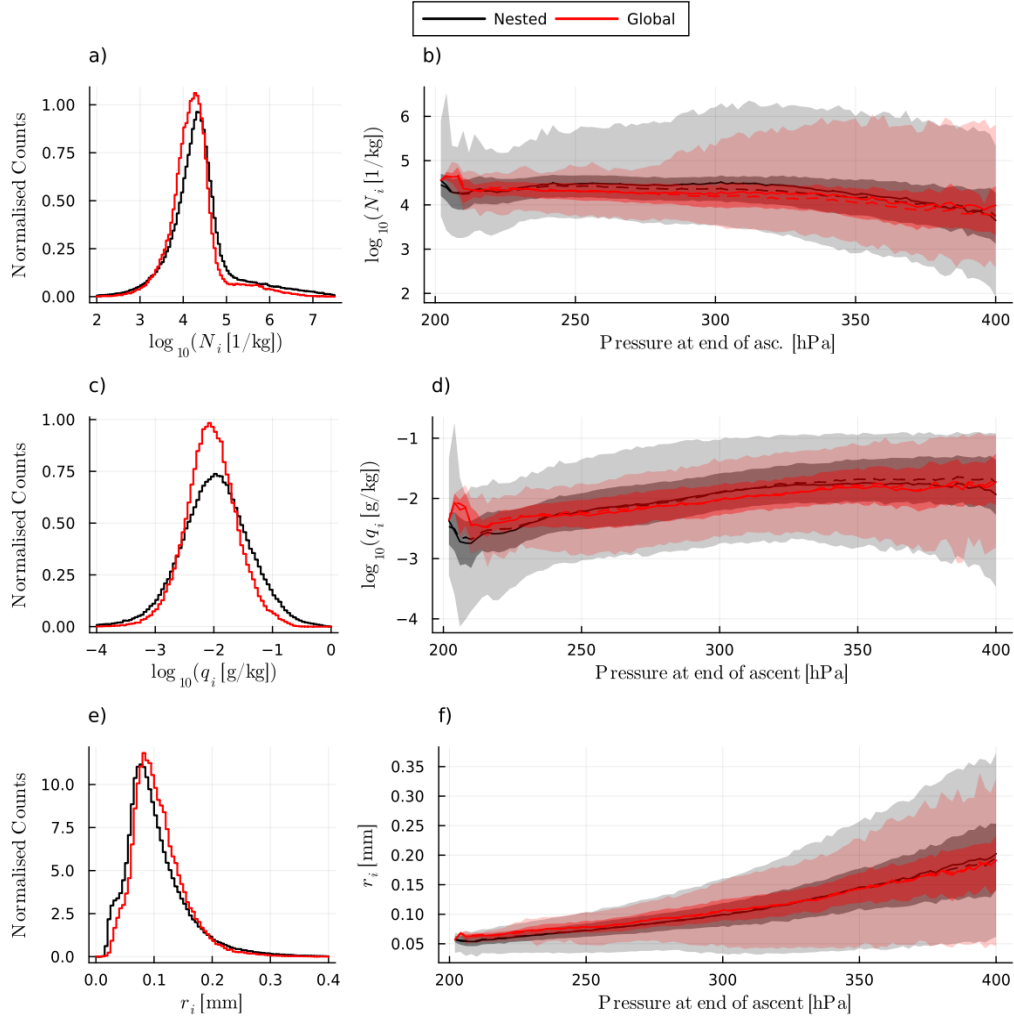


Figure 6. Normalised histograms of $\log_{10}(N_i)$ (a), $\log_{10}(q_i)$ (c) and r_i (e) at the end of the ascent. The distributions binned by pressure are shown in b) for $\log_{10}(N_i)$, in d) for $\log_{10}(q_i)$ and f) for r_i . Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

Schwenk and Miltenberger (2024) determined that high vertical velocities allow for more ice crystals to remain suspended in an air parcel, and that the less time spent in the ascent means that there is also less time for the growth of larger ice crystals. Therefore, the differences in ascent velocity and time between the two simulations can plausibly explain the higher N_i and q_i for pressures larger than 225 hPa in the nested simulation, as well as the overall smaller r_i . However, we also found that for pressures below 225 hPa, this behaviour is inverted, with N_i and q_i being larger in the global simulation (differences in r_i are unchanged). This inversion cannot be explained by differences in ascent velocity. Instead, we hypothesize that the convection parameterization, which can transport ice to high altitudes in a single time-step, might lead to the higher N_i and q_i values for the global simulation above 225 hPa. This hypothesis is tested in the second part of this study.

The differences in vapour and hydrometeor content at the end of ascent that we discuss in this section suggest that trajectories in the two simulations undergo distinct (micro)physical processes during their ascent. This will be discussed in section 3.3. These results also raise the question of whether the condensation ratio (CR) and precipitation efficiency (PE) are different in the nested and global simulations and is discussed next.

Precipitation Efficiency, Condensation Ratio

We use the definitions of the Lagrangian precipitation efficiency PE and condensation ratio CR from Schwenk and Miltenberger (2024) to quantify the water budget of WCB air parcels, the efficiency of vertical moisture transport and the role of cloud processes in modifying the latter. CR quantifies the fraction of vapour present at the beginning the WCB ascent that is converted into hydrometeors by the end of the ascent. It is therefore a measure of how efficiently vapour is removed. PE quantifies the fraction of hydrometeors formed during the ascent that are lost by the end of the ascent. It is therefore a measure of how efficiently hydrometeors are removed.

Based on the overall lower q_v for the nested simulation discussed in the previous section we expect that the CR should be larger overall than in the global simulation, since a lower q_v generally indicates that more vapour has been converted and likely dominates over differences in the boundary layer water vapour field. The distribution of CR confirms this (Fig. 7 a). However, if considering CR for trajectories arriving at different altitudes in the outflow, CR is also larger in pressure bins where q_v -values are the same or similar (Fig. 7 c), indicating an overall greater fraction of vapour to condensate conversion in the nested simulation. This difference likely derives from differences in boundary-layer moisture field and representation of boundary-layer moisture transport in the two simulations: Trajectories in the nested simulation are more humid and warmer at the beginning of τ_{WCB} (Fig. A2), and CR strongly depends on the initial q_v .

However, the higher overall q_v at the beginning of the ascent the nested simulation is most likely not due to a higher q_v in the planetary boundary layer (PBL) WCB inflow region of the nested simulation. Rather, the higher initial q_v is likely a result of the algorithm that determines the beginning of τ_{WCB} : The algorithm goes backwards from the beginning of τ_{600} until the ascent velocity is smaller than 8 hPa h^{-1} , and since trajectories in the nested simulation generally experience higher vertical velocities, this point can be shifted further backwards than in the global simulation. This higher overall ascent velocity in the nested simulation therefore shifts the beginning of τ_{WCB} closer to the surface, where q_v is higher. We conclude that this is a potential weakness and source of error when calculating CR. Perhaps a future study should explore ways to account for this induced bias.

In the nested simulation, the PE has a more broad distribution than in the global simulation and is smaller on average (Fig. 7 b). The bias induced for CR from the τ_{WCB} algorithm does not effect PE, because it depends more strongly on the initial hydrometeor concentration and not the vapour content. When binned by pressure, PE is smaller in the nested simulation from approximately 250 hPa to 350 hPa, and larger below 225 hPa (Fig. 7 d), which is mostly consistent with differences in N_i and q_i as expected (Fig. 6 b and d). This consistency confirms that the differences in hydrometeor transport, rather than the differences in vapour-to-hydrometeor conversion, drive the differences in hydrometeor content at the end of the ascent between the two simulations.

The larger spread in PE values for the nested simulation results in some trajectories precipitating nearly all hydrometeors, with 5% of trajectories in the nested simulation having $\text{PE} > 99.999\%$, as opposed to just 0.5% in the global simulation. The retention of hydrometeors strongly depends on vertical velocities as well as the hydrometeor size and type. Since Schwenk and Miltenberger (2024) showed that convective activity changes which hydrometeor species are more abundant in an ascending air parcel, we are able to attribute the different spread of PE distributions to the more convective flow in the nested simulation.

In this section, we analysed the conditions of WCB trajectories at the end of their ascent. We found that trajectories in the nested simulation end their ascent at slightly lower pressures and temperatures than in the global simulation (Fig. 4). The vapour content is smaller overall in the nested simulation (Fig. 5 a), but especially so below pressures of 250 hPa (Fig. 5 b). RH_i is also smaller in the nested simulation and the distribution peaks below 100 %, whereas in the global simulation the distribution peaks above 100 % (Fig. 5 c and d). We are able to trace these differences to different relaxation timescales $\tau_{\text{sat,ice}}$ between the two simulations. $\tau_{\text{sat,ice}}$ is inversely proportional to N_i (Eq. 1), which at the end of the ascent is much higher in the nested simulation (Fig. 6 a and b), which explains why $\tau_{\text{sat,ice}}$ is smaller and thus also RH_i . By analysing the moisture transport through the lens of the CR and PE, we are able to trace these differences in hydrometeor content (and therefore vapour) to differences in hydrometeor transport during the ascent (quantified by PE), as opposed to differences in the vapour-to-hydrometeor conversion (quantified by CR).

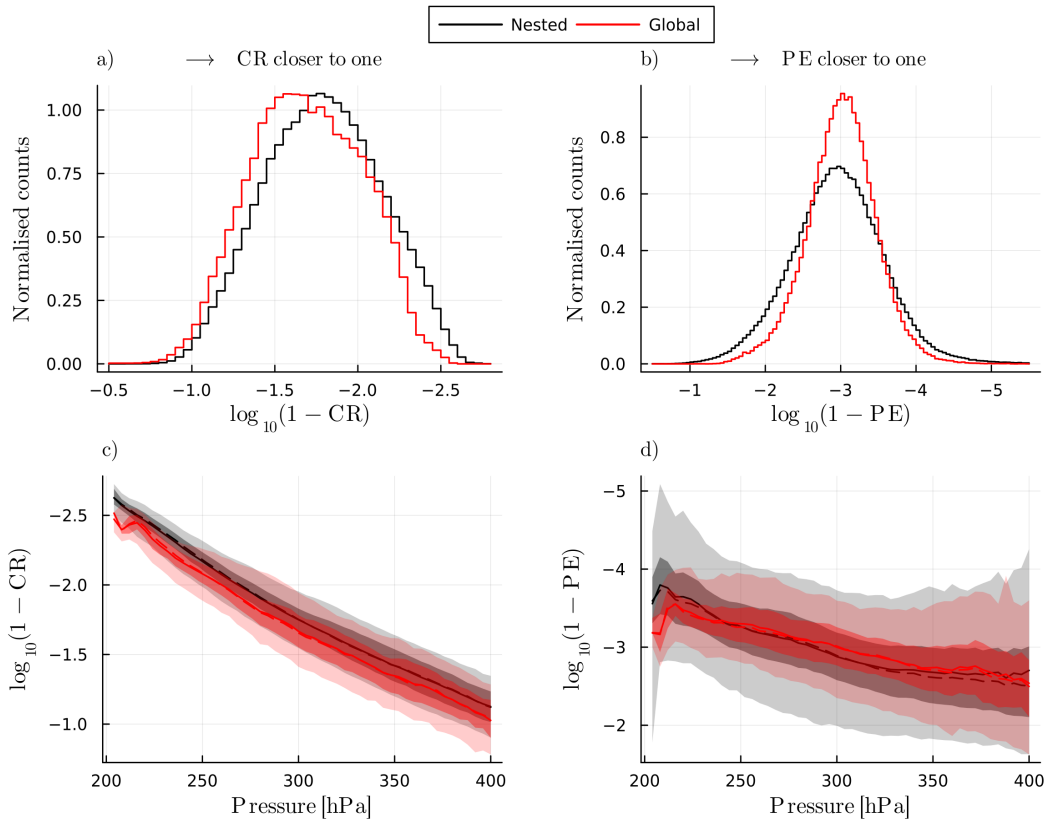


Figure 7. Normalised histograms of $\log_{10}(1 - \text{CR})$ in a) and of $\log_{10}(1 - \text{PE})$. The distributions binned by pressure are shown in c) and d), respectively, with mean (solid lines), median (dashed lines), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) for the nested (black) and global (red) simulation. We chose to display the parameters this way because it would otherwise be very difficult to visualize the difference in the CR and PE, which are both close to one and not spaced linearly. Larger negative values of $\log_{10}(1 - X)$ mean X is closer to 1. Therefore, values for CR and PE which are closer to one lie to the right in a) and b) (note the flipped x-axes), and to the top in c) and d) (note the flipped y-axes).

3.3 Microphysical composition and processes in the WCB ascent stage

In this section, we examine key differences and similarities in the microphysical processes and hydrometeor populations between the two simulations during TF2, the time period when the main WCB ascent occurs (Fig. A4). The aim is to develop a deeper understanding of what leads to the diagnosed differences in the outflow hydrometeor populations. In this analysis, it is important to note that the geographical distribution of trajectories is very similar between the simulations at each time step

(Fig. A1), allowing us to assume that any differences we observe arise from the simulated processes themselves rather than from geographical shifts between simulations.

Microphysical processes

Firstly, we examine integrated process rates that are accumulated by the end of the ascent. In a previous study, Schwenk and Miltenberger (2024) showed that convective WCB trajectories exhibit higher accumulated frozen precipitation flux — defined as the net loss of frozen hydrometeors along the trajectory ($q_{x,\text{in}} - q_{x,\text{out}}$), integrated over τ_{WCB} — while slow trajectories tend to produce more rain. We find that this pattern is similarly reflected in the global and nested simulations: Compared to the global simulation, the nested simulation shows a greater integrated frozen precipitation flux, a smaller integrated rain flux, and a higher total hydrometeor flux overall (Fig. 8,a). We also find higher integrated riming and condensation rates in the nested simulation, while the global simulation shows slightly higher integrated deposition rates (Fig. 8,b). These difference very likely arise from the missing representation of large (convection-related) vertical velocities in the global simulation and the impact of ascent timescale on the cloud microphysical pathways as identified by Schwenk and Miltenberger (2024).

Moisture loss via the convection parametrisation is nearly zero in the nested simulation (Fig. 8,b), since most of the ascent occurs within domains 2 and 3 (Fig. A4 a), where the (deep) convection scheme is switched off. This lack of the convection parametrization for most of the ascent in the nested simulation can explain the large differences observed in integrated condensation (cond) and turbulence (turbs) rates between the simulations (Fig. 8,b): Without the option to redistribute moisture through the convection parametrisation, the higher resolution model must instead condense and precipitate water vapour via the cloud microphysics parametrisation or lose moisture to the turbulence parameterization.

The greater moisture loss due to turbulence in the nested simulation also supports our interpretation that the atmospheric flow in the nested simulation has more fine-scale structure increasing also wind shear and thus the loss of moisture due to turbulence. In the global simulation, approximately 1.5 g kg^{-1} of moisture is lost due to the convection scheme, and moisture lost to the turbulence scheme is small. The clear signal from the convection parameterization demonstrates that the Lagrangian diagnostics are able to pick up on the contributions from the convection parameterization scheme, even though it does not influence the wind fields with which trajectories are advected. The negative sign also indicates that the contributions to the moisture budget mostly occur in the beginning of the ascent, where the scheme removes moisture from a parcel to redistribute it vertically, as opposed to at the end of the ascent, where the scheme takes moisture from the PBL and redistributes it into the parcel.

Binning the instantaneous (not integrated) microphysical process rates by pressure and temperature (Figs. A10 and A11) indicates that they differ primarily in magnitude rather than being shifted to different pressure or temperature regions. Notable exceptions are riming (Fig. A10 a, d) and the influx of graupel and rain (Fig. A11 a and b): in the global simulation, riming and the influx of graupel are largely absent below approximately 500 hPa and -10°C , whereas in the nested simulation, trajectories still undergo substantial riming and experience graupel influx at pressures around 400 hPa and temperatures as low as -20°C . This behaviour is due to the higher ascent velocities in the nested simulation, which allow for the production and retention of larger amounts of graupel and rain at lower pressures and temperatures.

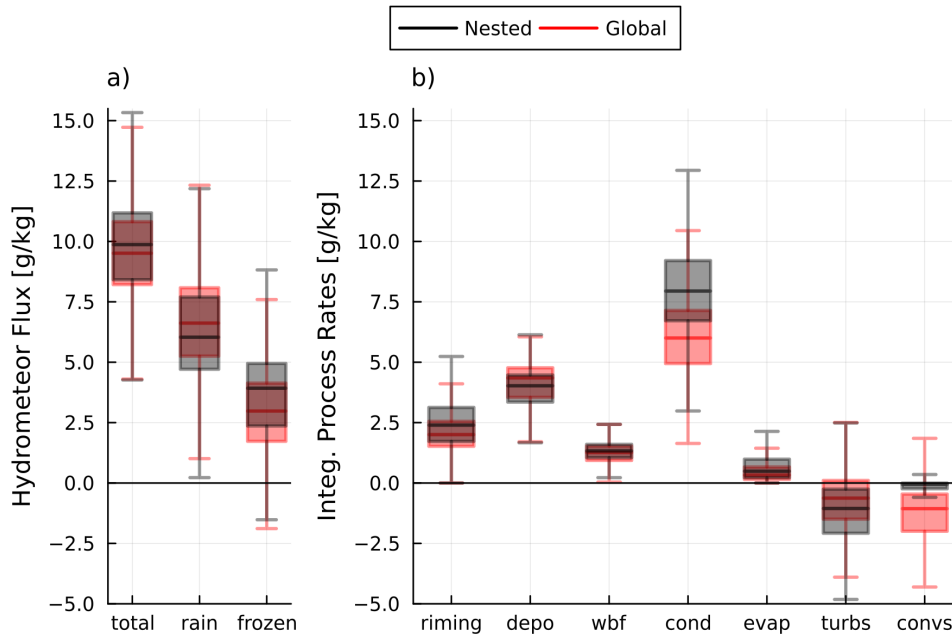


Figure 8. Box plots of a) the total hydrometeor flux, rain flux and frozen hydrometeor flux accumulated by the end of the ascent, as well as b) of the end-of-ascent accumulated microphysical process rates for riming, deposition, Wegener-Bergeron-Findeisen process, condensation, evaporation, turbulence parametrisation tendencies and convection parametrisation tendencies, respectively. All tendencies are shown separately for the nested (black) and global (red) simulation.

Higher vertical velocities keep particles in suspension longer which allows them to grow. This means that the nested simulation favours the production of graupel over ice and snow (supported by the larger integrated riming rate shown in Fig. 8 b). Trajectories in the global simulation consistently experience a slightly greater influx of ice and snow than those in the nested simulation (Fig. A11 c and d). We note that these differences only slightly change the pressure and temperature of glaciation (first time step where liquid water content is zero; Fig. A12). On average, air parcels in the nested simulation glaciate at 362 hPa and -32.2°C , compared to 356 hPa and -32.8°C in the global simulation. Even though we attribute most of the previously identified differences between the two simulations to the difference in vertical velocities, the process of reaching full glaciation is not strongly affected by this difference. This is probably because vertical velocities between the two simulations differ most strongly in the lower troposphere ($p < 500$ hPa; Fig. A7 b), whereas full glaciation occurs at around 360 hPa, where vertical velocities are much more similar (Fig. A12). However, vertical velocities do influence the hydrometeor content and type, meaning that some effect on the glaciation can be expected. Since the majority of trajectories in both simulations experience slow, slantwise ascent, it is possible that the differences induced by different vertical velocities at the glaciation point are minimal.

Hydrometeor content

During the ascent, i.e. TF2, the total hydrometeor and frozen hydrometeor mass mixing ratio are both higher in the nested simulation across all pressure bins (Fig. 9 a and b). The mass mixing ratio of liquid water (Fig. 9 c) is similar at pressures of approximately 600 hPa but larger in the nested simulation above and below this pressure. When examining the number concentrations for individual hydrometeor species (Fig. A13 and Fig. A14), we find that the largest difference in frozen hydrometeor content is seen for graupel, which has far higher mass mixing ratios and number concentrations in the nested

simulation (Fig. A13 c and f), especially below pressures of 600 hPa. The radii of ice and snow are smaller or similar in the nested simulation (Fig. A13 g and h), and the number concentrations mostly larger (Fig. A13 d and e). The mass mixing ratios are similar or slightly larger (Fig. A13 a and b). Especially for snow, the largest difference between simulations is seen for the size and number concentration, not the mass mixing ratio. Cloud droplet mass mixing ratios and number concentrations are higher in the nested simulation (Fig. A14 a and c), whereas radii are similar. For raindrops, the mass mixing ratios and radii are larger in the nested simulation (Fig. A14 b and f) but the number concentrations are far lower (Fig. A14 d).

These differences can mostly be explained by the different vertical velocities experienced by trajectories in the nested and global simulation. Higher vertical velocities mean that more hydrometeors can remain suspended in the ascending air parcel, which explains higher mass and number mixing ratios in the nested simulation. The larger number and mass mixing ratios of cloud droplets (Fig. A14 a and c) could also be due to differences in CCN activation. This process is also strongly dependent on vertical velocity, with higher vertical velocities leading to higher CCN activation rates and smaller cloud droplets. Therefore, higher vertical velocities also impact the size distribution of cloud droplets, and as a consequence thereof also influences the conversion rates of cloud droplets to raindrops, and the size of the ice particles that result from the freezing of cloud droplets. Therefore, differences in vertical velocities can also explain the different hydrometeor radii. The larger raindrops seen in the nested simulation are exempt from this explanation, and could instead be caused by the higher influx of graupel from aloft (Fig. A11 b) which is melted to form raindrops.

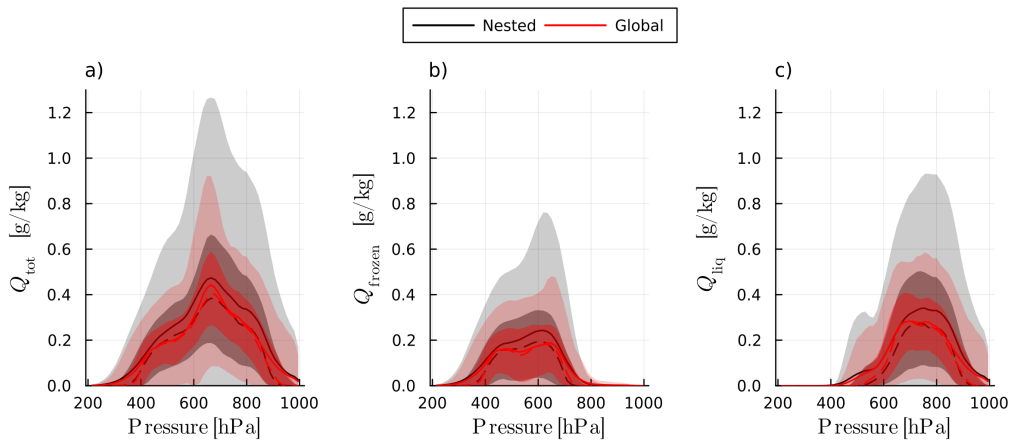


Figure 9. Total (a), frozen (b) and liquid (c) hydrometeor mass mixing ratios binned by pressure for TF2. Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

In summary, the trajectories in the two simulations undergo different microphysical processes during WCB ascent phase with arguably the most important difference being a shift of precipitation to the frozen phase in the nested simulation. During the ascent and for ice, snow, and cloud drops, the nested simulation tends to favour more and smaller hydrometeors, whereas for graupel and rain the opposite is the case. Differences in the vertical distribution of hydrometeors are minor for all hydrometeor species except for graupel, which is shifted substantially to lower pressures in the nested simulation. We can plausibly attribute all of these differences to the different ascent velocities experienced by trajectories in the two simulations.

3.4 Conditions after the end of ascent

In this section we investigate trajectories after they have completed their ascent, when they form the WCB outflow and remain mostly below pressures of above 500 hPa. This is the case in TF3, during which q_v and RH_i become overall very similar in the two simulations (Fig. 10 a, b). However, when binned by pressure, the mean and median q_v and RH_i are both higher in the global simulation than in the nested simulation below approximately 280 hPa (Fig. 10 a and b). In the case of q_v , this difference is relatively small in absolute values, but percentage-wise and at 250 hPa, q_v is 12% higher in the global simulation than in the nested simulation.

The vertical shift in ice-content we showed in Section 3.2 is exaggerated in TF3. q_i and N_i are larger in the nested simulation below approximately 350 hPa, and larger in the global simulation above this pressure (Fig. 10 c and d). This presumably due to the smaller r_i below 300 hPa at the end of the ascent in the nested simulation (Fig. 6 f), which can remain suspended at higher altitudes for longer. This indicates that the nested simulation also produces more high-level cirrus clouds. The higher N_i in the nested simulation at the lowest pressures could also enhance vapor-to-ice conversion efficiency and therefore explain the smaller RH_i . However, this difference is minor and the larger RH_i in the global simulation below approximately 280 hPa might instead (or also) be caused by the influx of moisture from the convection scheme. This might also explain why q_i in the global simulation is slightly larger at approximately 250 hPa and N_i increases in the same region.

These differences in vertical distribution of N_i and perhaps in q_v in the outflow of the WCB between the simulations raise an important follow-up question: do these differences translate into detectable changes in the radiative balance at the top of the atmosphere (TOA)? The larger N_i at lower pressures in the nested simulation could lead to lower cloud top temperatures, which could reduce OLR. However, the slightly lower q_v might trap less OLR, meaning these effects might cancel out, if they are present. This is investigated in Schwenk and Miltenberger (2025a).

4 Conclusions and Discussion

In this paper, we investigate and compare the moisture transport into the UTLS by a WCB that occurred on 23 September 2017 for two simulations with different grid spacings. Both simulations use the ICON model and air mass trajectory data from the ICON online trajectory module and run on a global grid with an approximate grid spacing of 13 km. One simulation additionally has a doubly-nested setup, with a convective-permitting resolution of approximately 3.5 km in the second nest which encompasses the main WCB ascent area. In this paper—which is part one of a three-part study—we analyse only the Lagrangian trajectory data. In Schwenk and Miltenberger (2025a) we analyse Eulerian data to verify the results from this study and additionally investigate differences in radiation. While previous studies such as Choudhary and Voigt (2022) have investigated how grid spacing influences WCB trajectories, our focus on the transport of moisture and hydrometeors into the UTLS is novel. Additionally, we use online trajectories which deliver far more accurate data than offline trajectories (Sprenger and Wernli, 2015; Miltenberger, 2014).

We expect that WCB trajectories in the higher-resolution ("nested") simulation ascend more rapidly and that the observed differences in moisture and hydrometeor transport between the simulations can largely be traced back to this difference in ascent dynamics. Choudhary and Voigt (2022) clearly found that higher resolutions lead to higher vertical velocities, and this is expected because in the lower-resolution ("global") simulation, convection is parameterised using the Tiedtke–Bechtold scheme,

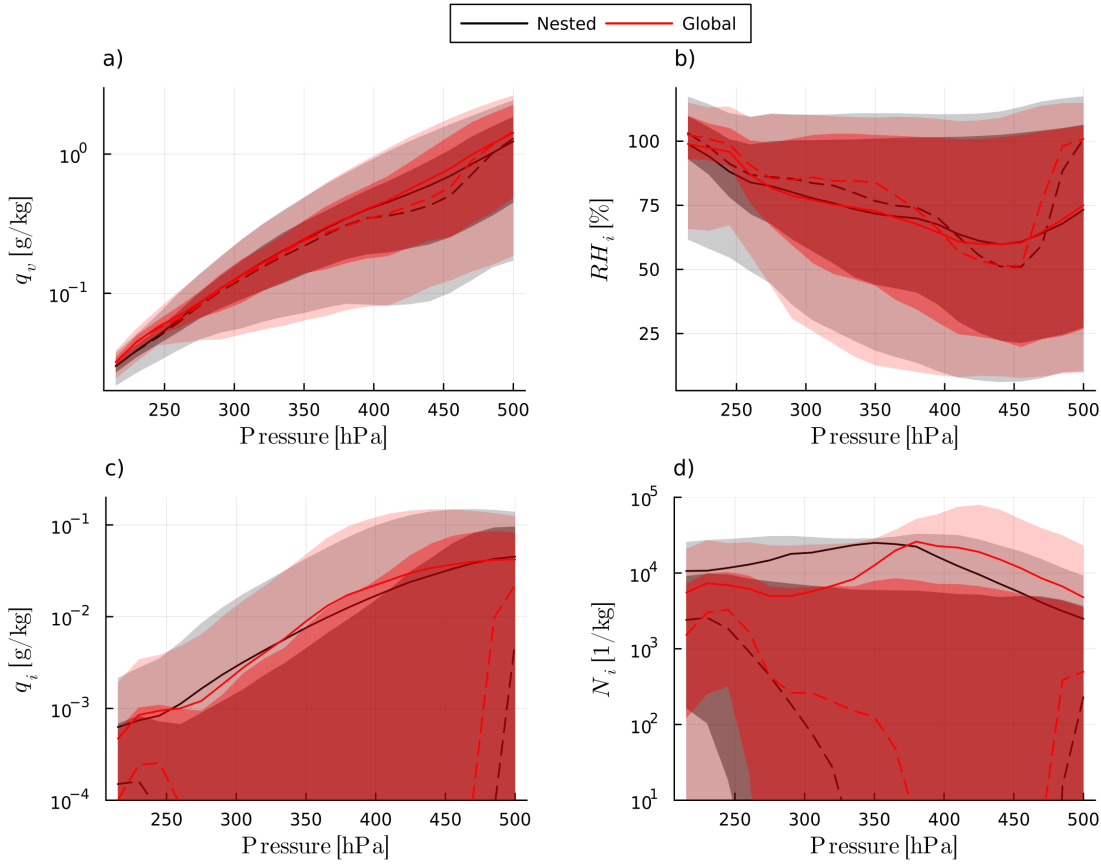


Figure 10. q_v (a), RH_i (b), q_i (c) and N_i (d) binned by pressure for TF3 (time after the ascent), with mean (solid lines), median (dashed lines), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) for the nested (black) and global (red) simulation.

which represents subgrid-scale convective mass fluxes but does not explicitly generate resolved vertical winds. By contrast, in the convection-permitting nested simulation, CAPE is depleted through explicitly resolved ascending motions, leading to stronger vertical velocities. We put forward the hypothesis that this difference in vertical velocities affects moisture and hydrometeor transport, because Schwenk and Miltenberger (2024) showed that trajectories undergoing rapid ascent differ substantially in these properties compared to those that ascend more slowly.

In line with our expectations, trajectories in the nested simulation ascend more rapidly. Not only are the τ_{600} and τ_{300} values on average much smaller (Fig. 2), but the $\max(\Delta p_{2h})$ values are also much higher even when τ_{600} are similar (Fig. A7 a), indicating that trajectories in the nested simulation experience more frequent and intense "bursts" of convective ascent than in the global simulation, regardless of ascent time. Trajectories in the nested simulation also end their ascent at slightly lower pressures and temperatures than those in the global simulation due to their more rapid ascent (Fig. 4), consistent with Choudhary and Voigt (2022). We additionally find that trajectories in the nested simulation tend to remain at lower pressures and temperatures for longer. This may occur because faster-ascending trajectories reach deeper into the UTLS, where the atmosphere could be stratified more strongly, preventing descent. These results show that the atmosphere (in the WCB region) of the nested simulation has stronger and more dynamic vertical motions compared to the global simulation. In this study, we are able to plausibly trace many of the differences observed between simulations to these differences in vertical velocities.

At the end of the ascent, trajectories in the nested simulation have slightly smaller q_v and RH_i than in the global simulation (Fig. 5 a-e), but this does not hold true across all pressure bins. q_v and RH_i are smaller in the nested simulation for pressures between approximately 225 hPa to 300 hPa, and higher below 225 hPa. This indicates that when changing from a convection parameterizing to a convection permitting simulation, the vertical humidity distribution in the WCB outflow changes. We are able explain these differences using the relaxation timescale for supersaturation over ice ($\tau_{\text{sat,ice}}$, Eq. 1), which is inversely proportional to N_i and r_i . N_i and q_i are higher in the nested simulation when RH_i is lower, and vice versa (Fig. 6 a-d), which makes differences thereof the likely explanation for the differences in RH_i and q_v . We attribute higher N_i and q_i in the nested simulation to higher ascent velocities that allow for the retention of more hydrometeors. The higher N_i and q_i in the global simulation, however, are not attributed to changes in vertical velocity but to the transport of hydrometeors by the convection scheme. r_i is generally smaller in the nested simulation (Fig. 6 e-f), but has less of an impact on $\tau_{\text{sat,ice}}$, because N_i spans many orders of magnitude and thus dominates the denominator in Eq. 1.

During the ascent (TF2), microphysical processes in the nested simulation shift to more frozen processes such as riming and trajectories also produce more frozen precipitation (Fig. 8, Fig. A10). Hydrometeor content and radii during the ascent also change; the nested simulation has higher graupel mass and number concentrations and raindrops are larger. For ice, snow, and cloud drops, the nested simulation tends to favor more and smaller hydrometeors (Fig. A13, Fig. A14). Assuming that differences in vertical velocity are the cause of these differences, these results align with those of Schwenk and Miltenberger (2024), suggesting that differences in vertical velocity are a plausible mechanism for these changes.

After the end of the ascent, when trajectories are in the outflow of the WCB (TF3), differences in q_v and RH_i are much smaller, but in the global simulation, q_v is slightly larger overall and RH_i is larger below approximately 280 hPa (Fig. 10 a and b). This indicates that the differences found at the end of the ascent can persist for some time and influence the WCB outflow. On the other hand, the differences in q_i and N_i in TF3 are much more pronounced than at the end of the ascent, with both q_i and N_i being much larger in the global simulations for pressures larger than approximately 350 hPa and much smaller below (Fig. 10 c and d). However, at approximately 250 hPa in the global simulation, q_i is slightly larger than in the nested simulation, and N_i increases slightly, which might be caused by the transport of hydrometeors by the convection parameterization. Whether these differences in UTLS humidity and hydrometeor content translate to differences in outgoing longwave radiation is discussed in Schwenk and Miltenberger (2025a).

The results of this study support our hypothesis that higher vertical velocities in convection-permitting simulations not only alter the ascent characteristics but also have a substantial effect on the transport of moisture and hydrometeors into the UTLS by a WCB, as well as the resulting cloud composition and microphysical processes during the ascent. These findings are important because they suggest that convection-parameterizing simulations, which remain widely used in weather forecasts and especially in climate models, may inaccurately represent WCB cloud properties and humidity in the outflow (if we assume that higher-resolution models represent reality more accurately). This has potentially important implications for how these models simulate Earth's radiative budget or the upper level flow.

There are some important limitations to consider when interpreting our results. First, the trajectory data are not guaranteed to be domain-filling, meaning that certain key regions within the WCB may

be under-represented or missing entirely. Second, as we have noted, the convection parameterization does not influence the wind-field in the global simulation, meaning that trajectories do not ascend faster even when they are in a region where convection is triggered. As a result, the convection parameterization may generate high N_i and low q_v values in the WCB outflow, while the trajectories themselves do not directly capture these effects. In Schwenk and Miltenberger (2025a), we therefore independently verify the findings presented here and demonstrate that the results remain consistent and the conclusions hold.

Finally, it is common within the modelling community to compare outputs from different simulations and draw conclusions without directly validating them against observational data. In this study, we highlight differences between the models and operate under the assumption that the higher-resolution simulation provides more accurate results; however, for this particular case, that assumption still needs to be verified. This is the aim of the third part of this study (Schwenk and Miltenberger, 2025b), where we develop and implement a method compare the simulation output presented here to satellite observations.

Code and data availability. Code and data will be made available after publication and will appear under the TPChange community on zenodo: <https://zenodo.org/communities/tpchange>. The code will also appear on my GitHub page <https://github.com/CornelisSchwenk/>.

Author contributions. CS and AM designed the experiment and conducted the numerical simulations. AM implemented the online diagnostics, and CS wrote the post-processing code. CS and AM worked jointly on the interpretation of the results. CS drafted the final manuscript with contributions from AM.

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Appendix A: Additional Figures

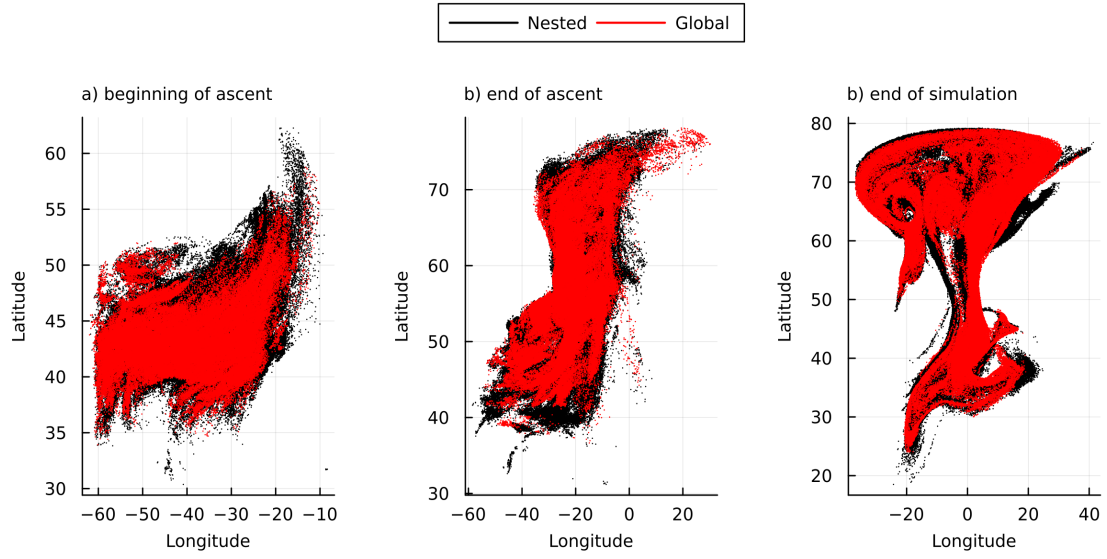


Figure A1. Longitude and latitude points of trajectories in the beginning (a), end (b) of ascent and at the end of the simulation (c) for the nested (black) and global (red) simulation.

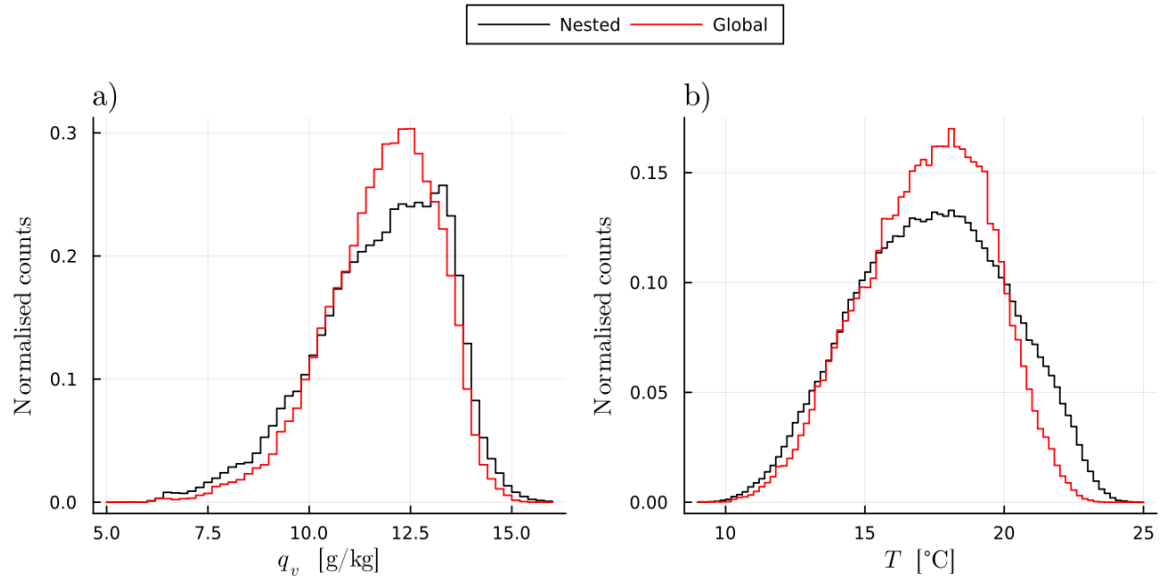


Figure A2. Normalized histograms of q_v (a) and temperature (b) at the beginning of τ_{WCB} for the nested (black) and global (red) simulations.

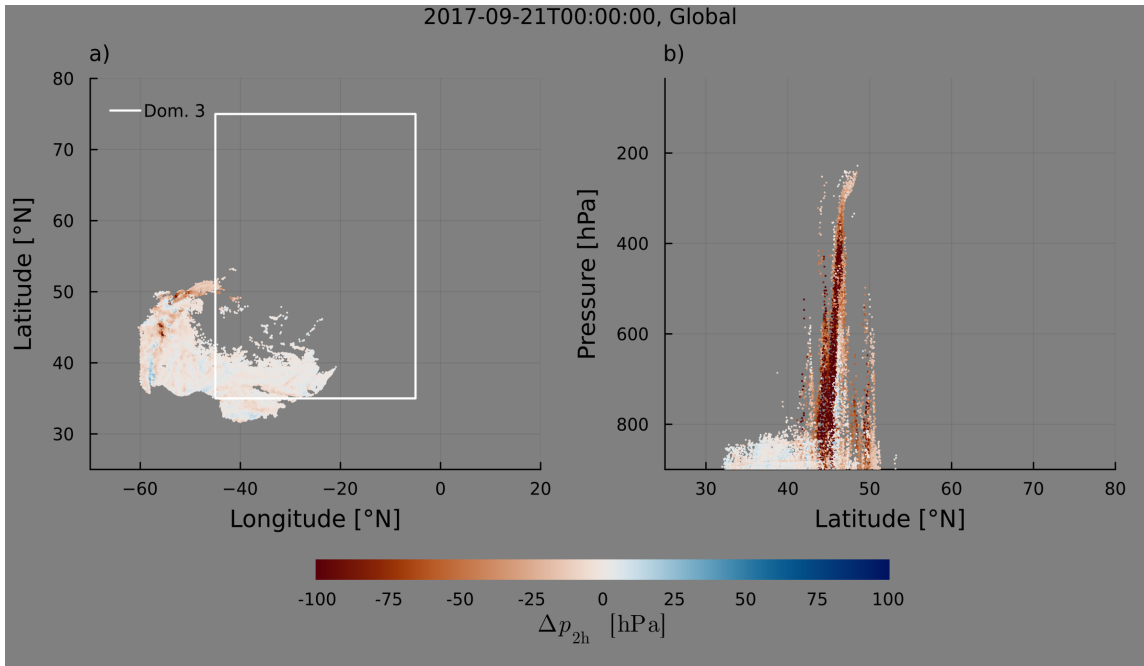


Figure A3. Trajectory longitude-latitude (a) and latitude-pressure (b) positions at the center of TF1 (00:00 on the 21.09.2017) with color indicating Δp_{2h} for the global simulation.

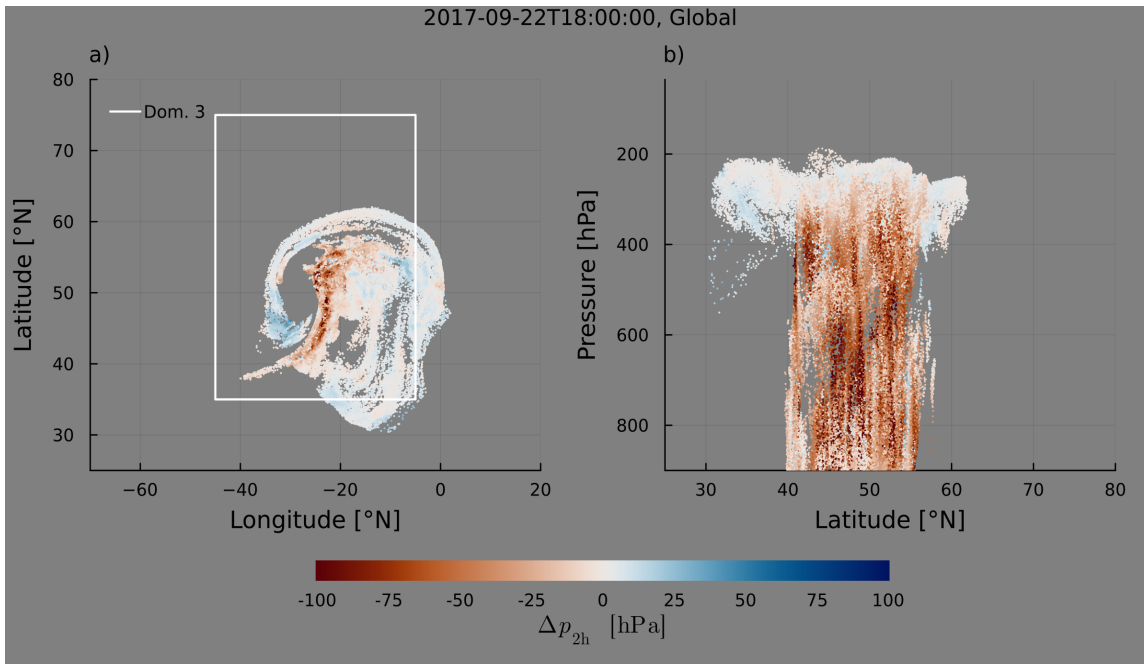


Figure A4. Trajectory longitude-latitude (a) and latitude-pressure (b) positions at the center of TF2 (18:00 on the 22.09.2017) with color indicating Δp_{2h} for the global simulation.

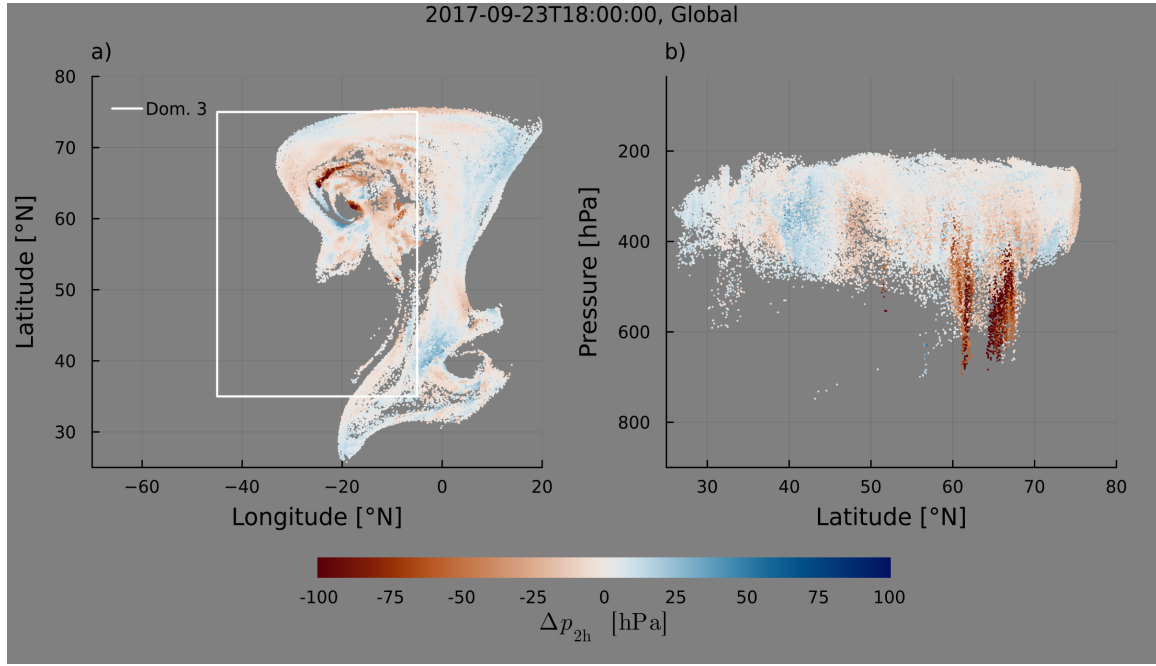


Figure A5. Trajectory longitude-latitude (a) and latitude-pressure (b) positions at the center of TF3 (18:00 on the 23.09.2017) with color indicating Δp_{2h} for the global simulation.

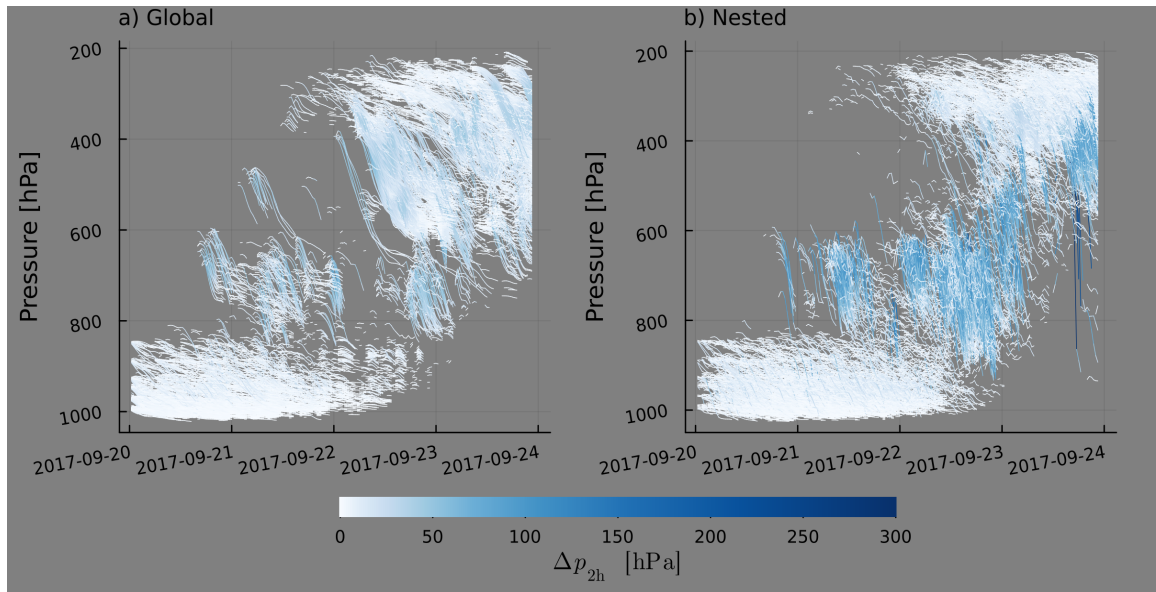


Figure A6. Plot of all positive Δp_{2h} -values (indicating descent) for each trajectory over simulation run-time for the global (a) and nested (b) simulation.

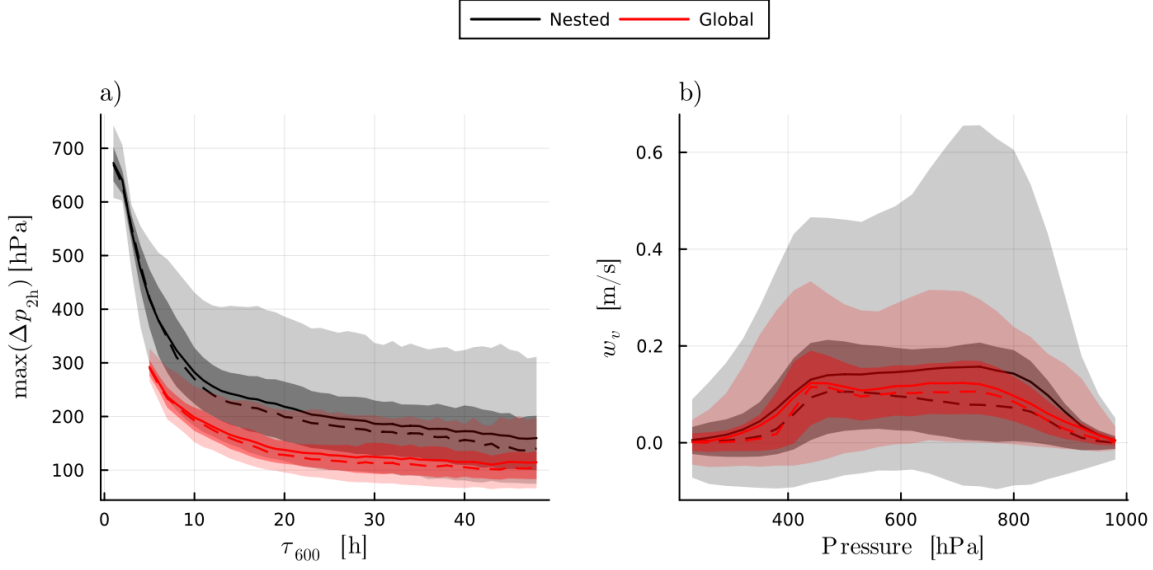


Figure A7. a) The maximum pressure velocity over 2 h ($\max(\Delta p_{2h})$) over τ_{600} , and b) the vertical velocities experienced by trajectories in TF2 binned by pressure, for the nested simulation in black and the global simulation in red, with mean (solid lines), median (dashed lines), inter-quartile ranges (light shaded) and 5th to 95th percentile ranges (dark shaded).

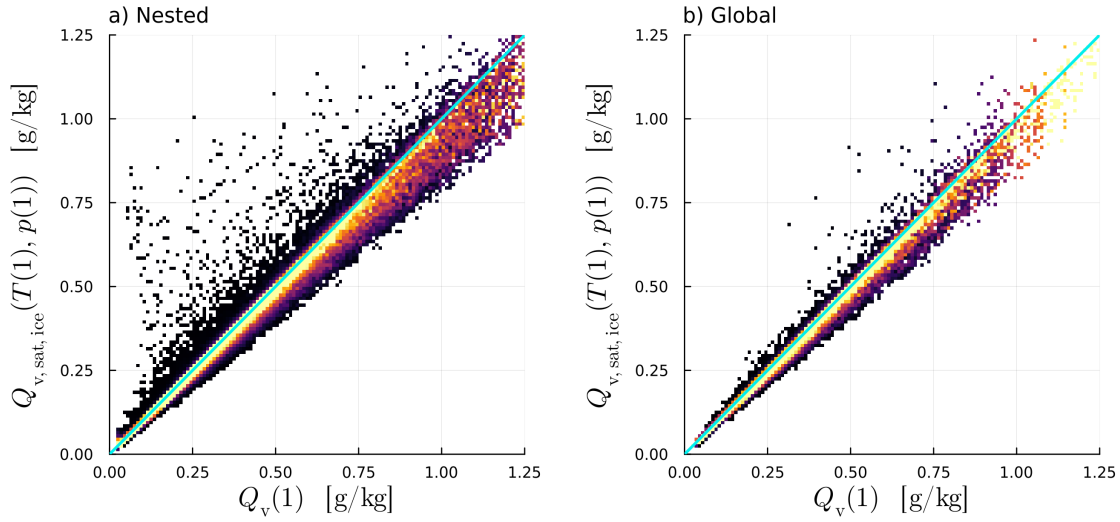


Figure A8. Two-dimensional histograms of calculated saturation specific humidity (over ice) given temperature and pressure at the end of the ascent (y axis), over the actual specific humidity (x axis) for the nested (a) and the global (b) simulation.

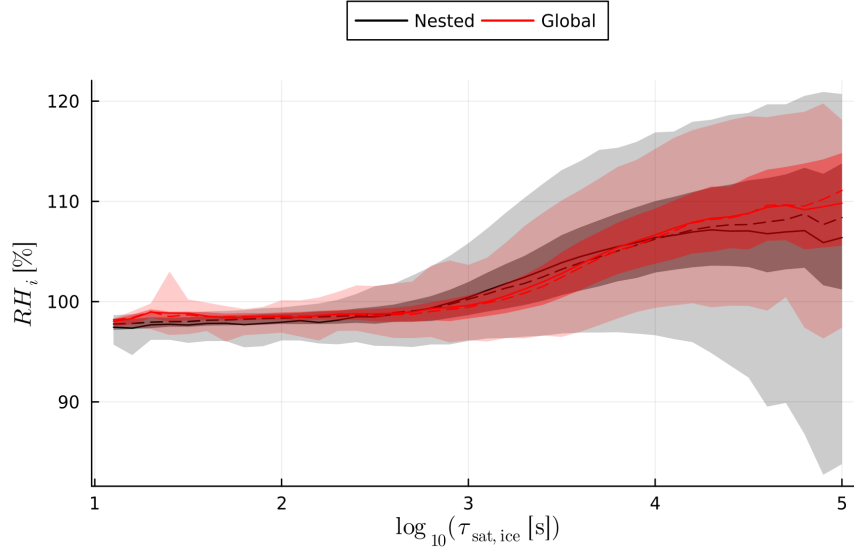


Figure A9. Behaviour of RH_i with $\log_{10}(\tau_{\text{sat, ice}})$ for the nested (black) and the global (red) simulation, with mean (solid line), median (dashed line), 5th-95th percentile ranges (light shaded) and inter-quartile ranges (dark shaded). Only trajectories with absolute vertical velocities smaller than 0.1 ms^{-1} are included for plotting.

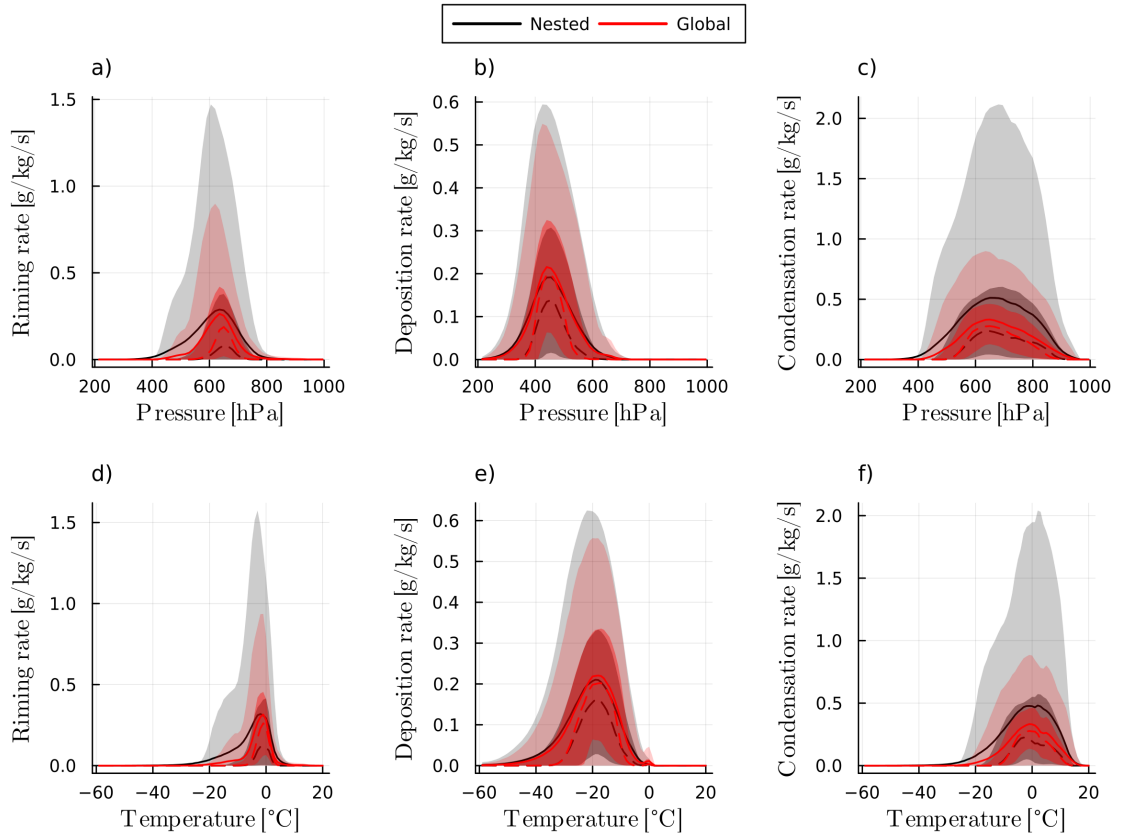


Figure A10. Instantaneous microphysical process rates during TF2 binned by pressure (top row) and temperature (bottom row) for riming (a and d), deposition (b and e) and condensation (c and f). Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

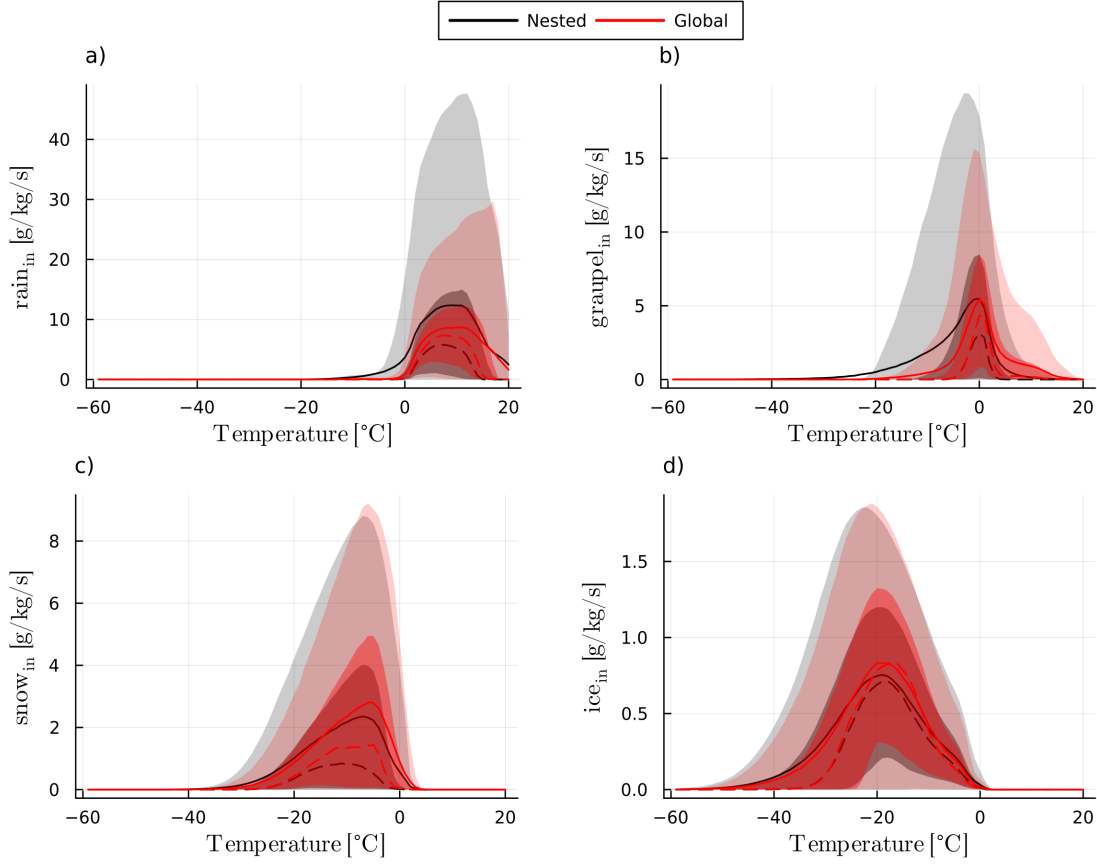


Figure A11. Instantaneous microphysical process rates during TF2 binned by temperature for the influx of rain (a), graupel (b), snow (c) and ice (d). Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

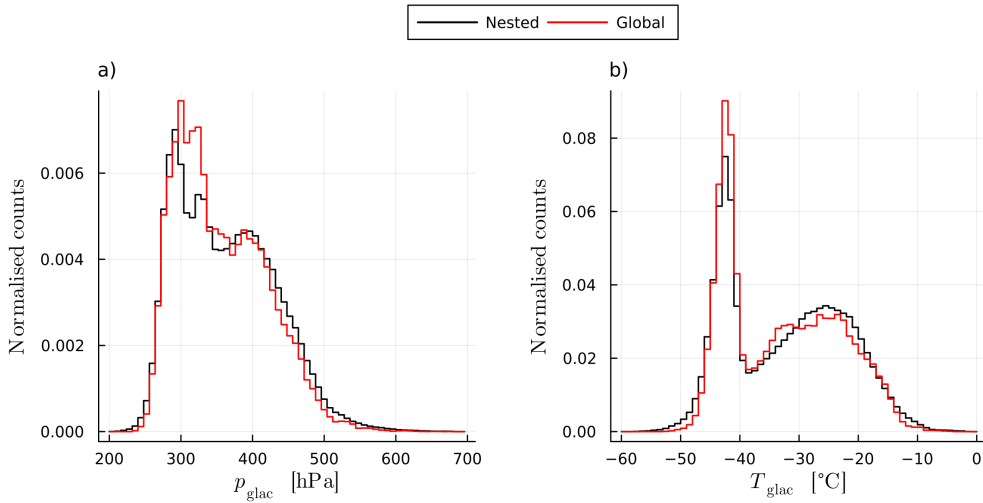


Figure A12. Normalized histograms of pressure (a) and temperature (b) of glaciation for the nested (black) and global (red) simulations.

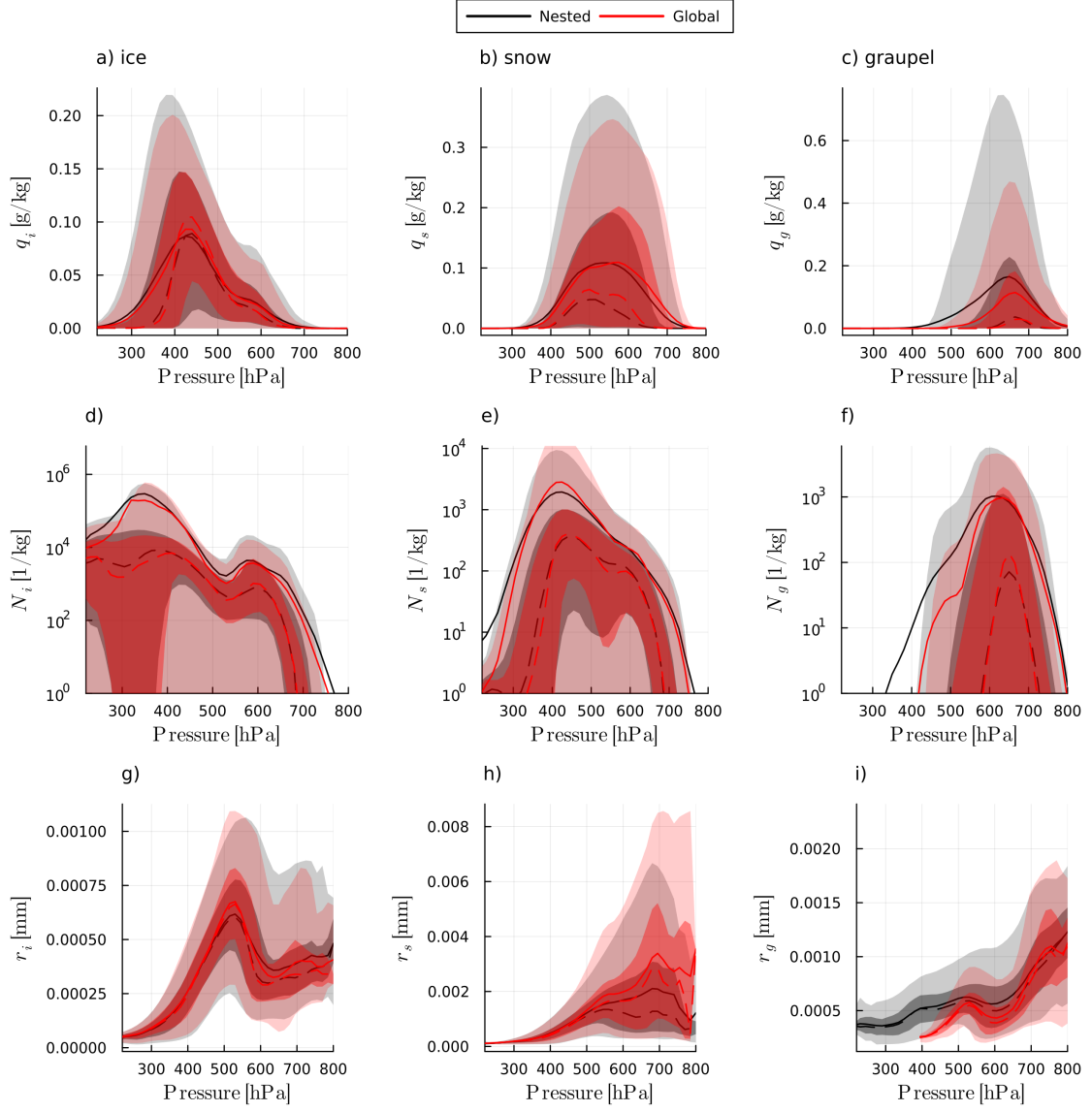


Figure A13. Hydrometeor mass mixing ratios (top row), number concentrations (middle row) and radius (bottom row) for ice (a, d, g), snow (b, e, h) and graupel (c, f, i) during TF2. Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

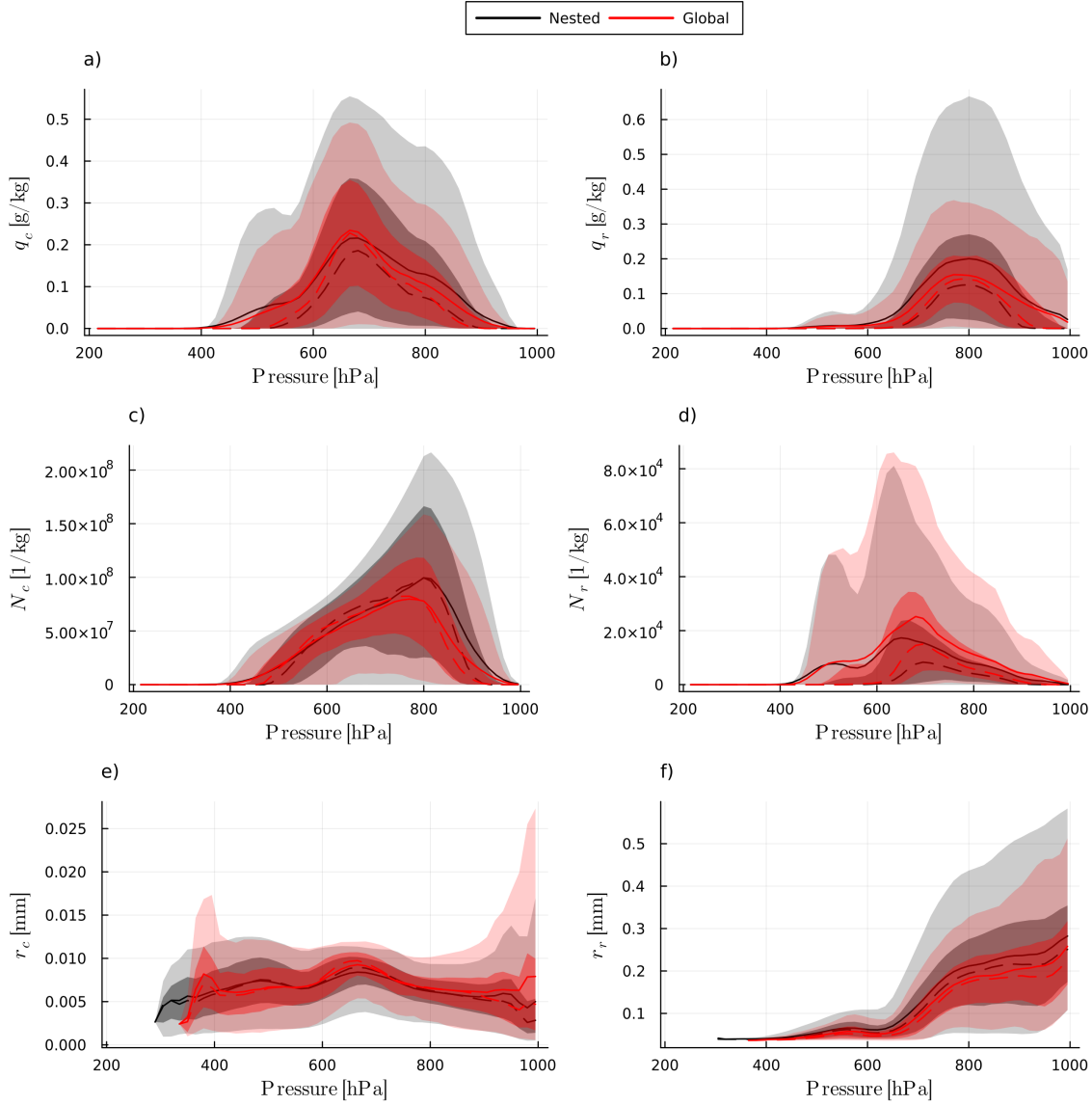


Figure A14. Hydrometeor mass mixing ratios (top row), number concentrations (middle row) and radius (bottom row) for cloud droplets (a, c, e) and rain during TF2. Mean (solid line), median (dashed line), inter-quartile ranges (dark shaded) and 5th to 95th percentile ranges (light shaded) are plotted for the nested (black) and global (red) simulation.

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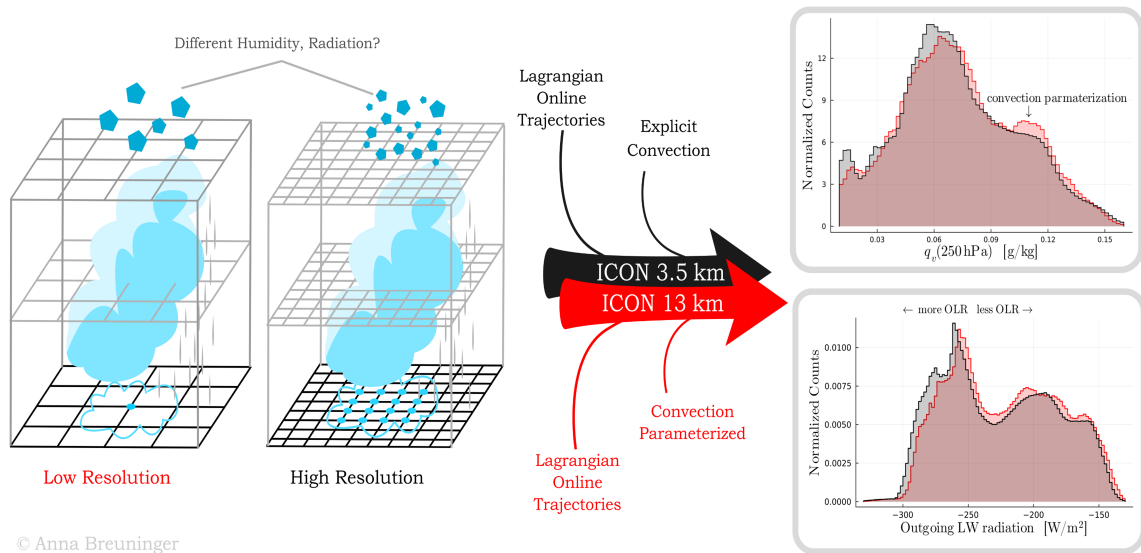
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Chapter 5

The Role of Scale - Part II: Eulerian Perspective

Research Question IIa

How do the modeled moisture transport, the governing microphysical processes, and the radiative effects of WCBs differ when transitioning from a convection-parameterizing simulation to a convection-permitting one?



This Chapter is in preparation to be submitted to *Atmospheric Chemistry & Physics* under the title "Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS)— Part II: Eulerian Perspective" (Schwenk and Miltenberger, 2025b). For author contributions, see Section 2.5.

Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS) — Part II: Eulerian Perspective

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Abstract. Warm conveyor belts (WCBs) are important components of extratropical cyclones, carrying water vapor and hydrometeors into the upper troposphere and lower stratosphere (UTLS), where they play a role in modifying Earth’s radiative budget. Earlier studies have demonstrated that the horizontal grid resolution of models significantly influences WCB properties such as ascent rates, diabatic heating, and the altitude at which ascent ends. In this three-part study, we investigate the role of grid resolution in the modeled transport of water vapor and hydrometeors into the UTLS by WCBs. Here, we conduct a Eulerian analysis to complement and improve on investigations done in the first part that used Lagrangian online trajectories. We analyze how model grid-spacing impacts microphysical processes and the transport of moisture and hydrometeors into the UTLS by a WCB. We examine two ICOSahedral Nonhydrostatic (ICON) model simulations of a WCB event over the North Atlantic on 23 September 2017—one using a convection parameterization at approximately ~ 13 km grid spacing, and another convection permitting simulation at around ~ 3.5 km grid spacing. We find that the convection parameterising simulation produces a more humid UTLS in the WCB outflow than the convection permitting simulation, which can be traced back to (i) smaller ice crystal number concentrations leading in turn to longer timescales for the depletion of supersaturation, but also (ii) to the influence of the convection parameterisation, which introduces too much vapor and too few ice crystals into the UTLS. Clouds in the convection permitting simulation are thicker and contain more graupel, while cloud top temperatures are similar in both simulations. Despite this, the convection permitting simulation produces more outgoing long-wave radiation at the top of the atmosphere, which we demonstrate is due to altered UTLS water vapor content and not to altered cloud properties. These findings suggest that convection parameterising simulations might misrepresent the transport of hydrometeors and humidity into the UTLS by WCBs, which may impact the role of WCBs in global climate models concerning their radiative effect and possibly their influence on upper-level flow.

1 Introduction

In the upper troposphere and lower stratosphere (UTLS), variations in water vapor concentration drive one of the most important positive climate feedbacks (Li et al., 2024; Held and Soden, 2000; Dessler et al., 2013, e.g.), making water vapor a key greenhouse gas in this region (Schneider et al., 2010). Despite its critical role, the UTLS remains one of the least well-characterized parts of the atmosphere (Jeffery et al., 2022), and the processes governing its moisture content — including the mechanisms of moisture transport — are still subject to considerable uncertainty (Petzold et al., 2020; Jeffery et al., 2022; Zahn et al., 2014; Guo and Miltenberger, 2025). The most prominent source of global UTLS moisture is deep convection in the tropics (Lee et al., 2019; Ueyama et al., 2023; Dauhut et al.,

2018; Hassim and Lane, 2010; Corti et al., 2008; Gordon et al., 2024). However, warm conveyor belts (WCBs) have also been identified as a major source of moisture in the extratropical UTLS by experimental (Zahn et al., 2014) and climatological studies ((Guo and Miltenberger, 2025)).

WCBs are an intrinsic part of extratropical cyclones (ETCs), where large-scale ascending air streams produce the typical elongated and comma-shaped cloud-bands associated with them. WCBs are important for mid-latitude weather patterns and the large amounts of precipitation formed by WCBs can be hazardous (Pfahl et al., 2014; Eckhardt et al., 2004). During the ascent of planetary boundary layer (PBL) air to the UTLS, WCBs produce mixed-phase clouds and precipitation and ultimately inject water vapor and hydrometeors into the UTLS (Madonna et al., 2014; Schwenk and Miltenberger, 2024). Their clouds also interact with radiation, making WCBs important when modeling Earth's radiative budget (Joos, 2019). For these reasons, it is important that numerical weather prediction (NWP) and global climate models (GCMs) accurately represent WCBs.

However, the correct representation of WCBs and in particular the moisture transport is difficult because WCB airstreams are a highly dynamic, multi-phase phenomenon that undergoes complex microphysical processes at each stage of their life (Binder et al., 2020; Forbes and Clark, 2003; Gehring et al., 2020; Hieronymus et al., 2025). Sensitivity experiments have shown that the choice of microphysical parameterization schemes significantly affects WCB ascent characteristics (Mazoyer et al., 2021, 2023), while the specific microphysical parameters within these schemes strongly influence diabatic processes during ascent (Hieronymus et al., 2022; Neuhauser et al., 2023; Forbes and Clark, 2003). Oertel et al. (2025); Schwenk et al. (2025) further demonstrated that modifying microphysical parameters in NWP models — such as the capacitance of ice and snow — even within plausible ranges, can substantially alter diabatic heating and ice content in the WCB outflow.

Collectively, these studies highlight the complexity of microphysical processes in WCBs and the major challenges they pose for accurate modeling. This is an important issue because the incorrect representation of WCBs has been linked to degraded forecast performance in the extratropics (Rodwell et al., 2018; Grams et al., 2018; Pickl et al., 2023; Berman and Torn, 2019). Therefore, the combination of processes (and the associated model uncertainty) that typically occur within WCBs make them a region where model uncertainties rapidly grow and amplify forecast errors (van Lier-Walqui et al., 2012; Morrison et al., 2020; Posselt and Vukicevic, 2010). This issue also increases the difficulty of quantifying the past, current and future radiative impact of WCBs (Joos, 2019).

Another critical factor, which has recently attracted attention, is the role of embedded convection within WCBs (Oertel et al., 2019, 2020; Schwenk and Miltenberger, 2024). Oertel et al. (2020) showed that convective regions within WCBs generate stronger surface precipitation, more intense diabatic heating, and have distinct impacts on the potential vorticity structure of the outflow. Schwenk and Miltenberger (2024) additionally showed that convective parcels transport significantly more ice into the UTLS and undergo distinct microphysical processes compared to slower-ascending parcels. Furthermore, Hieronymus et al. (2025) and Schwenk et al. (2025) showed that regions of embedded convection have different parameter sensitivities compared to more slantwise ascending regions. The insight from these previous high-resolution, convection-resolving studies raises an important question: Can low-resolution, convection-parameterizing simulations (which are typically used for global forecasting and climate predictions) adequately represent WCBs, particularly regarding their moisture transport into the UTLS?

Several studies have shown that increasing model resolution from convection-parameterizing to con-

vection-permitting scales can improve NWP model performance with respect to simulated top-of-atmosphere cloud-radiative effects (Senf et al., 2020) as well as precipitation patterns and the diurnal cycle (Vergara-Temprado et al., 2020). Choudhary and Voigt (2022) found that for simulations of WCBs, finer grid spacing resulted in more WCB air parcels ascending faster and reaching higher altitudes, alongside increased diabatic heating at all pressure levels, primarily from cloud-phase changes. This suggests that microphysical processes during WCB ascent are strongly resolution-dependent, likely tied to the role of convective air parcels (Schwenk and Miltenberger, 2024).

In this three-part study, we investigate the role of grid resolution in the modeled transport of water vapor and hydrometeors into the UTLS by WCBs. In Part-I (Schwenk and Miltenberger, 2025a) we used Lagrangian online trajectories to examine how model grid spacing affects the microphysical processes and the transport of moisture and hydrometeors into the upper troposphere and lower stratosphere (UTLS) by a warm conveyor belt (WCB). Specifically, we analyzed two simulations of a WCB event over the North Atlantic on 23 September 2017, conducted with the ICOSahedral Non-hydrostatic (ICON) model: one convection-parameterizing (~ 13 km horizontal grid spacing) and one convection-permitting (~ 3.5 km grid spacing). We hypothesized, and our analysis strongly suggests, that the higher prevalence of larger vertical velocities in the high-resolution simulation is the driving force behind many of the differences between the two simulations.

Our results from Schwenk and Miltenberger (2025a) indicate that the WCB outflow in the high-resolution simulation is drier. We attributed this to shorter relaxation timescales for supersaturation over ice driven by higher ice number concentrations. We also found differences in the microphysical processes experienced during ascent and in the fraction of frozen precipitation, which we plausibly linked to the stronger vertical motions in the high-resolution run. We provided a detailed insight into the microphysical processes and moisture conversion during ascent, because Schwenk and Miltenberger (2024) determined that these processes are sensitive to vertical velocities.

We also noted some important limitations of the Lagrangian analysis: the trajectory dataset is not domain-filling, meaning parts of the WCB may be underrepresented, and in the low-resolution simulation, the vertical motion of trajectories is not influenced by the convection parameterization scheme. In particular, the latter means we are only able to see the impact of convective moisture transport if a WCB trajectory passes over a grid-point where the convection scheme is triggered but not its impact on UTLS parcels not undergoing WCB ascent. Therefore, the aim of this second paper is expand our the analysis from Schwenk and Miltenberger (2025a) with an Eulerian analysis of the simulation data focusing less on the microphysical pathways in the WCB ascent but more strongly on the radiatively important UTLS moisture and hydrometeor content in the WCB outflow region. The second aim is to address the radiative implications: the drier WCB outflow in the high-resolution simulation could lead to more outgoing longwave radiation (OLR) at the top of the atmosphere (TOA), while the higher ice number concentrations could have the opposite effect. Our central research question for this paper is therefore: How does increasing model resolution from convection-parameterizing to convection-permitting affect the transport of moisture and hydrometeors into the UTLS, and how does it influence OLR at TOA?

The paper is structured as follows: First, we describe the methodology of the paper (section 2), which differs from Schwenk and Miltenberger (2025a) in that we perform an Eulerian analysis as opposed to a Lagrangian analysis. We then briefly verify the results of Schwenk and Miltenberger (2025a) from the Eulerian analysis (where possible; sections 3 and 4) and then analyze the differences in OLR at TOA and the causes of these differences (section 5).

2 Methods

This study is the second part of a three-part study. In Schwenk and Miltenberger (2025a) and in this paper, we examine two simulations of one WCB case study, run at convection-permitting and convection-parameterising resolutions. Furthermore, the case study, model setup (and simulation output) for the convection-permitting simulation are the same as in Schwenk and Miltenberger (2024), to which readers are directed to for details on the WCB case study and online trajectory analysis. Here, we summarize the key features of the case and setup in the following section. The subsequent sections describe the procedure of the Eulerian analysis, in particular our approach to identify WCB outflow airmasses from a Eulerian perspective.

2.1 WCB Case and ICON Model Setup

The WCB investigated in this study occurred over the North Atlantic on 23 September 2017, and we use the Icosahedral Nonhydrostatic (ICON) model (v2.6.2; Zängl et al. (2014)) for our simulations (see also Schwenk and Miltenberger, 2024). Both simulations were initialized with the operational ICON global analysis at 00:00 UTC on 20 September 2017, when the extratropical cyclone began to form off the coast of Canada, and ran for 96 hours until 00:00 UTC on 24 September 2017, by which time the WCB began to dissipate over Northern Europe.

Two simulations of this case at different resolutions have been conducted. The first simulation uses a R03B07 global grid which results in an approximate effective grid spacing of 13 km. This simulation is called the "global" simulation. The second simulation includes two higher-resolution domains that are nested into the global domain, and is therefore called the "nested" simulation. The nested regions were chosen such that they cover the main region of WCB ascent. The two nested domains use a R03B08 grid (~ 6.5 km grid spacing) and a R03B09 grid (~ 3.3 km grid spacing). Neighboring grids interact through two-way coupling (Zängl et al., 2022). The resolution of the highest-resolution domain was chosen to permit (deep and embedded) convection explicitly, while shallow convection remains parameterized. In the global domain deep convection is parameterized using the Tiedtke–Bechtold scheme (Tiedtke, 1989; Bechtold et al., 2008). Sub-grid processes such as turbulence, orographic drag, and radiation, are parameterized using the standard ICON schemes (in all domains). For the parameterization of clouds microphysical processes, we use the two-moment scheme by Seifert and Beheng (2005). This scheme represents six hydrometeor species (cloud droplets, rain, ice, snow, graupel, and hail). In contrast to the nested simulation, the lower-resolution ("global") simulation only has one domain on the same R03B07 grid with an approximate grid spacing of 13 km. In this simulation, deep convection is parameterized by the Tiedtke–Bechtold scheme everywhere.

Although this study focuses on Eulerian simulation output, we use trajectory data as the basis to produce our Eulerian WCB masks (see Section 2.2). The trajectory data was produced by the online-trajectory module, Miltenberger et al. (2020); Oertel et al. (2023). In both simulations, trajectories were initiated at 3-hour intervals throughout the 96-hour simulation and output recorded every 30 minutes. The starting regions was selected to represent the WCB air stream as well as possible (Fig. 1) and starting positions were selected randomly from grid points within this region at six vertical levels spanning from approximately 1000 to 800 hPa. Selection of actual WCB trajectories from the full trajectory data-set was performed as described in Schwenk and Miltenberger (2025a) and in Schwenk and Miltenberger (2024): Most importantly a WCB parcel has to ascend 600 hPa in no more than 48 h and be within a visually defined region at a specific time.

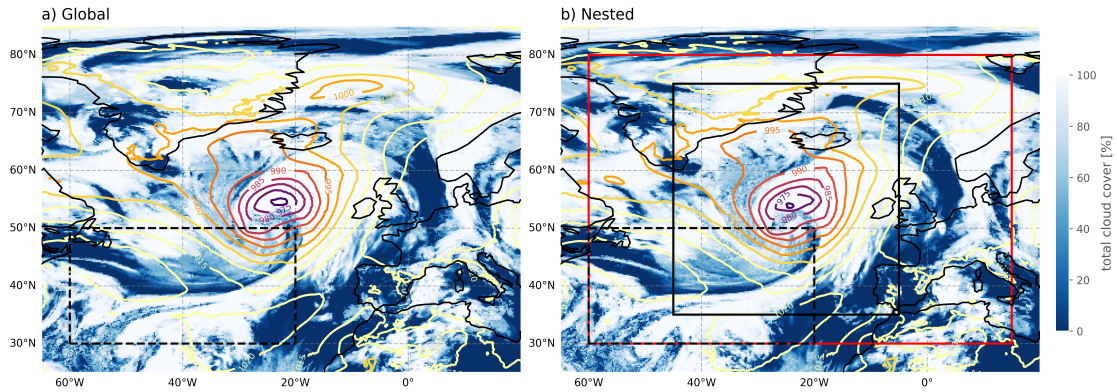


Figure 1. Simulation setup for the global simulation in a) and the nested simulation in b). Online trajectory starting area from Schwenk and Miltenberger (2025a) is marked by the dashed lines; the nested domain 2 (red) and 3 (black) boundaries are indicated in b) by the solid lines. Data shows the ICON output for the total cloud cover and sea-level pressure on 23.09.2017 00:00 UTC.

2.2 Eulerian WCB mask

A direct, point-by-point comparison of the Eulerian output from the two simulations is neither feasible nor meaningful, since the simulated fields inevitably shift in time and space relative to each other. Therefore, we created an algorithm that can isolate the WCB region in the Eulerian data for each simulation, using the WCB trajectory data as a basis. The goal is to produce masks that encompass as much of the WCB as possible and exclude non-WCB airmasses, and to then compare the simulations using a range of statistical measures, such as pressure-bin distributions and their mean and median. We only produce masks for the area encompassed by domain 3 (Fig. 1 b), because this is where the differences between simulations arise most strongly due to the differences in grid-spacing.

Our algorithm is split into two parts: First, we use the WCB trajectories to produce an initial two-dimensional (time-dependent) mask at each model output pressure level according to the following steps (schematic shown in Fig. 2):

1. At every pressure level, WCB trajectories that are located at pressures within ± 25 hPa of that level are selected (Fig. 2 a).
2. The number of identified trajectories at a given pressure level is represented on a Eulerian longitude-latitude grid (grid spacing 0.5° for both simulations; illustrated in Fig. 2 b). We restrict this to domain 3 for both simulations.
3. The gridded product is normalized and smoothed using a Gaussian filter with $\sigma = 1$ (Fig. 2 c).
4. The smoothed gridded product is converted to a binary mask, where only grid points with a normalized trajectory count larger than 0.005 are retained as WCB grid points (Fig. 2 d).
5. Finally, the WCB mask is re-gridded back to the original ICON grids for selecting the WCB related grid points in the native output of all simulations.

The resulting mask is very 'patchy' and leaves out large regions of the WCB where there are few or no trajectory points (Fig. 2 d). This reflects the limitation of the Lagrangian analysis presented in (Schwenk and Miltenberger, 2025a) related to the non-domain filling properties of the Lagrangian data. Therefore, we add a crucial second step in creating the final Eulerian WCB outflow mask:

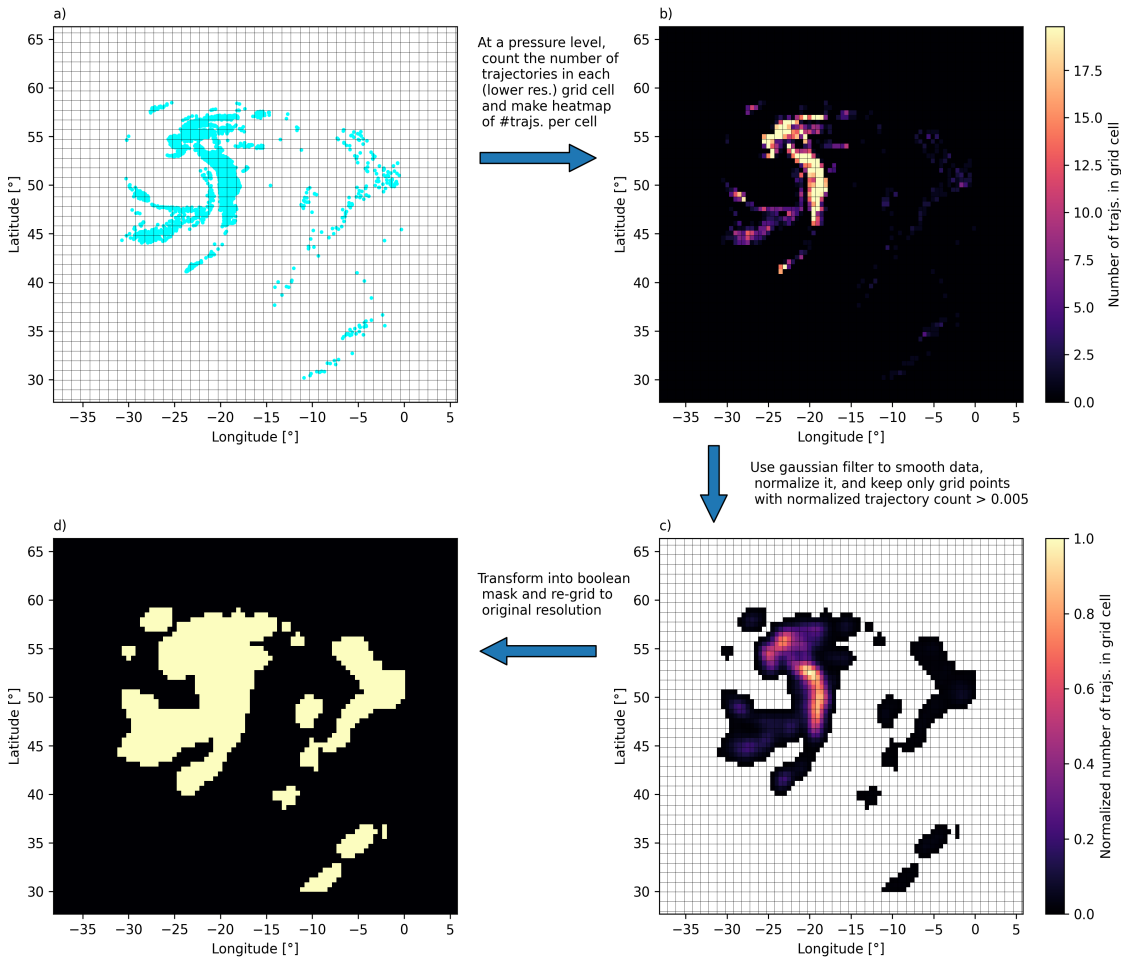


Figure 2. Schematic for the first step of creating the Eulerian WCB masks using the WCB trajectory data: a) at each time step and for each model level, select trajectories close to model level, b) count the number of trajectories per longitude-latitude bin (bin width of 0.5°), c) use Gaussian filter to smooth the data, which is then normalized, and d) create binary mask using a threshold of normalized trajectory count and re-grid to original grid size. This figure serves as an illustration, therefore the grid boxes etc. are not to scale.

1. Small, isolated regions of the mask that are more than 350 (100) grid points away from the largest centroid of the mask for the nested (global) simulation are removed (see for example the bottom left region in Fig. 3 a). The aim is to remove as many outliers, i.e. regions not related to the main WCB outflow, as possible for the next step (even if they are again included in the mask at a later step).
2. The mask is repeatedly dilated (ca. 60 times in the global simulation and 300 times in the nested simulation; optimal number of iterations determined through trial and error). After each iteration, all grid points for which the specific humidity (q_v) is smaller than a custom threshold value for that pressure level are discarded. This allows the mask to slowly 'crawl' forward and fill in the WCB area without reaching into regions outside of the WCB, where q_v is much smaller (Fig. 3 b).
3. At the halfway point of iterations, the mask is smoothed using a Gaussian filter. This enables the mask to 'jump' across filaments of low q_v within the WCB area instead of having to go around them and removes small isolated regions of the mask that are surrounded by low q_v and have therefore not dilated.

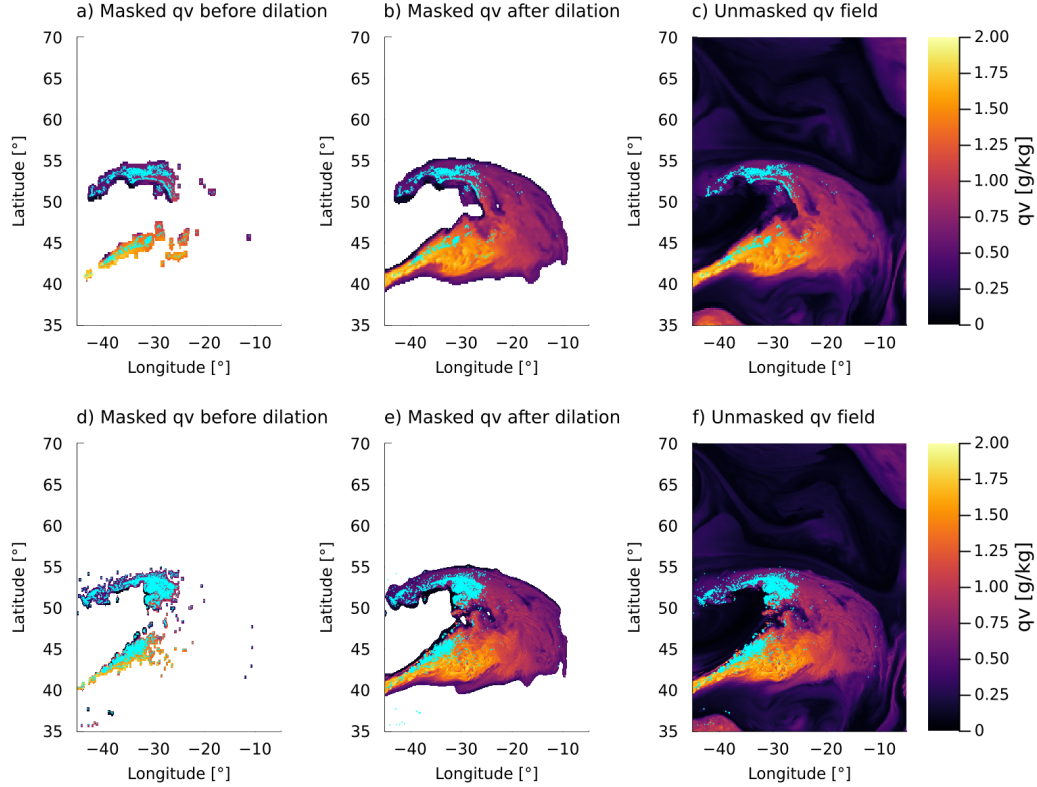


Figure 3. q_v at 400 hPa for the global simulation (top row) and nested simulation (bottom row) at 02:00 UTC on the 22. September, 2017. The area within the WCB mask is shown before (a, d) and after (b, e) the dilation. The total q_v field in (c, f). Trajectory points within ± 25 hPa of this pressure level are plotted in cyan.

There is a marked improvement in coverage of the WCB area by the dilated mask (Fig. 3 b and e) compared to the initial mask (Fig. 3 a and d). The masks that we produced for both simulations during TF2 and TF3 are visually very similar in terms of area (when collapsed) and volume (Fig. 3 b and e; Fig. A2 b and Fig. A3 b; otherwise not shown). The Area and volume of the masks are also similar in magnitude, but consistently larger in the nested simulation by about 5%, yet no more than 7% (not shown). This means that we should not expect substantial biases to arise from the WCB masks having different geographical coverage. Hence, the outlined algorithm can produce WCB masks that cover the entire WCB influenced area (and eliminate outliers) starting from non-domain filling trajectory data (and include outliers). In theory, one trajectory per pressure level could suffice to fill the WCB area at that pressure level through dilation. Therefore, the advantage of the masking procedure is that relatively sparse trajectory data can be used to produce the Eulerian WCB masks. Another advantage is that the resulting mask 'wraps around' the WCB in three dimensions, meaning it can exclude grid points that are below or above the WCB but not directly part of it (Fig. A1 b). To obtain all longitude-latitude points that across which the WCB airstream passes at any altitude, the mask can be "collapsed" onto two dimensions (Fig. A1 a). In this way, two-dimensional variables such as total cloud cover or column-integrated rain mass can be retrieved.

Despite the overall benefits of the outlined algorithm to identify WCB influenced parts of the atmosphere, there are also some limitations. Although this method goes beyond trajectory data, it still relies on it. If there are too few trajectories at a given pressure level, regions of the WCB could still be excluded, even with dilation. The dilation procedure can also cause the mask to extend into unrelated weather systems that are geographically close to the WCB and also have a high q_v . This was particularly problematic for pressure levels close to the PBL, where q_v is generally high and the WCB

'touched' other mesoscale systems. The associated 'muddy' q_v field makes it difficult to prevent the masking procedure from dilating too far. For this reason, we only ran the second part of the masking algorithm for pressure levels below 600 hPa. For higher pressures, we only use the WCB mask after the first step (i.e. only based on Lagrangian data). However, as we are mostly interested in WCB outflow properties in our analysis, the impact of this on our analysis is expected to be small.

The outlined masking algorithm further relies on several human choices that make its automatic translation to other case-studies not straightforward. For instance, the precise implementation for this study included numerous custom thresholds, primarily determined 'by eye' (see the published code for specifics). Modifications were made to the q_v field (to prevent the mask from dilating into a certain region), the number of dilation iterations, and the final mask itself for certain pressure levels and time steps. This was an intensive process involving a lot of manual double-checking and is feasible for the case study presented here. However, for implementation in a climatological study, the algorithm would need to be adapted to ensure that the mask never dilates into unrelated weather systems because manually double-checking every time step would be too time-consuming. Nevertheless, the general approach we have conceptualized here (using trajectories and dilation according to level-specific q_v thresholds) could form the basis of an effective algorithm for this purpose. The benefits of this method compared to already established WCB detection algorithms such as from Quinting and Grams (2022), is that this process is inherently grid-scale agnostic and highly adaptable to be as inclusive or exclusive as desired (by tuning the respective thresholds). It also does not rely on pre-training a machine learning algorithm on a specific combination of variables, but instead only requires q_v .

2.3 Definition of time frames for analysis based on WCB ascent stages

In Schwenk and Miltenberger (2025a) we defined three main time frames (TF) during which the WCB trajectories experience different parts of their ascent. During TF2, which is defined as beginning on 00:00 UTC 2017-09-22 and lasting to 06:00 UTC 2017-09-23, most trajectories begin and end their WCB ascent. Most of this ascent also occurs within the boundaries of domain 3 for both simulation. We will therefore focus on TF2 to investigate the WCB during the main ascent-phase. In TF3, defined as beginning on 00:00 UTC 2017-09-23 and ending at 20:00 UTC 2017-09-23, most trajectories remain above 500 hPa and are part of the WCB outflow. However, in TF3 many trajectories leave the boundaries of domain 3 in both simulations. In spite of this, we restrict ourselves to domain 3 for the analysis of the outflow in TF3, because this is where the grid-spacings between simulations are the most different.

3 Differences in WCB properties during ascent-dominated stage (TF2)

3.1 Vertical velocity

WCB trajectories in the nested simulation experience more intense vertical motion overall, during both ascending and descending segments of the overall WCB ascent (Schwenk and Miltenberger, 2025a). This suggests that the nested simulation has more chaotic and intense vertical motions. The Eulerian results clearly confirm this: For example, at 00:00 UTC 2017-09-23 and at 600 hPa, the masked vertical velocity (w) field in the global simulation shows large regions of coherently ascending vertical motion (Fig. 4 a), whereas in the nested simulation the same field is more patchy with small regions of larger positive and negative vertical velocities (Fig. 4 b). The distribution of w for this time and pressure is centered around 0 m s^{-1} for both simulations, but for the nested simulation it shows longer tails to higher and lower values (Fig. 4 c). Interestingly, the difference in the tail to

negative w is greater between simulations than towards positive w , suggesting that the global simulation may not accurately represent descending motion in WCBs. Overall, this behavior is consistent across TF2 and across pressure bins (not shown).

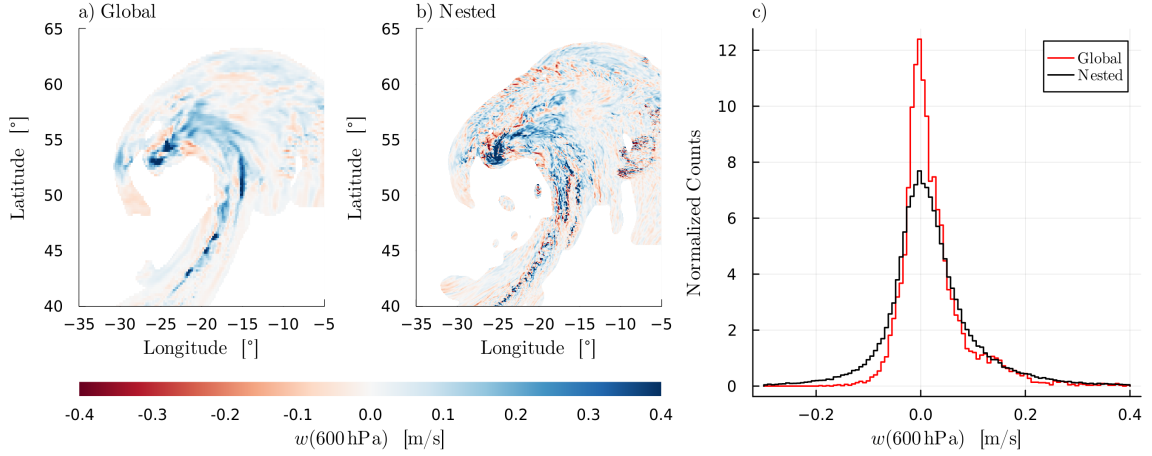


Figure 4. Masked vertical velocities w at 00:00 UTC 2017-09-23 and 600 hPa for the global (a) and nested simulation (b). A normalized histogram of w at this time and pressure (c) is shown for the global simulation (red) and the nested simulation (black).

3.2 UTLS moisture and condensate content

In Schwenk and Miltenberger (2025a) we determined that at WCB outflow pressure levels, WCB trajectories have both a larger relative humidity over ice (RH_i) as well as higher specific humidity (q_v) in the global simulation compared to the nested simulation. In that paper, the detailed Lagrangian analysis suggests that this is most likely due to higher ice number concentrations (N_i) and smaller ice radii (r_i) in the nested simulation, which lead to smaller (faster) relaxation timescales for supersaturation over ice ($\tau_{\text{sat,ice}}$). We additionally speculated that larger q_v values in the global simulation could be due to the parameterized convective moisture and hydrometeor transport in that simulation. However, due to the non-domain filling nature and missing explicit representation of parametrized convective motion in the Lagrangian analysis, the importance of differences in explicit and parameterized water transport are difficult to assess from the Lagrangian perspective. Therefore, in this section we provide an overview of the moisture conditions in the UTLS during TF2 over the Eulerian WCB outflow and compare them with the results and hypotheses proposed in Schwenk and Miltenberger (2025a) based on Lagrangian data.

Overall, the Eulerian analysis these confirms findings with respect to the distribution of RH_i and q_v values (Fig. 5 and 6). From 200 to 250 hPa, the mean and median for RH_i are larger in the global simulation than in the nested simulation (Fig. 5 a and b). Additionally in the global simulation, about 30 % of grid points within the WCB mask have RH_i greater than 100 % at pressure levels between 200 to 350 hPa. This is consistently higher than in the nested simulation, for which this figure is around 25 %. For q_v the mean and median are larger in the global simulation at 200 and 250 hPa (Fig. 6 a and b), while they are almost identical both simulations at 400 and 450 hPa (6 c and d). At 250 hPa (Fig. 6 b), there is a pronounced third peak for q_v at approximately 0.12 g kg^{-1} in the global simulation that is absent in the nested simulation. These results indicate that in the global simulation, the UTLS outflow of the WCB is more moist on average, and also has a greater fraction of grid points that experience supersaturated conditions with respect to ice.

The geospatial distributions of RH_i , and especially of regions of supersaturation, are qualitatively similar in both simulations (Fig. 7). This indicates that the grid-resolution does not necessarily lead to large horizontal shifts or to completely different spatial moisture distributions overall. Instead, similarly to the distribution of w (Fig. 4 a and b), the key difference is the 'patchiness' of the RH_i -field; the global simulation exhibits large contiguous areas of supersaturation (Fig. 7 d-f), whereas in the nested simulation these areas are broken up into smaller fragments by regions that experience descending motion (Fig. 7 a-c). This additionally indicates that one reason for the smaller RH_i and lower frequency of supersaturated grid-cells in the nested simulation might be the larger abundance of descending motion (Fig. 4 c). This is an interesting result for those who study supersaturated regions in the UTLS (and for instance their fractal characteristics).

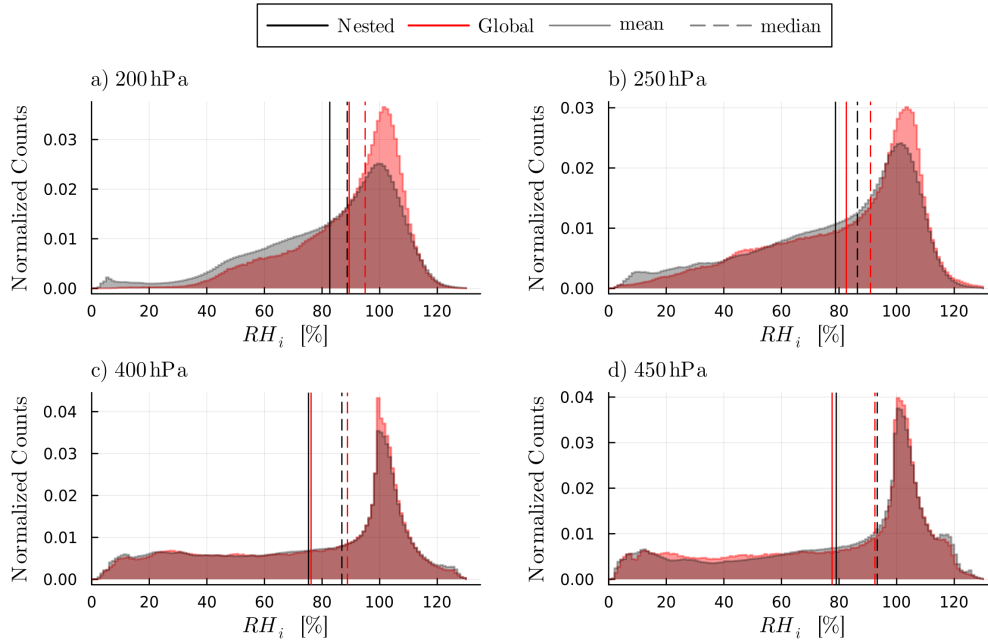


Figure 5. Normalized RH_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

The variables characterizing the ice hydrometeor population (i.e. N_i , q_i and r_i) also show broadly similar behavior in the Lagrangian and Eulerian analysis: Both N_i (Fig. 8) and q_i (Fig. A5) are larger in the nested simulation for pressures of 250 hPa or higher. r_i is much smaller in the nested simulation than in the global simulation and there is a sharp peak at small r_i for pressures of 400 hPa or lower, which is absent from the global simulation (Fig. 9). Schwenk and Miltenberger (2024) determined that convectively ascending air parcels contain ice crystals with much smaller r_i values than air parcels that ascend more slowly. Therefore, the peak at small r_i in the nested simulation can be very likely attributed to convectively ascending air, and its absence in the global simulation suggests that the convection parameterisation produces much larger r_i for detrained ice.

We examine the same hypothesis from Schwenk and Miltenberger (2025a) that larger N_i in the nested simulation can cause the smaller q_v and RH_i . While larger values for N_i decrease $\tau_{\text{sat,ice}}$, smaller r_i do the opposite (Schwenk and Miltenberger, 2025a; Schwenk et al., 2025). For pressures of 250 hPa and higher, the nested simulation has larger N_i and smaller r_i than the global simulation. However, N_i still determines the overall response of $\tau_{\text{sat,ice}}$, because its increase is stronger than the decrease of r_i in the nested simulation (note: logarithmic x-scale in Fig. 8 and linear x-scale in Fig. 9): in the nested simulation $\tau_{\text{sat,ice}}$ is either similar or slightly smaller (Fig. A4), which could

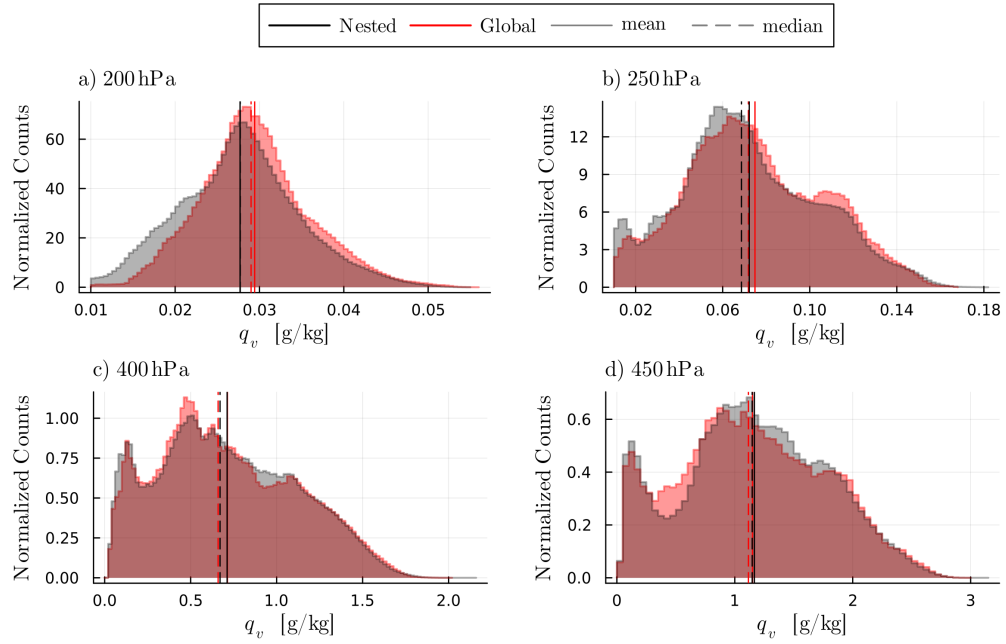


Figure 6. Normalized q_v distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

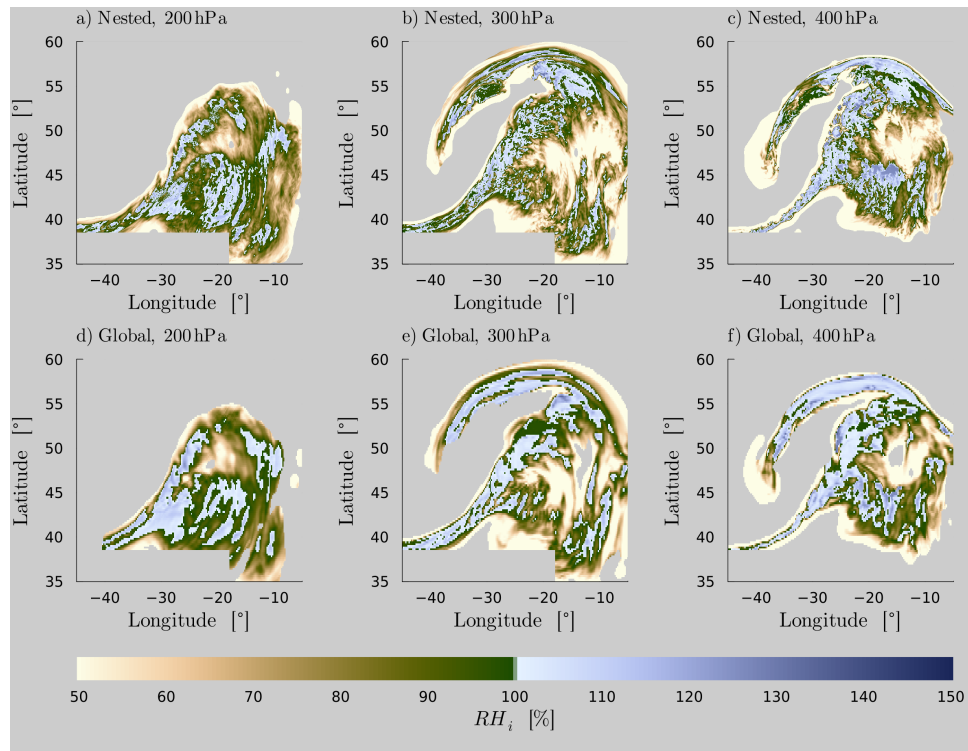


Figure 7. RH_i fields at 12:00 UTC 2017-09-22 for the nested simulation (top row) and the global simulation (bottom row) at 200 hPa (a and d), 300 hPa (b and e) and 400 hPa (c and f).

explain the smaller q_v and RH_i . However, in the highest part of the WCB outflow, i.e. 200 hPa, differences in saturation adjustment timescales due to different ice particle populations are not able to explain the shifts in the RH_i and q_v distribution. N_i is larger in the global simulation rather than smaller (Fig. 8 a), and RH_i as well as q_v is still larger than in the nested simulation. Instead, as we also hypothesized in Schwenk and Miltenberger (2025a), the differences in q_v and RH_i at 200 hPa arise due to the convection parameterisation.

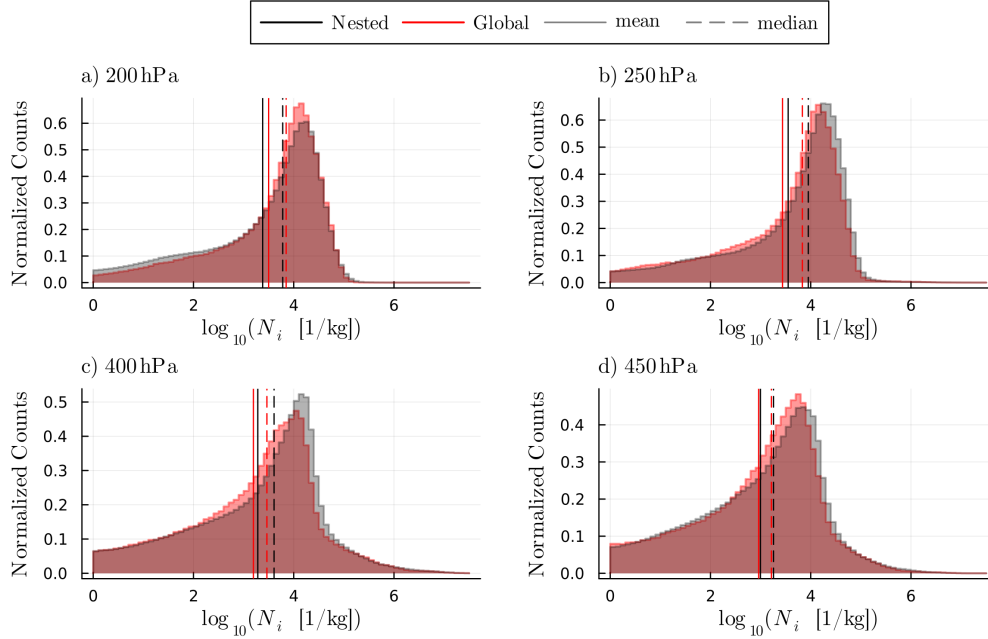


Figure 8. Normalized N_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

In TF2 and for the global simulation, 11% of grid-points in the collapsed WCB mask are influenced by convection. We define a grid-point as "convective" when the convective precipitation over the preceding 15 min exceeds 0.01 kg m^{-2} . This definition holds only for that point in time, i.e. we do not "track" the convective areas. At 200 and 250 hPa, the convective data points in the WCB mask have a very different q_v to that of the non-convectively influenced data points: The former on average show much larger q_v values (Fig. 10 a and b). As these areas account for a substantial proportion of the data points, it is obvious that this shifts the q_v distribution at 200 hPa to larger values (Fig. 6 a) and introduces a second maximum at $q_v = 0.12 \text{ g kg}^{-1}$ at 250 hPa (Fig. 10 b, Fig. 6 b). For pressures of 400 hPa and larger, the data results in bi-modal q_v distributions with one peak at small q_v and another at large q_v (Fig. 10 c and d).

These differences in the q_v distributions for convective and non-convective regions clearly show that the convection parameterization produces moister conditions in the WCB outflow than regions where there is no convection (in that time step; a background signal from convection is expected in the "non convective" areas). In the nested simulation, where convection is not represented by a convection parameterisation, similar patterns cannot be identified, i.e. either the shift at 200 hPa nor a bimodality of the q_v distribution at higher pressures is evident. Schwenk and Miltenberger (2024) also identified substantial convective motion in the high-resolution simulation of this WCB case. Therefore, the differences in the q_v distributions suggests that the convection parameterisation overestimates the amount of moisture transported into the UTLS.

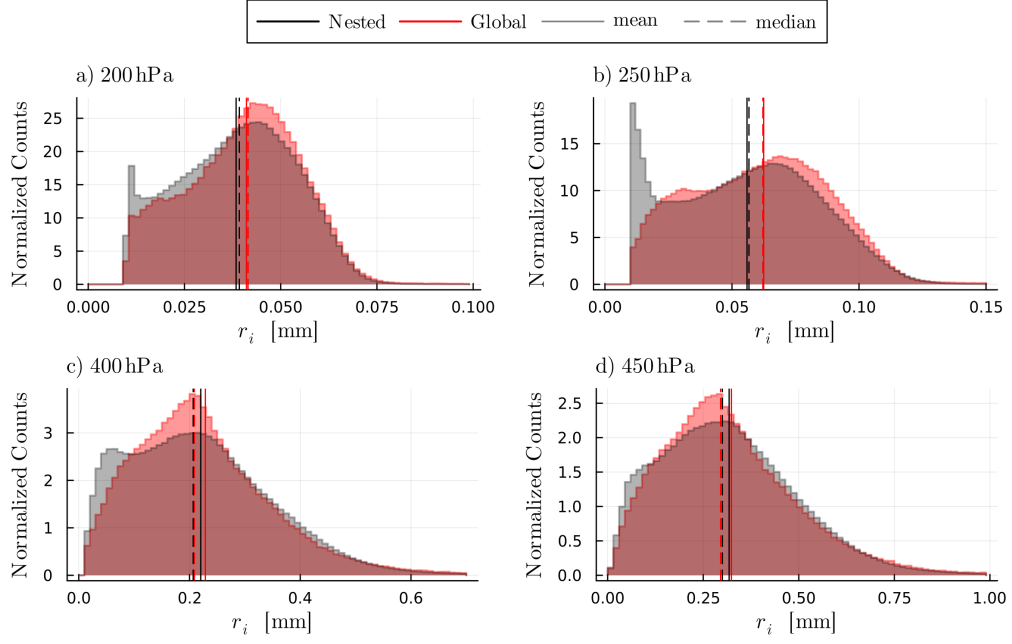


Figure 9. Normalized r_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

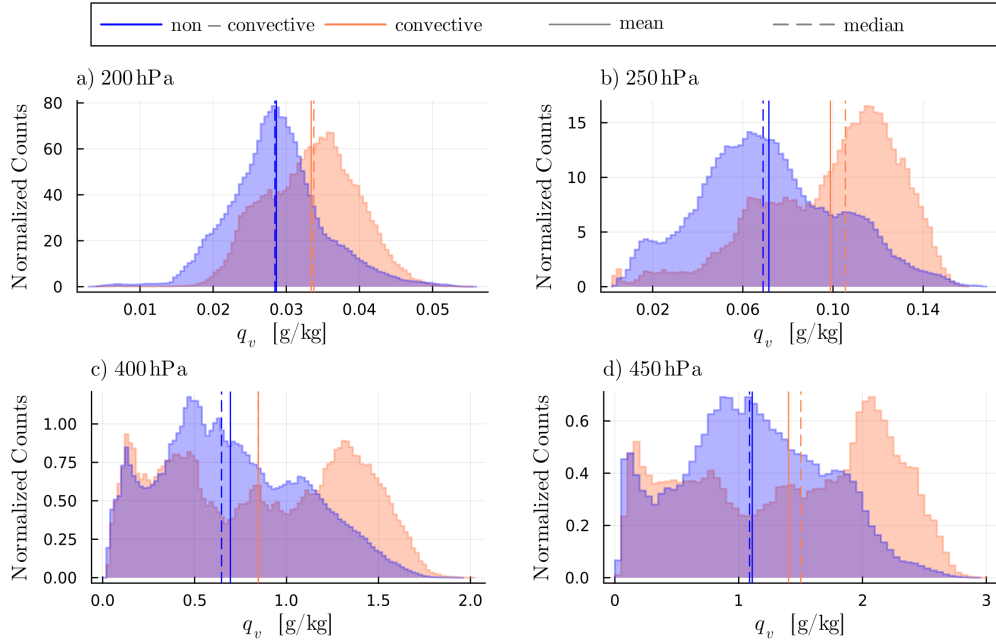


Figure 10. Normalized q_v distributions at different pressure levels for grid-points that experience convection (orange) and that do not (blue) in the global simulation only, in TF2. The mean (solid) and median (dashed) values are indicated by vertical lines.

3.3 Cloudiness

The total, low, medium and high cloud cover values reflect the enhanced patchiness of simulated fields in the nested simulation (Fig. 11). The nested simulation experiences a greater proportion of grid points with cloud cover of any type greater than 95% than the global simulation. However, they also experience a greater proportion of grid points with total cloud cover less than 5%. This is not entirely unexpected, because grid points in the global simulation represent a larger area and therefore must average out the cloud cover, leading to less extreme values. However, the only value that reverses this trend is the percentage of grid-points with a low cloud cover smaller than 5%, which is larger in the global simulation. Grid-points that are influenced by the convection parameterization in the global simulation produce higher low cloud cover values. Therefore, a much stronger preference for low-cloud production by the convection parameterization might explain this difference. However, any deeper analysis into the cloud cover is beyond the scope of this study.

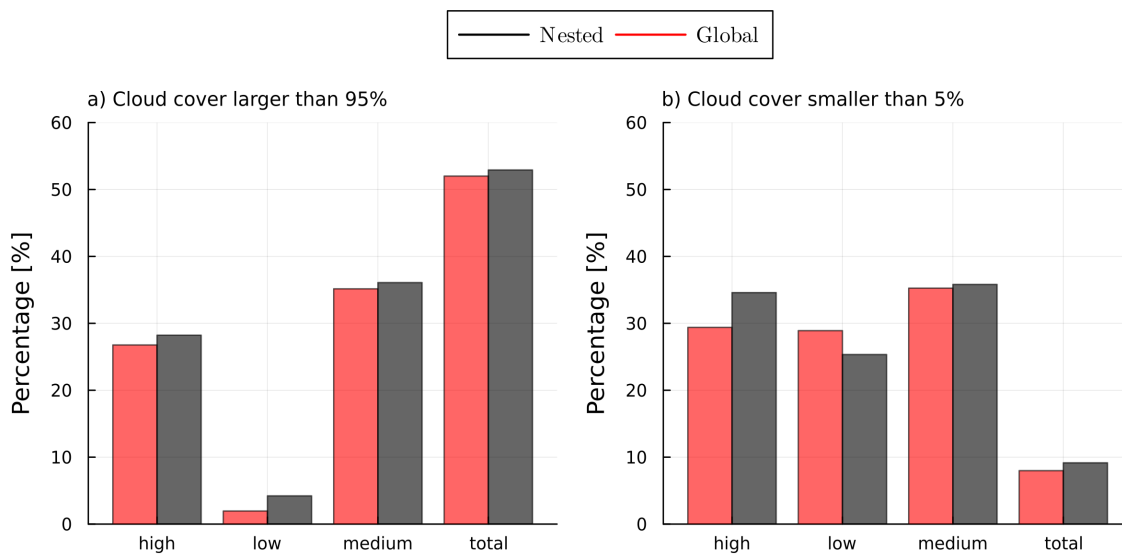


Figure 11. Bar chart showing the percentage of grid points during TF2 that have high, low, medium and total cloud cover larger than 95% (a) and smaller than 5% (b) for the global simulation (red) and the nested simulation (black).

In Schwenk and Miltenberger (2025a), we investigated the Lagrangian hydrometeor content at pressure levels larger than 500 hPa and concluded that overall, the nested simulation produces more hydrometeors, which should result in thicker clouds. The largest differences were found for graupel and rain, which are much higher in the nested simulation. We attributed this to the higher vertical velocities in the nested simulation that allow for the retention of larger hydrometeors.

For the Eulerian analysis, we do not investigate the hydrometeor content at pressure bins larger than 500 hPa, because in this lower region of the atmosphere the WCB masks are not produced using dilation (see Sec. 2.2) and therefore will mostly represent the trajectory data. Instead, we investigate the total column integrated hydrometeor mass for longitude-latitude points that lie within the collapsed WCB mask (i.e., all points that have at least one WCB grid-point above them; Fig. A1 a). Overall, the total column integrated hydrometeor mass (Fig. 12 c) is larger in the nested simulation. The difference is largest for rain (Fig. 12 b) and graupel (Fig. 12 c), and smallest for snow (Fig. 12 e). This is in line with our findings from Schwenk and Miltenberger (2025a) and therefore indicates that the WCB in the nested simulation does indeed produce thicker clouds with more graupel and rain.

In summary, for the time frame during which most of the WCB ascent occurs (TF2), we were able to confirm many of the results from Schwenk and Miltenberger (2025a) and support the hypotheses we formulated to explain the differences observed between the simulations. In particular, we are able to clearly identify the influence of the convection parameterisation on q_v in the global simulation, which was not possible in Schwenk and Miltenberger (2025a). Our findings suggest that, although the Lagrangian trajectories did not fill the entire WCB area and in the global simulation were not necessarily directly influenced by the convection parameterization scheme, the results obtained from them are largely valid for the entire WCB. This lends confidence to the results from Schwenk and Miltenberger (2025a) that cannot be verified from the Eulerian data, such as the accumulated microphysical process rates during ascent, and the results for the condensation ratio and precipitation efficiency. In the following we discuss TF3, the time frame during which most of the WCB trajectories are part of the WCB outflow.

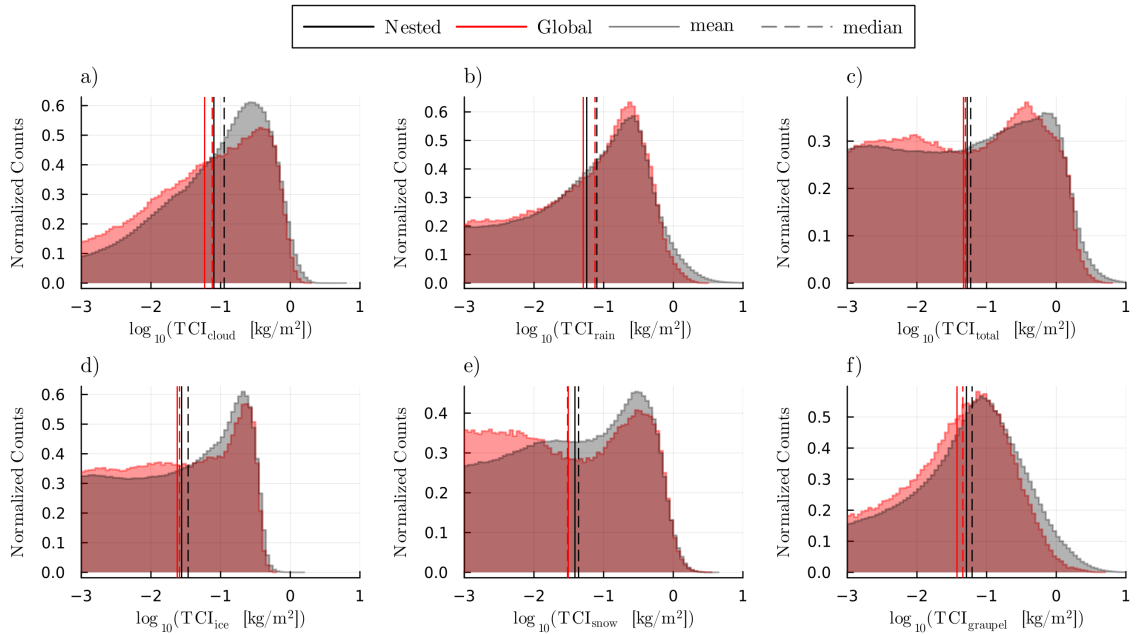


Figure 12. Normalized distributions of total column integrated hydrometeor mass in TF2 for cloud droplets (a), rain drops (b), total hydrometeor mass (c), ice (d), snow (e) and graupel (f). The mean (solid) and median (dashed) values are indicated by vertical lines.

4 Differences in WCB properties during outflow-dominated stage (TF3)

In Schwenk and Miltenberger (2025a), we established that in TF3, the trajectories of the global simulation remained slightly more moist than in the nested simulation at pressures below 250 hPa. In our Eulerian results, we observe similar behavior for RH_i and q_v , even though this is not necessarily reflected in the mean and median values. For RH_i , the distributions at 200 and 250 hPa in the nested simulation have slightly longer tails for larger RH_i than the global simulation (Fig. 13 a and b). However, the peaks of the distributions in the global simulation are shifted slightly towards larger RH_i values (Fig. 13 a and b). At 400 and 450 hPa, RH_i distributions are similar in both simulations and bimodal, but the second peak at smaller RH_i is larger in the global simulation, which results in smaller mean and median RH_i (Fig. 13c and d). For q_v , the distributions at 200 and 250 hPa (Fig. 14 a and b) are very similar to those in TF2 (Fig. 6 a and b), with the difference that q_v is no longer much larger in the global simulation. However, at 250 hPa the second peak at $q_v = 0.12 \text{ g kg}^{-1}$ for the global simulation remains, and the mean is slightly higher. For pressures of 400 hPa and higher, the mean and

median q_v are similar but the peaks of the multi-modal distributions are shifted relative to each other (Fig. 14 c and d).

Similarly to TF2, we can attribute our observations for TF3 to the influence of the convection parameterisation, since the distributions for q_v in these areas are very similar to TF2 (Fig. 10). However, during TF3 and in the global simulation, convective grid points in the collapsed mask account for only around 6% of the data points, compared to 11% for TF2. This difference explains why the differences in mean and median values are not as pronounced, and implies that the differences in UTLS moisture conditions between simulations are largest when convective activity is strongest.

Our Eulerian results for the ice content in TF3 do not necessarily reflect our Lagrangian results from Schwenk and Miltenberger (2025a). In that paper, we determined that in TF3 both q_i and N_i are larger in the global simulation for pressures larger than 400 hPa, and smaller below (except for q_i at 200 hPa, which is larger in the global simulation). However, in our Eulerian results we find that N_i (Fig. 15) and q_i (Fig. A6) are smaller in the global simulation from 200 to 450 hPa. At first, this seems unexpected because the data presented here for the global simulation takes convective grid points into account, and these produce much higher N_i and q_i in the UTLS than non-convective grid points (Fig. A7 and A8, respectively for TF2).

Since the Lagrangian data does not fill the entire WCB area, we assumed that the effects of the convection parameterisation would be less noticeable in the Lagrangian trajectory data than in the Eulerian data. Therefore, we expected the Eulerian results to more strongly represent these grid points and to have higher N_i and q_i than the Lagrangian trajectories (at least compared to the nested simulation). However, in our results we find the opposite to be true: Lagrangian trajectories represent convective areas more strongly than the Eulerian data. In the 30-hour window defined by TF2, 50% of trajectories in the global simulation spend over 10 hours above grid points flagged as convective in our mask (more than 30% of the time; not shown). However, convective grid points only make up 11% of grid points in the Eulerian WCB mask. Therefore, this suggests that online trajectories in the global simulation represent convective grid points more strongly than the Eulerian data does. This could explain why N_i and q_i are larger at certain pressures during TF3 in the global simulation and the Lagrangian data from Schwenk and Miltenberger (2025a), but not so in the Eulerian data presented here.

Another reason why the results for ice content in TF3 from Schwenk and Miltenberger (2025a) are different may be that, during TF3, many trajectories extend beyond the boundaries of domain 3, whereas our Eulerian analysis is confined to this area. On average, these trajectories have been part of the WCB outflow for longer than those inside domain 3. It is possible that the differences produced by the simulations within this domain propagate into the other domains (where the simulations have the same grid spacing) in unexpected ways.

5 Results - Radiation

In Schwenk and Miltenberger (2025a), we hypothesized that the differences in UTLS hydrometeor and vapor content between the two simulations might also be reflected in differences in cloud top temperature (CTT) as well as outgoing longwave radiation (OLR) and at the top of the atmosphere (TOA). To test this hypothesis in this study, we calculated the CTT by taking the collapsed two-dimensional mask at each time step and grid point, and setting the temperature at the first model level (from above) at which the total hydrometeor mass exceeds $10^{-7} \text{ kg kg}^{-1}$ as the CTT. We find that the

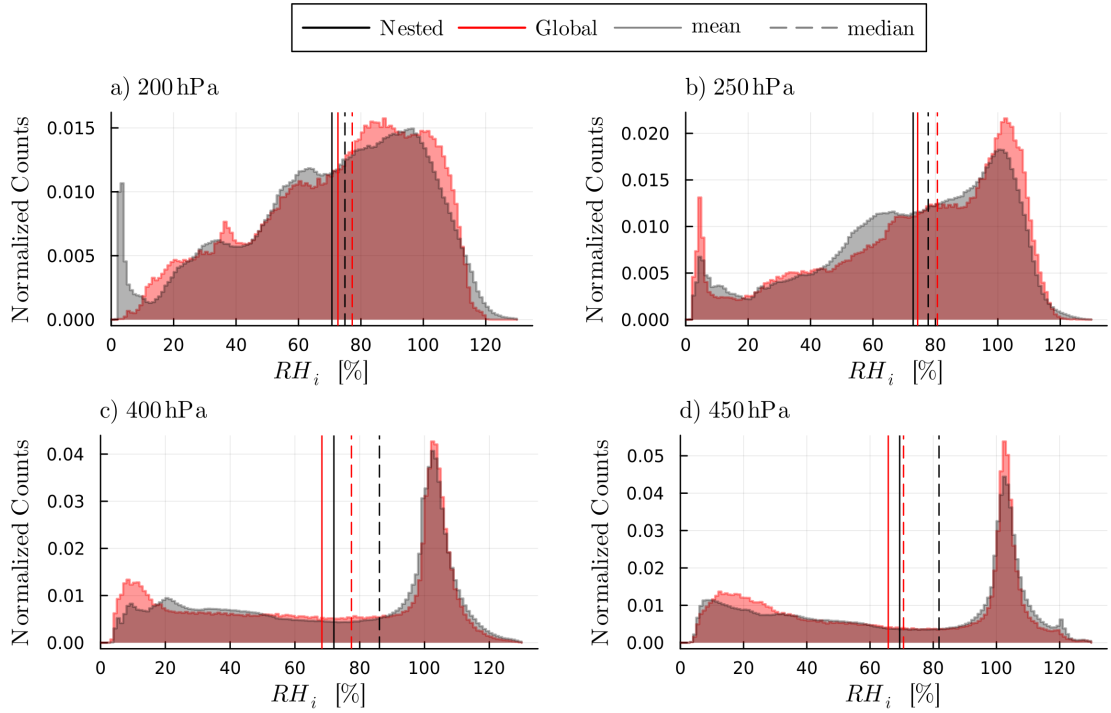


Figure 13. Normalized RH_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF3, with mean (solid) and median (dashed) values indicated by vertical lines.

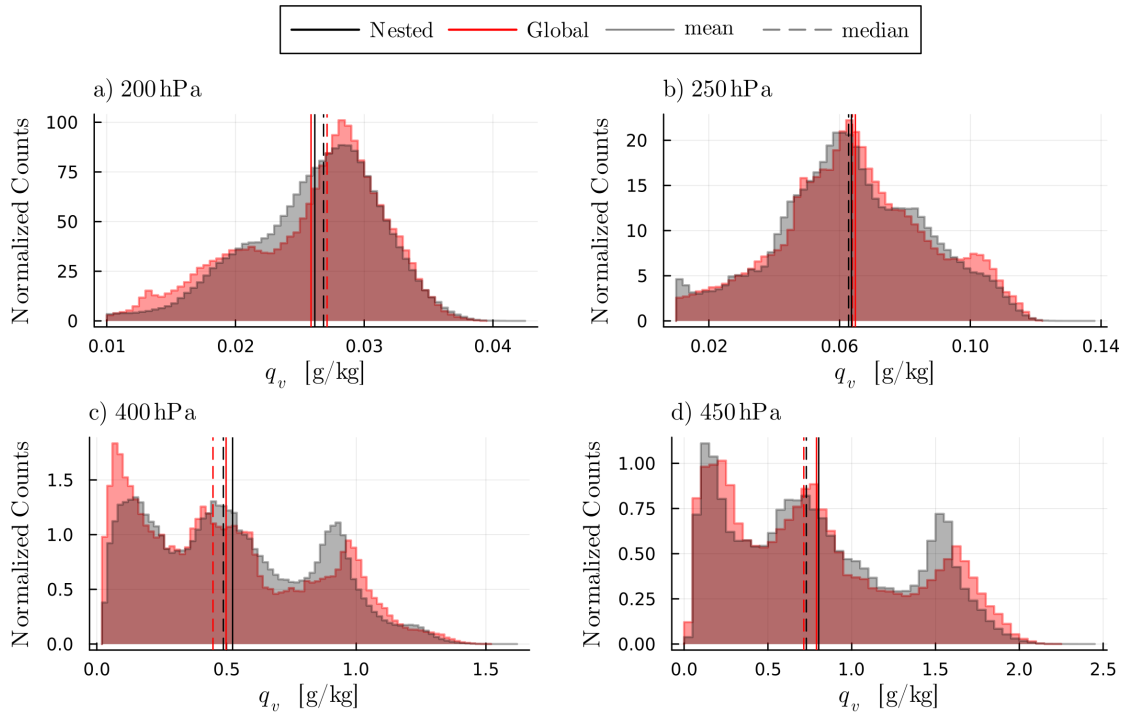


Figure 14. Normalized q_v distributions at different pressure levels for the global (red) and the nested (black) simulations in TF3, with mean (solid) and median (dashed) values indicated by vertical lines.

differences in CTT between simulations are almost negligible for both TF2 and TF3 (Fig. 16 a and b). The distributions are very similar and the peaks are located at approximately the same temperatures. However, for TF2 (Fig. 16 a), the mean CTT of -49.5°C in the global simulation is slightly smaller than the mean of -48.8°C in the nested simulation. This is due to a slightly more pronounced second peak at approximately -60°C in the nested simulation (Fig. 16 a). However, this slight difference in CTT during TF2 is not able to explain the larger difference we find for OLR (Fig. 16 b and c), which is clearly shifted to more negative values in TF2 in the nested simulation compared to the global simulation (meaning more outgoing radiation).

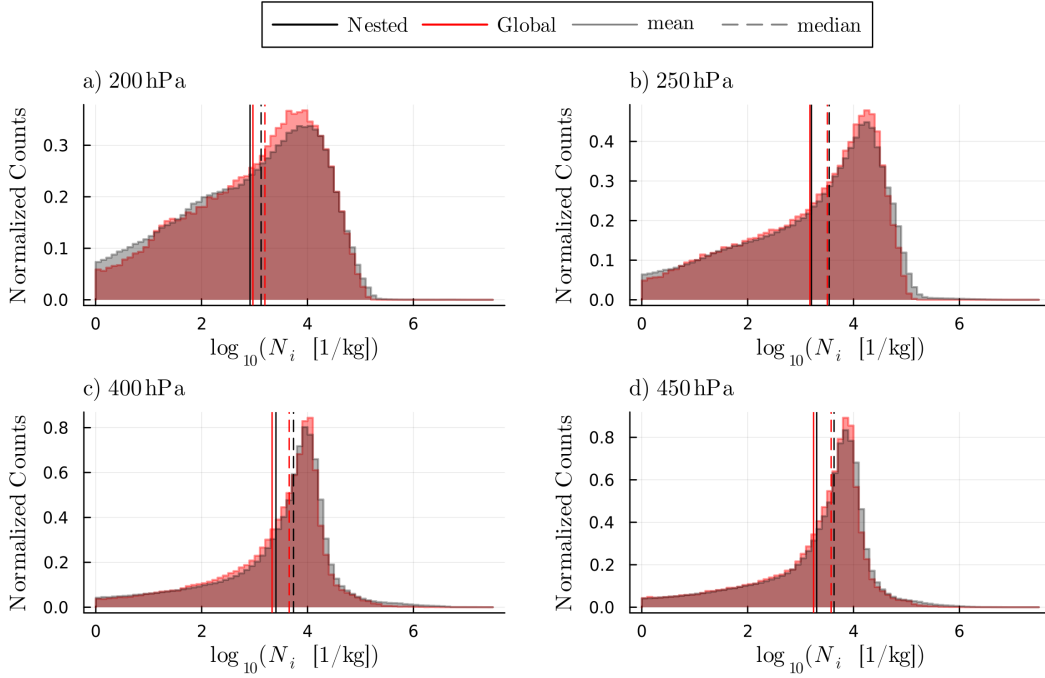


Figure 15. N_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF3, with mean (solid) and median (dashed) values indicated by vertical lines.

The mean OLR in TF2 is -226.6 Wm^{-2} in the nested simulation, which is 4 Wm^{-2} smaller than the mean of -221.5 Wm^{-2} in the global simulation. For the median, this difference is 7 Wm^{-2} (-230.5 Wm^{-2} in nested, -223.3 Wm^{-2} in global). This difference is still present but much smaller during TF3, with mean (median) values of -223.0 Wm^{-2} (-225.3 Wm^{-2}) in the global simulation, and -223.6 Wm^{-2} (-222.6 Wm^{-2}) in the nested simulation. However, the distributions of OLR in TF3 are slightly different, whereas the distributions in TF2 are the same, only shifted. Nevertheless, the relatively large differences in OLR clearly cannot be caused by the minor differences in CTT. Instead, the differences we found for q_v could provide a more plausible explanation.

Upper tropospheric q_v can trap longwave radiation and act as a greenhouse gas. In TF2 and at pressures of 200 and 250 hPa, q_v is larger in the global simulation than in the nested simulation (Fig. 6 a and b). In TF3, q_v at pressures from 200 to 450 hPa are neither substantially smaller or larger on average, but the distributions are shaped differently (Fig. 14). This behavior fits OLR, which is more negative for the nested simulation in TF2, and the distribution is shaped differently in TF3. The peak for CTT is situated at approximately 60°C , which is also approximately the mean temperature at the 200 hPa level during TF2 and TF3. Therefore, the differences in upper-level q_v could lead to the differences in OLR.

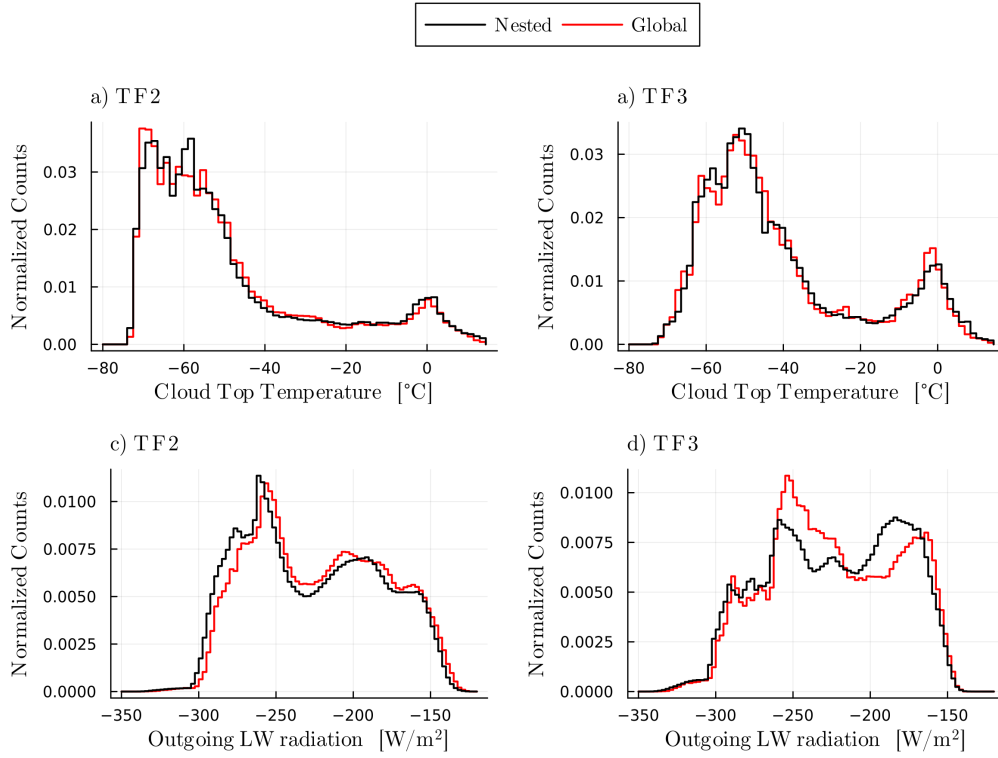


Figure 16. Normalized histograms of cloud top temperature during TF2 (a) and TF3 (b), and the outgoing longwave radiation at the top of the atmosphere during TF2 (c) and TF3 (d) for the nested (black) and global (red) simulation.

To determine whether this hypothesis is plausible, we split the OLR and CTT data into two groups according to q_v at 200 hPa; the first group has $q_v(200 \text{ hPa}) > 0.03 \text{ g kg}^{-1}$, and the second $q_v(200 \text{ hPa}) \in (0.01, 0.02) \text{ g kg}^{-1}$. In both the global and the nested simulations and as long as CTT is between approximately -50°C and -20°C , OLR is much more negative (more outgoing LW radiation) for the second group, when $q_v(200 \text{ hPa})$ is smaller (Fig. 16). When CTT is smaller than -50°C or larger than -20°C , there is not difference in OLR between the groups. For CTT below -20°C this could indicate the following: when cloud tops are situated at around 200 hPa, OLR does not depend on q_v at that pressure level, but instead on CTT. However, when cloud tops are situated below this level at higher pressures (i.e temperatures), q_v can trap the emitted radiation. This clearly shows that differences in upper tropospheric q_v between the simulations could cause the substantial differences observed for OLR.

6 Conclusions and Discussion

This paper is the second part of a three-part study investigating the role of simulation scale for WCB moisture transport into the UTLS. In Schwenk and Miltenberger (2025a) and in this paper, we examine a case study of a WCB that dominated the weather over the northern Atlantic on 23 September 2017, using two ICON simulations: one run at a convection-parameterising resolution of $\sim 13 \text{ km}$ ('global' simulation) and one run at a convection-permitting resolution of $\sim 3.5 \text{ km}$ ('nested' simulation). In the Schwenk and Miltenberger (2025a), our focus was primarily on analyzing Lagrangian trajectory data. A comparison of the WCB outflow in the two simulations revealed substantial differences in UTLS ice content and moisture. We also observed differences in the microphysical processes experienced during the WCB ascent. However, the Lagrangian data does not necessarily cover the

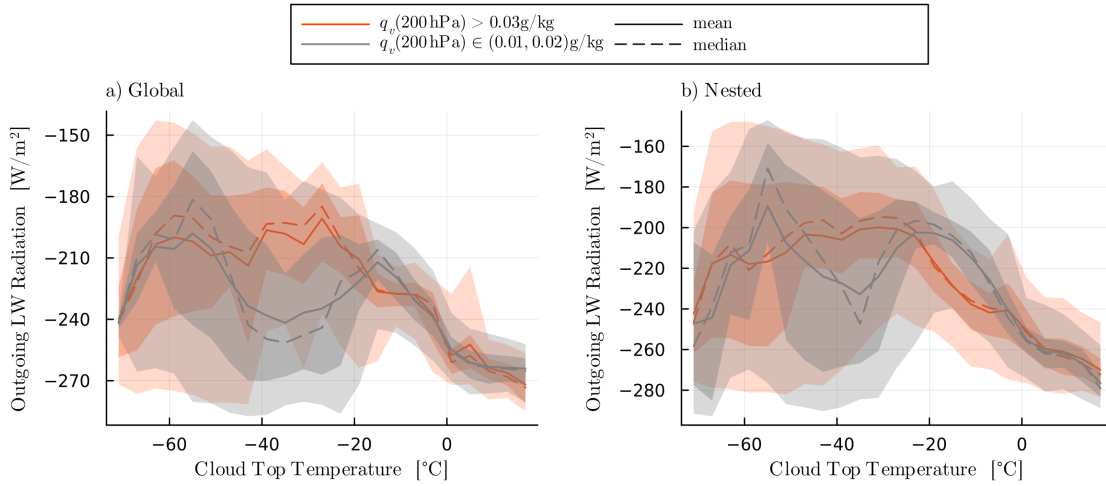


Figure 17. Outgoing longwave radiation at the top of atmosphere as a function of cloud top temperature, but split according to q_v at 200 hPa for the global simulation (a) and the nested simulation (b). Values for which q_v is larger than 0.03 g kg^{-1} are plotted in orange, values for which $q_v \in (0.01, 0.02) \text{ g kg}^{-1}$ in grey. Mean (solid) and median (dashed) lines are plotted, with the 5th to 95th percentiles shaded in light, and the interquartile range in dark.

entire WCB area, meaning important regions of the WCB may not be accurately represented. Additionally, the convection parameterisation, which redistributes moisture and hydrometeors vertically, does not affect the Eulerian wind fields in the model, but these wind fields are used in the Lagrangian trajectory module to advect air parcels. This means that, in the global simulation, trajectories traveling through convective regions might not ascend appropriately, which could introduce an additional bias into the data. Therefore, in this paper, we conducted an additional Eulerian comparison of the simulation outputs to verify or improve upon the results of the Lagrangian analysis.

A crucial aspect of a robust Eulerian analysis focusing on a specific area is the creation of a mask to isolate that area. In this paper, we present a novel and efficient algorithm for producing a Eulerian WCB mask from trajectory data. At each time step and pressure level, the relevant trajectory positions are gathered and converted into a preliminary Eulerian mask. This mask is then gradually expanded ('dilated') to fill the entire WCB area at that pressure level using level-specific q_v thresholds (Section 2.2). To our knowledge, this approach has not been used before, and it could be useful for creating multiple WCB masks using only sparse trajectory data in climatological studies. This could save computational resources while improving the representation of WCB in such studies. However, the algorithm needs to be refined first, as the exact implementation of the masking algorithm in this study involves many custom thresholds and manipulations that cannot be achieved on a climatological scale. Nevertheless, the algorithm presented here could be a useful starting point for future improvements and could be an alternative or supplement to the machine learning WCB identifying algorithm by Quinting and Grams (2022).

In Schwenk and Miltenberger (2025a), we mainly attributed the differences in hydrometeor and vapor content in the WCB outflow between simulations to the different ascent velocities. The trajectories in the nested simulation exhibit much more rapid periods of ascent and descent, leading us to hypothesize that the nested simulation exhibits much more dynamic and chaotic vertical flow overall. Here, we verify this assumption by showing that, although the vertical velocity distributions within the WCB mask in both simulations are centered around zero, the distributions in the nested simulation extend to larger negative and positive values (Fig. 4). Higher vertical velocities allow for

the retention of more and larger hydrometeors, which was the reason why trajectories in the nested simulation showed much higher total hydrometeor content and specifically more graupel during the ascent. Here, we were able to generalize this for the entire WCB by examining the total column integrated hydrometeor content underneath the WCB mask (Fig. 12, which is larger overall in the nested simulation, and much larger in the case of graupel and rain.

Regarding UTLS moisture, Schwenk and Miltenberger (2025a) found that the WCB outflow in higher regions of the atmosphere is more moist in the global simulation than in the nested simulation. We attributed this to the different N_i values, which were much higher in the nested simulation, allowing for a faster reduction of supersaturation by reducing the relaxation timescale of supersaturation over ice $\tau_{\text{sat,ice}}$. In this study, we also find that in the global simulation, the WCB outflow is more moist at pressures of approximately 350 hPa and below. This was observed for both the specific humidity (Fig. 6) and the relative humidity over ice (Fig. 5). For pressures larger than 200 hPa, this can be attributed to higher N_i as in Schwenk and Miltenberger (2025a) (Fig. 8). However, at 200 hPa this explanation is not valid, because N_i is not higher in the nested simulation, but the differences in vapor remain.

We therefore identified a second mechanism to explain the differences in humidity: the influence of the convection parameterization in the global simulation. In the global simulation, regions within the WCB mask that exist over areas of convective precipitation have much higher q_v values than all other points within the WCB mask (Fig. 10). If the representation of upper-level moisture and hydrometeor content in a convective atmospheric column by the convection parameterization were consistent with that in an explicitly resolved convective column, there would be no difference in upper-level moisture between the two simulations. However, Schwenk and Miltenberger (2024) found that convective ascending air in a convection permitting simulation produces very high N_i and thus small $\tau_{\text{sat,ice}}$, leading to RH_i values that peak below 100%. This results in less upper-level moisture for a convective ascending atmospheric column, not more. Therefore, the difference in convective ascending air between the convection-parameterizing global simulation and the convection-permitting nested simulation is another factor contributing to differences in upper-level moisture.

In the global simulation, the convection parameterization produces larger ice-crystals in the WCB outflow than one would expect given the results from Schwenk and Miltenberger (2024), which show that fast ascending air produces smaller ice crystals than slower ascending air. The distribution of r_i -values for the global simulation lacks the bi-modalilty that is prominent in the nested simulation, especially for pressures of 350 hPa and smaller (Fig. 9). Assuming that a simulation that allows for convection better represents reality than one using convection parameterization, the results presented here for upper-tropospheric moisture and hydrometeor content suggest the latter fails to accurately simulate moisture transport into the UTLS.

In TF3, after the primary ascent phase has passed, we find slightly different results than in Schwenk and Miltenberger (2025a), with humidity and ice differences remaining mostly the same as in TF2. We hypothesize that this is because the trajectory data from the global simulation over-represents areas of convection, while the Eulerian mask encompasses larger regions, including those without convection. Overall, differences remain and hint at weaknesses of the convection parameterization.

We also investigated differences in the spatial distribution of variables between the two simulations, and found that some variables are different in their 'patchiness'. The global simulation shows large contiguous areas of similar RH_i , whereas these are broken into smaller fragments in the nested sim-

ulation (Fig. 7). This increased patchiness also leads to the WCB in the nested simulation having a higher proportion of grid points that are both fully clouded and not clouded at all (Fig. 11). This could be an interesting topic for further research, especially since the characteristics of supersaturated regions in the UTLS are increasingly being studied (Schuh et al., 2025). Our limited analysis of geographical shifts in the distribution of variables over time and space indicates that differences are minimal. However, at certain pressure levels and times, there are noteworthy differences in ice content (not shown) that are not important for the overall budget analyzed here but could be an interesting topic for future research.

The differences we have identified between the two simulations lead to no observable difference in cloud top pressures or temperatures, but nevertheless to different amounts of OLR and TOA (Fig. 16). In the nested simulation, more radiation escapes into space. Since cloud top temperatures, which typically limit OLR, remain the same, the increase in outgoing radiation must be caused by the vapor content in the upper troposphere. We show that this is a plausible explanation by filtering OLR values q_v at 200 hPa, where one can see that OLR become more negative when q_v is small (Fig. 17). This further suggests that simulations that parameterize convection might falsely (over)estimate the contribution of WCBs to the radiative budget of Earth because we have shown that they may transport too much vapor into the UTLS.

In summary, in this paper we (i) developed a novel and potentially useful algorithm for creating WCB masking in Eulerian data using Lagrangian trajectories; (ii) clearly showed that simulations that parameterize convection may be substantially biased when modeling the transport of moisture into the UTLS by WCBs, and (iii) demonstrated that a purely Lagrangian analysis produces similar results to the much more cumbersome Eulerian analysis. This means that future studies examining only Lagrangian data in WCBs can be (carefully) considered very robust in representing the entire WCB, as long as the data is not too sparse.

Outlook

Our results show that the convection parameterization produces WCB outflow that is too moist, with too few ice crystals that are too large, when compared to the higher resolution simulation. This indicates that global climate models might overestimate the transport of moisture into the UTLS and incorrectly depict the properties of cirrus clouds in WCB outflows. Given the assumption that a future increase in sea surface temperatures will increase the occurrence of convection, this bias could increase further still. A larger study comparing WCB simulations at different resolutions could therefore be worthwhile. Such a study could compare atmospheric columns in convective areas between simulations and aim to improve the representation of vertical transport of moisture and representation of cirrus clouds by the convection parameterization scheme.

However, the conclusion that the convection parameterization leads to an unrealistic UTLS moisture bias is based on the assumption that the simulation output from the convection-permitting simulation better represents reality. However, this assumption may not be valid. Therefore, model output for simulations run at different resolutions should be compared to measurements, to determine the extent to which a convection-permitting simulation is more consistent with reality than a convection parameterizing simulation. This is precisely the aim of the third part of this study (Schwenk and Miltenberger, 2025b), where we translate the simulation output from the nested and global simulations from this study into synthetic satellite measurements using RTTOV (Saunders et al., 2018), and compare them to measurements from the Meteosat Second Generation satellite (Schmetz et al., 2002) using object tracking (e.g. TOBAC; Sokolowsky et al. (2024)) and fractal analysis (as in Freis-

chem et al. (2024)). The aim is to quantitatively measure consistency between simulation output and observations.

Code and data availability. Code and data will be made available after publication and will appear under the TPChange community on zenodo: <https://zenodo.org/communities/tpchange>. The code will also appear on my GitHub page <https://github.com/CornelisSchwenk/>.

Author contributions. CS and AM designed the experiment and conducted the numerical simulations. CS wrote the post-processing code. CS and AM worked jointly on the interpretation of the results. CS drafted the final manuscript with contributions from AM.

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Appendix A: Additional Figures

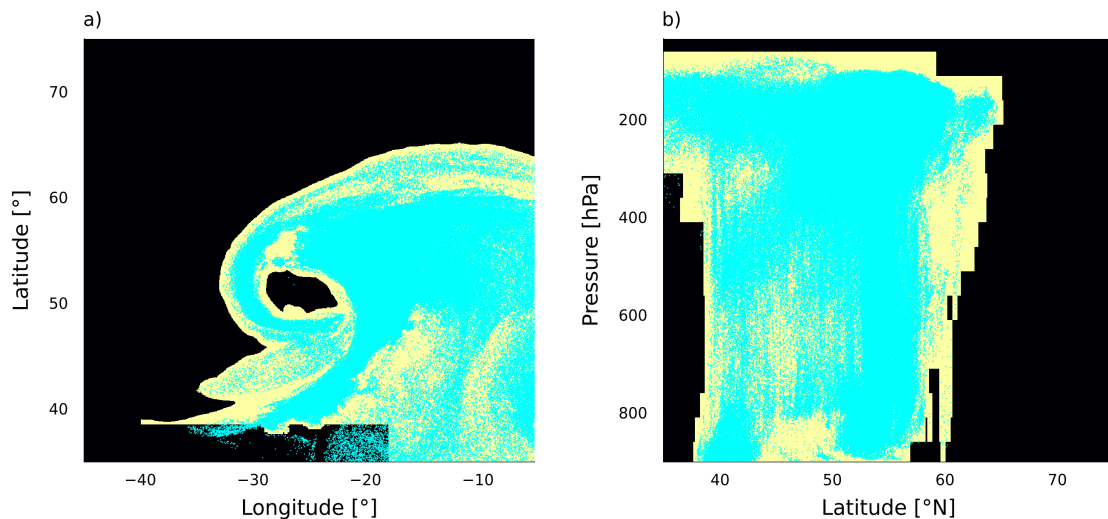


Figure A1. Example depiction of how the Eulerian WCB masks (black is false, yellow is true) "wraps" around the WCB trajectory positions (cyan) as seen from above (a) and from the side (b) at the model run-time of 144 h (00:00 UTC on the 23. September, 2017) for the nested simulation within the boundaries of domain 3.

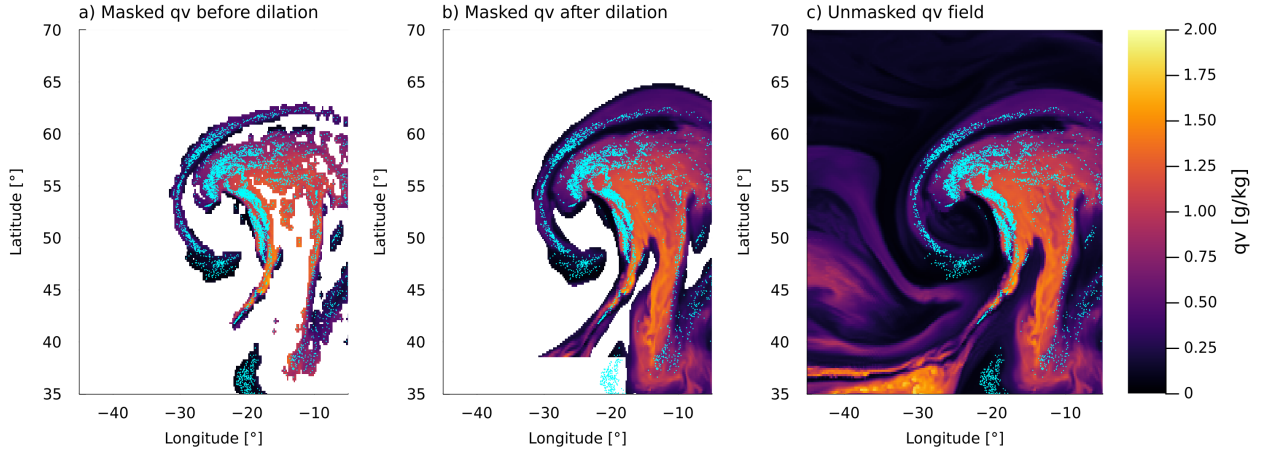


Figure A2. Heatmaps showing q_v at 400 hPa on 00:00 UTC 23 September 2017 for the global simulation in domain 3 masked before dilation (a), masked after dilation (b), and unmasked (c). Cyan points indicate trajectory positions in the vicinity of the 400 hPa pressure level.

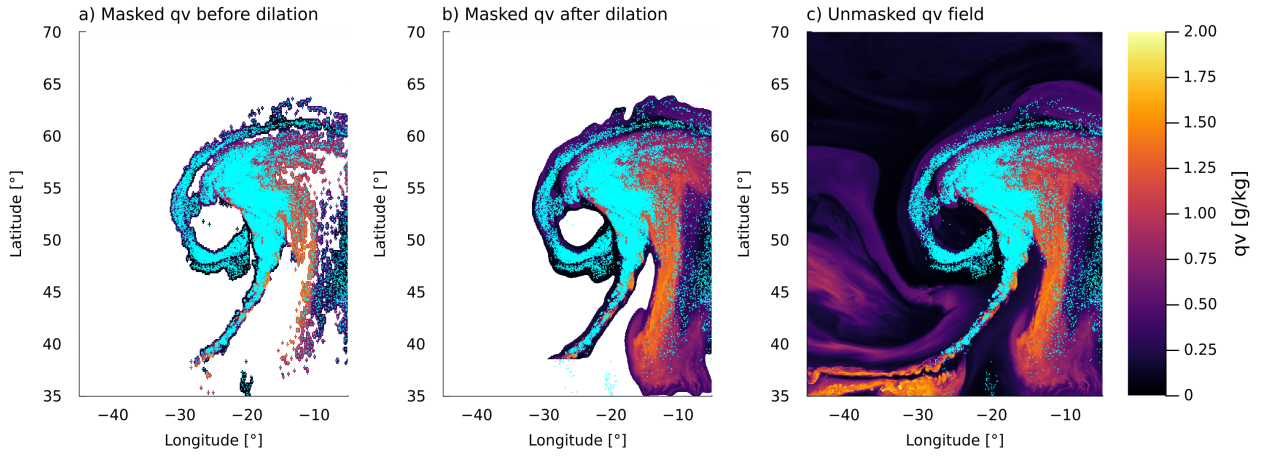


Figure A3. Heatmaps showing q_v at 400 hPa on 00:00 UTC 23 September 2017 for the nested simulation in domain 3 masked before dilation (a), masked after dilation (b), and unmasked (c). Cyan points indicate trajectory positions in the vicinity of the 400 hPa pressure level.

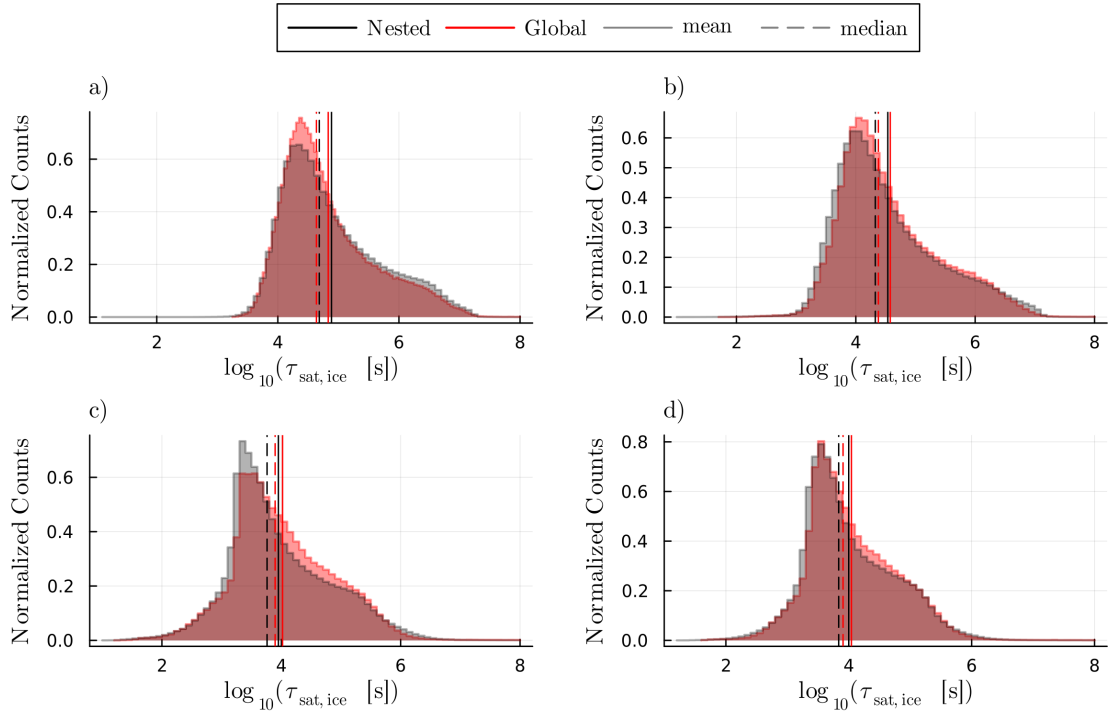


Figure A4. Normalized $\tau_{\text{sat,ice}}$ distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

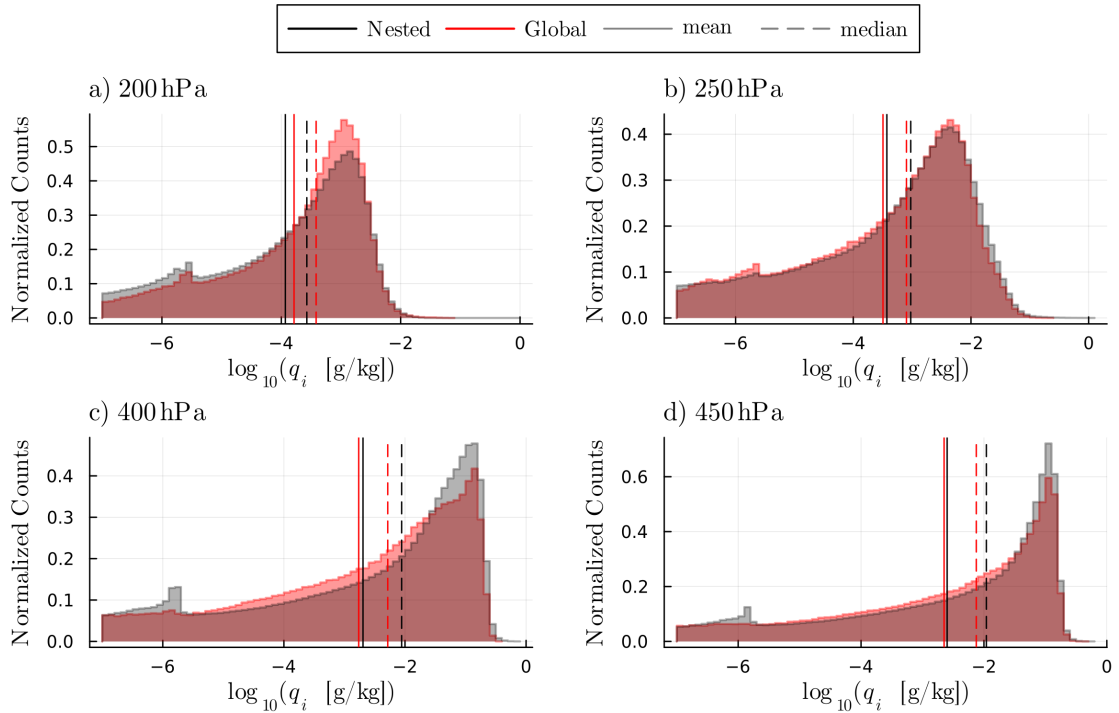


Figure A5. Normalized q_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF2, with mean (solid) and median (dashed) values indicated by vertical lines.

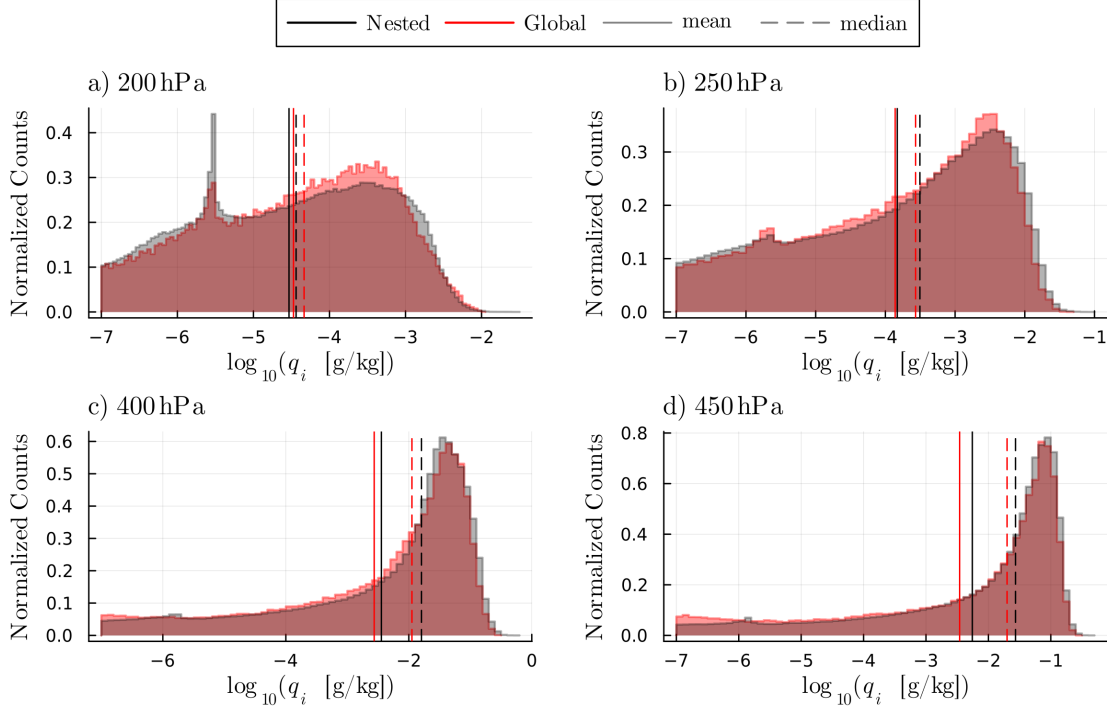


Figure A6. Normalized q_i distributions at different pressure levels for the global (red) and the nested (black) simulations in TF3, with mean (solid) and median (dashed) values indicated by vertical lines.

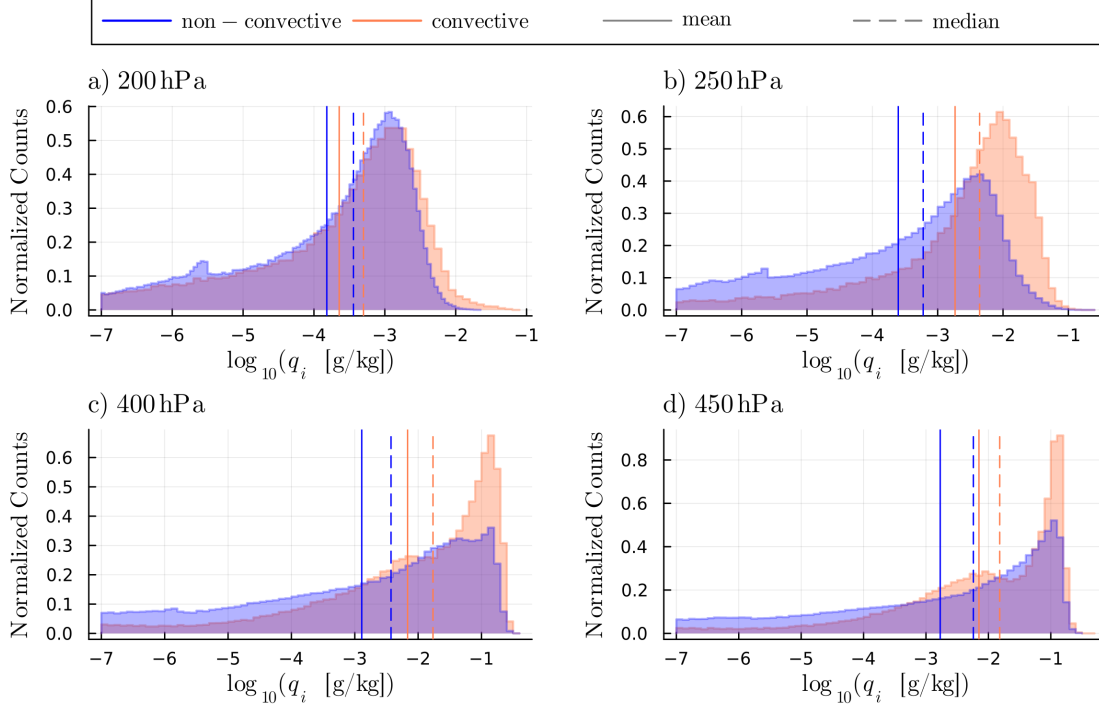


Figure A7. Normalized q_i distributions at different pressure levels for grid-points that experience convection (orange) and that do not (blue) in the global simulation only, in TF2. The mean (solid) and median (dashed) values are indicated by vertical lines.

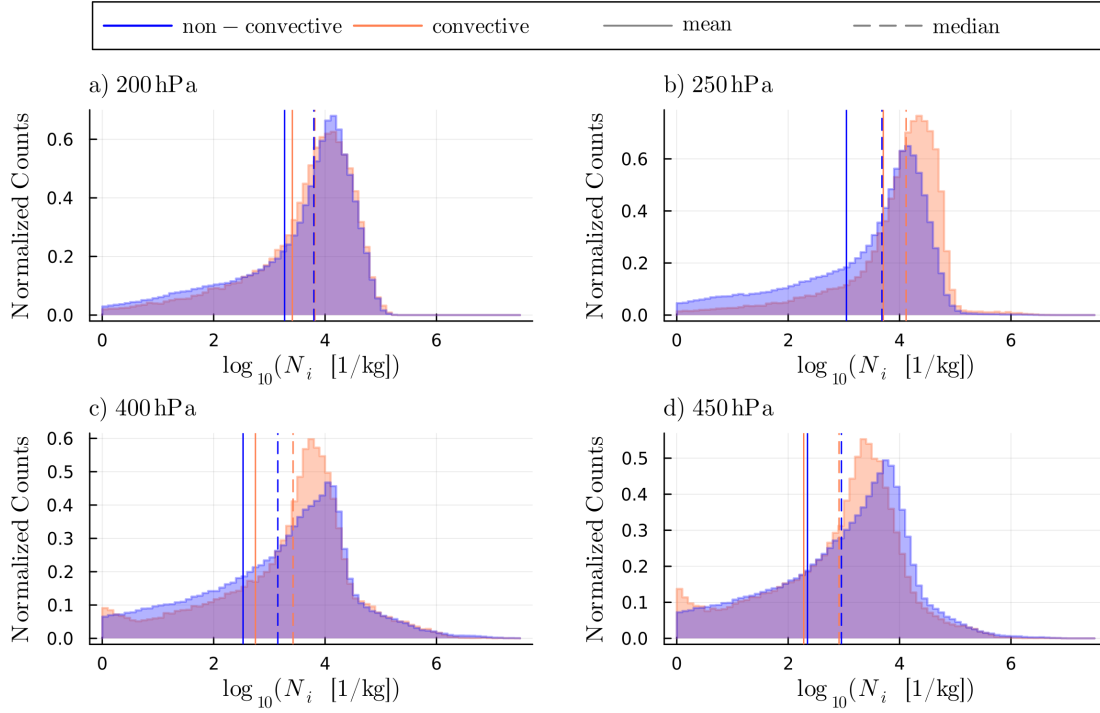


Figure A8. Normalized N_i distributions at different pressure levels for grid-points that experience convection (orange) and that do not (blue) in the global simulation only, in TF2. The mean (solid) and median (dashed) values are indicated by vertical lines.

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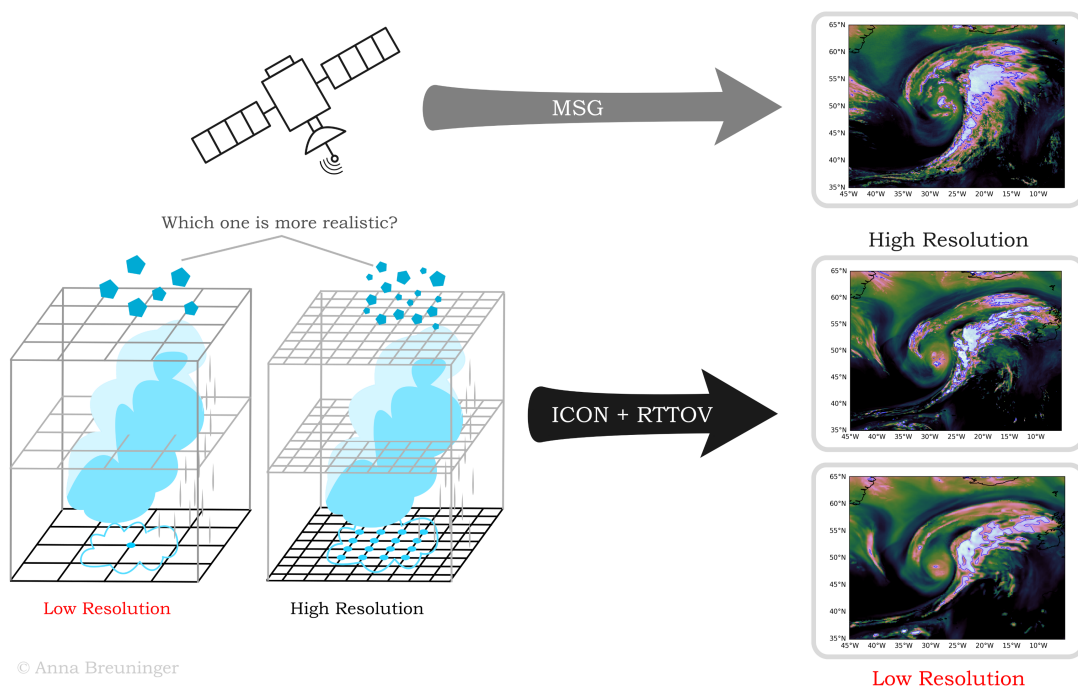
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Chapter 6

The Role of Scale - Part III: Comparison with Observations

Research Question II b

Are convection-permitting simulations better at simulating UTLS moisture in the outflow of a WCB than convection-parameterizing simulations?



This Chapter is in preparation to be submitted to *Atmospheric Chemistry & Physics* under the title "Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS)— Part III: Comparison with Observations" (Schwenk and Miltenberger, 2025c). For author contributions, see Section 2.5.

Effects of Model Grid Spacing for Warm Conveyor Belt (WCB) Moisture Transport into the Upper Troposphere and Lower Stratosphere (UTLS)— Part III: Comparison with Observations

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Abstract. In the field of atmospheric sciences, studies examining weather phenomena using numerical weather prediction (NWP) models often compare the results of simulations run with different configurations, such as different models, resolutions and parameterization schemes. When comparing models with different grid-resolutions, it is often assumed that higher-resolution simulations more accurately represent reality. However, this is not necessarily true. When studies reveal significant differences between higher and lower resolution simulations that could have far-reaching consequences, it is necessary to validate whether it is even likely that the higher resolution is more realistic. However, this is a very difficult task, especially for regions of the atmosphere where measurements are sparse or unreliable. In this study, we present a new method for comparing NWP simulations with geostationary satellite observations that focuses on the upper troposphere / lower stratosphere. This method involves translating NWP model output into a synthetic satellite image using the radiative transfer model RTTOV. Then, we perform fractal analysis and object tracking with TOBAC on the resulting products related to cloud height, phase, and particle size. We applied this method to the warm conveyor belt (WCB) case study from previous sections of this study, which analyzed the case using two ICON simulations run at convection-permitting (high) and convection-parameterizing (low) resolutions. Our results show that the higher-resolution simulation more accurately represents the three products than the lower-resolution simulation. Therefore, we can have confidence in the results from the two previous parts of this study. However, both simulations failed to represent the cold front in the southern regions of the WCB, meaning that we must recognize that the higher-resolution simulation still substantially falls short of fully representing reality.

1 Motivation

Throughout Parts I (Schwenk and Miltenberger, 2025a) and II (Schwenk and Miltenberger, 2025b) of this study, as well as in Schwenk and Miltenberger (2024) and Schwenk et al. (2025), we have investigated the modeled transport of moisture to the upper troposphere / lower stratosphere (UTLS) by warm conveyor belts (WCBs) and produced many interesting results. In a high resolution ICON simulation of a WCB case study, we shed light on the role of convectively ascending airstreams and highlighted the notable differences to their gradually ascending counterparts with respect to the vertical transport of vapor and hydrometeors, as well as to microphysical processes experienced during the ascent (Schwenk and Miltenberger, 2024). We compared this high resolution simulation to another simulation of the same WCB case, only with a lower resolution, and found substantial differences in UTLS moisture that were shown to have a noticeable effect the radiative budget at the top of the atmosphere (Schwenk and Miltenberger, 2025a, b). Finally, we examined multiple ICON simulations of another WCB case study with different parameter configurations and found that UTLS moisture in

the WCB outflow is highly sensitive to important microphysical parameters, such as ice capacitance and ice nucleating particle concentrations (Schwenk et al., 2025). At the risk of breaking the tacit expectation in the scientific community that scientific writing must always be conveyed with sobriety, we enthusiastically highlight our satisfaction with the novelty and scope of these results, which substantially address long-standing uncertainties and knowledge gaps in UTLS and WCB research.

However, while following our analysis and the conclusions derived therefrom, the attentive reader may have found themselves generally questioning the overall validity of our statements. After all, almost all the data we have examined up to this point in the thesis have been the ICON model's representation of reality, as opposed to reality itself. It would be valid for us to point out that this is a common approach in atmospheric sciences: results of model analyses are assumed to be our "best guess" of reality and therefore presented as representative of it. Most modeling scientists will be familiar with this assumption and usually, caveats are formulated in the conclusion to emphasize it clearly. The readability of any text would also be greatly diminished if every sentence were reformulated as follows: "The outflow contains more ice at 200 hPa" becomes "In the model, the outflow contains more modeled ice at the similarly modeled and uncertain 200 hPa level".

Nevertheless, in our experience, when this assumption is implied too often and one reads multiple modeling studies, the distinction between reality and the model representation thereof becomes blurred subliminally. Additional common assumptions, such as the natural superiority of models with higher spatial and temporal resolutions or more complicated, computationally expensive parameterization schemes exacerbate this issue, leading to a confused layering of assumptions. As a result, the findings of modeling studies are often presented, reused, cited, and treated as truth without emphasizing the crucial assumptions made throughout. This leads to exasperation among experimental scientists, who shake their heads in irritation and mutter, "It's only a model!"

Experimental scientist will be correct in pointing out that conclusions derived from model studies, specifically pertaining to the superiority of one model (configuration) over another, should be substantiated by measurements. To this, a modeling scientist might counter: "Please direct me to the nearest database which contains reliable, accurate, as well as spatially and temporally comprehensive measurement data of in-cloud humidity, ice mass, shape and size at 200 hPa over the Northern Atlantic ocean on the 23rd of September, 2017." Both scientists would probably laugh to the faint sound of decreased funding and rejected proposals, after which the modeler would add: "Even if this data were at hand, our model does not necessarily produce comparable data to flight measurements. The ICON grid-box variables for ice represent spatially averaged ice crystal size and mass distributions, while a measurement flight will invariably make point measurements of individual ice crystals in four dimensional space-time. Model uncertainties also introduce shifts in time and space, making direct comparisons difficult if not impossible. We would need a spatially averaged ice budget in the outflow of the WCB, or something similar, to verify our results." At this point, an atmospheric scientist working with satellite data would join the conversation and kindly point out that geostationary satellite data might be the best option for such an endeavor, given the high spatial and temporal coverage in a region of the atmosphere that is otherwise notoriously difficult to measure. One would, of course, still have to work around the problem of model shifts as well as satellite retrieval algorithms with large uncertainties. Such a hypothetical conversation could have taken place during a coffee break at the 2024 International Conference on Clouds and Precipitation on Jeju island, South Korea.

To satisfy attentive readers and frustrated experimental and modeling scientists, we developed a method to compare numerical weather prediction (NWP) model output for the upper troposphere

to geostationary satellite measurements. While not all elements of this method are novel (similar work has been done by Freischem et al. (2024)), to our knowledge, it has not been implemented in this fashion before. We translate NWP model output into synthetic satellite data using a radiative transfer model, and analyze both simulated and measured geostationary satellite products using fractal analysis and object tracking. The goal of this method is to expand the range of methods available for objectively and robustly verifying model results or validating the superiority of one model over another. We present a limited use case relevant to the context of Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b), but we are confident that this method can be expanded to other regions of the atmosphere and to other atmospheric phenomena besides WCBs.

2 Introduction

Since the emergence of numerical weather prediction (NWP) models, their output has been compared to observations for validation and to identify shortcomings (Charney, 1949; Charney et al., 1950; Shuman, 1989). A plethora of methods have been developed for this purpose, and they range from simple, point-wise statistics, such as root-mean-square error (RMSE), to more sophisticated approaches, such as object-based diagnostics and continuous ranked probability scores (CRPS). There are common problems that arise for many methods, such as the representation error, which comes from the fact that observations often represent a very local point in spacetime, whereas model grid values are an area-average over a larger cell and a time-average over the output interval (Janjić et al., 2017). The full scope of these methods and their shortcomings can only be adequately summarized in entire books, such as Jolliffe and Stephenson (2011). However, we can summarize that most of these methods have been developed and employed for operational weather forecasts and tropospheric variables, which have a more direct impact on the ground-dwelling human civilization than upper atmospheric phenomena. Therefore, there has been much less effort in comparing NWP model output to upper tropospheric variables such as cloud top height, cirrus occurrence or relative humidity using satellite or radiosonde data. Nevertheless, recent studies have explored this possibility. For example, Freischem et al. (2024) used multi-fractal analysis to compare outgoing longwave radiation (OLR) at the top of the atmosphere, as measured by satellites, with NextGEMS km-scale NWP model data (Segura et al., 2025). Carminati et al. (2019) and Newman et al. (2020) used global NWP models, as well as radiosonde data, to validate and calibrate satellite measurements. To properly compare NWP model output to satellite products, the latter two studies employed radiative transfer models to translate NWP model output into synthetic satellite images.

Radiative transfer models (RTMs) are models whose goal it is to solve the radiative transfer equation (RTE) given relevant atmospheric variables and surface conditions (Eyre, 1991; Vidot, 2015). One of the most widely used RTMs is the radiative transfer model for TOVS (RTTOV), which was developed by the European Centre for Medium-Range Weather Forecasts (ECMWF) (Vidot, 2015; Saunders et al., 2018). For the sake of completeness, we clarify that TOVS is an acronym for *TIROS operational vertical sounder*, and that TIROS is an acronym for *television infrared observation satellites*. The RTTOV model has been used for a variety of applications, such as enabling physical retrievals from satellite data (Li et al., 2000), proposing which instruments should fly on future satellites (Andrey-Andrés et al., 2018), and, crucially for this work, producing simulated imagery from numerical weather prediction (NWP) models (Saunders et al., 2018).

RTTOV can simulate the output of various satellite instruments, including passive infrared (IR) and passive microwave (MW) sounders and visible (VIS) regions of the electromagnetic spectrum (Vidot, 2015). Generally, the output depicts the brightness temperature (BT) of a certain channel, such

as the IR $39\mu\text{m}$ channel from the Spinning Enhanced Visible and InfraRed Imager (SEVIRI) on the Meteosat second generation (MSG) satellite. Translating NWP model output into synthetic satellite data is generally viewed as more reliable than translating satellite products into comparable NWP data using retrieval algorithms due to the difficulty of objectively quantifying the error in satellite retrievals (Thépaut, 2006). The method we propose in this paper aims to compare synthetic satellite data, obtained from NWP model output using RTTOV, with geostationary satellite measurements. To this end, we use object tracking and fractal analysis, which have two advantages: they do not introduce substantial errors from small shifts in time and space between simulations and measurements, and they compare statistical metrics instead of point-by-point values.

In atmospheric science, "object tracking" is the process of identifying regions of interest in two or three dimensions using satellite, radar, or model data and estimating their trajectories over time with a sequence of images (Raut et al., 2021). Since the 1970s, various techniques have been developed to track thunderstorms and convective clouds (Leese et al., 1971; Crane, 1979; Reid, 1979). Currently, there are multiple simple as well as complex tracking algorithms designed for specific applications and types of data. Examples include the Thunderstorm Identification, Tracking, Analysis, and Nowcasting (TITAN) software (Dixon and Wiener, 1993), which uses radar reflectivity thresholds to identify thunderstorms; the Tracking Algorithm for Mesoscale Convective Systems (TAMS, Núñez Ocasio et al. (2020)), which combines the area-overlapping and Lagrangian centroid projection techniques to track objects from satellite data; and the fully automated Tracking of Organized Convection (TOOCAN) algorithm (Fioleau and Roca, 2013), which uses clustering methods to track cloud clusters over time from satellite data. Another example noteworthy for its sophistication is the "Grab 'Em, Tag 'Em, Graph 'Em" (GTG) algorithm (Whitehall et al., 2014), which uses graph theory to track cloud clusters from satellite data. While these algorithms and techniques have proven useful, they remain specialized to various data types and target variables. This shortcoming led to the development of the Tracking and Object-Based Analysis of Clouds (TOBAC, Sokolowsky et al. (2024)) tool, which was designed to provide an efficient, multipurpose tracking algorithm that is both variable- and grid-agnostic and is used in this study.

In addition to object tracking we employ fractal analysis in our proposed method. Fractals are self-similar geometric structures with non-integer "dimensions" that describe how their complexity changes with scale (Feder, 1988). Early on, the field of fractal research recognized that fractals are omnipresent in the atmosphere; in his book "Fractals", Feder (1988) lists cumulus clouds as one of the first examples of fractals in nature, stating that they "consist of big heaps with smaller bulges that have smaller bumps with bumps on them and so on down to the smallest scale you can resolve." Therefore, many studies have investigated the fractal characteristics of clouds and storms (Cahalan and Joseph, 1989; Cahalan et al., 1994; Siebesma and Jonker, 2000; Lovejoy and Mandelbrot, 1985) as well as other atmospheric phenomena such as turbulence (Sreenivasan and Meneveau, 1986; Rees et al., 2024) and ice supersaturated regions (Schuh et al., 2025). Apart from pure scientific curiosity, fractal analysis has proven useful for various applications, such as the characterization of cloud types (Batista-Tomás et al., 2015) or in highlighting shortcomings of simulations with respect to observations (Freischem et al., 2024).

This paper serves as proof of concept for our proposed method, which we apply to a limited case study of a warm conveyor belt (WCB). WCBs are large and coherent ascending airstreams that form within extratropical cyclones (Madonna et al., 2014) and they play a large role in transporting hydrometeors, water vapor, aerosols, and trace gases into the upper troposphere / lower stratosphere (UTLS) (Zahn et al., 2014; Schwenk and Miltenberger, 2024). The clouds produced during the as-

cent interact with incoming shortwave and outgoing longwave radiation (Joos, 2019), and the water vapor that is injected into the UTLS acts as a strong greenhouse gas (Schneider et al., 2010).

The WCB case study investigated in this paper is the same case examined in Schwenk and Miltenberger (2025a), Schwenk and Miltenberger (2025b), and Schwenk and Miltenberger (2024). In the former two studies, we found that a WCB simulation with a high, convection-permitting resolution produced very different UTLS outflow conditions than a simulation with a lower resolution and convection parameterization. Although the higher resolution simulation had larger upper-level ice number mixing ratios, it showed smaller relative and specific humidities, leading to smaller OLR. Additionally, we identified potential limitations in how the lower-resolution model represents water vapor and ice transport through the convection parameterization scheme, suggesting that operational weather models and global climate models (GCMs) may overestimate water vapor and underestimate ice content in the outflow of WCBs. This latter conclusion hinges on the assumption that the higher resolution simulation provides data that is more accurate and realistic. While it is generally accepted that increasing NWP model resolution from tens of kilometers to a few kilometers improves model performance (Prein et al., 2015; Clark et al., 2016; Weusthoff et al., 2010), studies have shown that large biases can remain, especially when modeling convection (Prein et al., 2025). The extent of any improvements in the representation of upper tropospheric variables also remains unclear.

We write this paper to (i) challenge the crucial assumption of the superiority of a higher-resolution model and (ii) possibly lend more credibility to the conclusions we formulate in Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b). To our knowledge, our method has not been used in this way before, and we hope our work inspires others to adapt and improve upon it using larger, more varied datasets.

3 Methods

In this section, we briefly discuss the NWP model configurations used for simulating the WCB case study. We then provide an overview of the RTTOV model, fractal analysis, object tracking tools, and satellite data that we use.

3.1 Case study, ICON model setup

The same case study is examined as in Schwenk and Miltenberger (2024, 2025a, b). The extratropical cyclone that gave rise to our WCB case study formed off the Eastern coast of Canada on 21 September 2017, and by the 23 September 2017, a clear WCB cloud band formed over the Northern Atlantic. It made landfall over Iceland and began to dissipate over Northern Europe early on the 24 September 2017. For more details on the WCB case, see Schwenk and Miltenberger (2024). The Icosahedral Nonhydrostatic (ICON) model (v2.6.2; Zängl et al. (2014)) was used to conduct the two simulations, which were initialized with operational ICON global analysis at 00:00 UTC on 20 September 2017. Both simulations ran for 96 hours until 00:00 UTC on 24 September 2017. The simulations in this study were run with a slightly different configuration than previously in order to accommodate the RTTOV model, which requires model output for very high regions of the atmosphere. However, the only difference is the number of model levels, which was increased from 90 to 115 model levels for both simulations.

The higher resolution, convection permitting simulation is referred to as the "nested" simulation. This model configuration included two higher-resolution domains running on a R03B08 grid (~ 6.5 km resolution) and a R03B09 grid (~ 3.3 km resolution), nested into the global domain running on a

R03B07 grid (~ 13 km resolution). Two-way coupling allows neighboring grids to interact with each other (Zängl et al., 2022). The highest-resolution domain resolves (deep and embedded) convection, while the global domain parameterizes deep convection using the Tiedtke–Bechtold scheme (Tiedtke, 1989; Bechtold et al., 2008). Sub-grid processes are parameterized by the standard ICON schemes. Cloud microphysical processes are parameterized by a two-moment scheme from Seifert and Beheng (2005) which represents six hydrometeor species (cloud droplets, rain, ice, snow, graupel, and hail). The lower resolution simulation runs on only the global domain defined above, where deep convection is parameterized by the Tiedtke–Bechtold scheme.

In this study, we focus on the area defined by the third domain, since this is where differences between the two models due to their different resolutions are most apparent. Additionally, we only investigate the time during which the WCB is mostly situated within this domain and experiences the largest amount of ascending motion. In Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b) we call define this time as time-frame 2 (TF2), which begins on 00:00 UTC 2017-09-22, and lasts to 06:00 UTC 2017-09-23.

3.2 Satellite radiance simulation using RTTOV

Top-of-atmosphere radiances for the geostationary Meteosat-Second-Generation Spinning Enhanced Visible and InfraRed Imager (MSG-SEVIRI) thermal channels were computed using the standard RTTOV model configuration (Saunders et al., 2018) and the most recent version v14.0 released in February 2025 (EUMETSAT, 2025). The RTTOV model coefficients for SEVIRI are provided by the EUMETSAT NWP Satellite Application Facility (SAF). For details on the RTTOV model, we direct the reader to Saunders et al. (2002) and Saunders et al. (2018). We adapted a Python wrapper written by (Kugler, 2025) for the WRF model to work with ICON model output. We simulated eight MSG-SEVIRI infrared and water-vapour channels: WV_062, WV_073, IR_039, IR_087, IR_097, IR_108, IR_120, and IR_134 (Fig. 1). To ensure an apples-to-apples comparison, the MSG-SEVIRI data is interpolated to the output longitude-latitude grids of the nested and global simulations, respectively. Thus, RTTOV output from the respective models is compared to satellite data of the same resolution. However, this does not account for the fact that a model with a certain grid spacing always requires approximately seven grid cells to fully represent features such as convective clouds. In contrast, MSG-SEVIRI data simply downscales reality, resolving everything to a coarser grid.

We chose to focus our analysis on three "products" that are each formed by combining two MSG-SEVIRI channels. A greater variety of more sophisticated products are not investigated within the scope of this study. We define the cloud top height ("height") product as the brightness-temperature difference (BTD) between the SEVIRI water-vapour channels at $6.2\ \mu\text{m}$ and $7.3\ \mu\text{m}$ (WV_062–WV_073; see Fig. 2 c), which exploits their differential absorption to enhance overshooting tops and high-level moist features (NOAA/NESDIS/STAR (2018)). The cloud phase ("phase") product discriminates the cloud phase using the BTD between the thermal-window ($10.8\ \mu\text{m}$) and the $8.7\ \mu\text{m}$ SEVIRI channels (IR_108–IR_087; see Fig. 2 b), and is a well-established technique for separating liquid-water from ice clouds (EUMeTrain, 2021). Finally, the cloud particle size ("size") product is inferred from the BTD between the shortwave IR ($3.9\ \mu\text{m}$) and the thermal-window ($10.8\ \mu\text{m}$) channels (IR_039–IR_108), leveraging the high reflectance sensitivity of small hydrometeors at $3.9\ \mu\text{m}$ (NOAA/NESDIS/STAR, 2018).

For each product we identify threshold values by eye that are used to mask the data (Fig. 2). For RTTOV output, which is slightly different from the MSG-SEVIRI data, the thresholds are BTD values of $-7\ \text{K}$, $-0.5\ \text{K}$, and $4\ \text{K}$, for the height, phase and size products, respectively (Fig. 2). For the

MSG-SEVIRI data, these thresholds are -4 K, -0.5 K, and 6 K. The threshold targets for TOBAC are designated as maximum, minimum and maximum for the height, phase and size, respectively (i.e., is the area within the contour of the threshold a local maximum or minimum? See below for details). The binary, masked fields are then used in the fractal analysis and object tracking. We do not use multiple or more rigorously determined thresholds nor do we conduct multi-fractal analysis within the scope of this study, since we present a limited use-case as a proof of concept.

3.3 Fractal analysis

A common method for fractal analysis (in two dimensions) is to determine the fractal dimension (α) of an object using its logarithmic area (A) to perimeter (P) ratio (Lovejoy, 1982; Batista-Tomás et al., 2015; Schuh et al., 2025):

$$\alpha = \frac{\log(A)}{\log(P)}. \quad (1)$$

This equation stems from the assumption that the relationship between the area A and the perimeter P is described by a power law, with the exponent α . Smaller values for α indicate more complex fractal behavior. The theoretical limits are objects with (i) zero area but finite perimeter; (ii) non-zero area but infinite perimeter; and (iii) zero area and infinite perimeter, all of which produce maximally self-similar objects with $\alpha = 0$.

In our analysis, we produce masked fields of our products (i.e white contours in Fig. 3) and the fractal dimension is calculated for each individual object (closed contour) that covers 10 grid cells or more. We calculate the real perimeter and area in units of m^2 from latitude-longitude grid-spacings (Δx and Δy , in radians) using basic spherical geometry:

$$A^{\text{grid cell}}(\text{lat}, \Delta x, \Delta y) \approx \underbrace{R \cdot \Delta y}_{\Delta y_{\text{meters}}} \cdot \underbrace{R \cdot \cos(\text{lat}) \cdot \Delta x}_{\Delta x_{\text{meters}}(\text{lat})}, \quad (2)$$

With R the radius of Earth. The perimeter is calculated by adding all the sides (Δy_{meters}) and tops ($\Delta x_{\text{meters}}(\text{lat})$) of a grid cell within an object if it has no neighbors on the respective sides or top/bottom (similarly to Schuh et al. (2025)). At each time step, the mean fractal dimension is calculated for all objects within domain 3. In this way, we can observe the evolution of fractal characteristics over time.

3.4 Object tracking with TOBAC

Object tracking for the height, phase, and size products was performed using the Tracking and Object-Based Analysis of Clouds (TOBAC) Python package v1.6.0. TOBAC detects features as local maxima or minima within two- or three-dimensional gridded fields, using threshold-based segmentation routines to delineate individual objects. TOBAC then links objects across time steps by matching centroids and shape characteristics to construct temporally consistent trajectories (tobac development team, 2025). We investigate the lifetimes of these trajectories or "tracks", which provides comparable information on the temporal evolution of our three products between observations and simulations.

4 Qualitative analysis of satellite products

The MSG-SEVIRI products modeled by RTTOV in domain 3 during TF2 for the global and the nested simulation have nearly identical distributions and show only minor differences (Fig. 1). There

is a shift in the peak WV_073 brightness temperature (Fig. 1 b), which is located at a higher value of approximately 247 K for the global simulation, versus approximately 242 K for the nested simulation. Since this channel is used for water vapor retrievals, this could reflect the different UTLS vapor conditions identified in Schwenk and Miltenberger (2025b). For all infrared products, and counting alphabetically from Figures 1 c) to h), the bump or mode at approximately 280 K, 245 K, 280 K, 280 K, 255 K, and 275 K is slightly more pronounced in the global simulation than in the nested simulation (but not shifted). These channels are sensitive to many different variables and could reflect the differences in upper tropospheric hydrometeor content determined in Schwenk and Miltenberger (2025b). However, in contrast to the methods we used in Schwenk and Miltenberger (2025b), we did not create WCB masks to isolate the WCB region, meaning that we are looking at the entirety of domain 3 which contains other weather systems and clouds than the WCB.

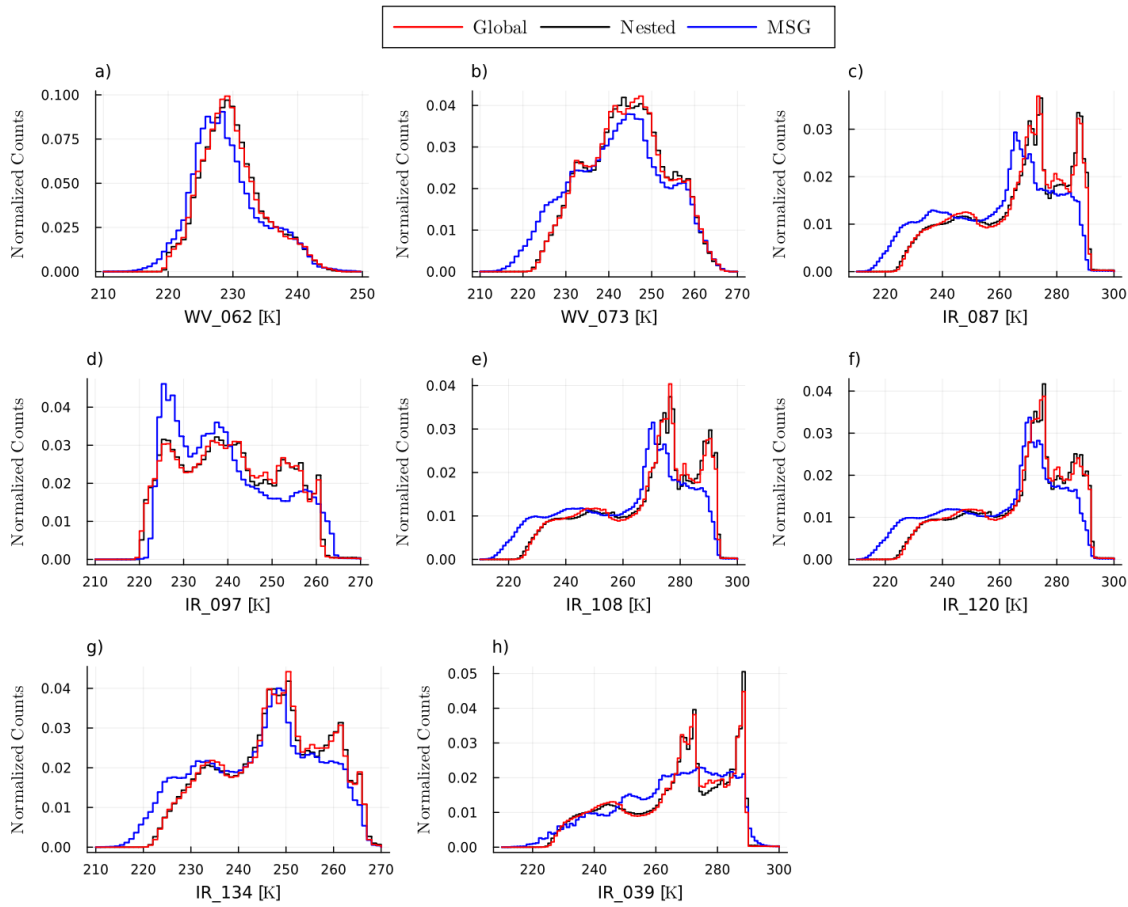


Figure 1. Normalized histograms of the measured (blue) and simulated (red for global simulation, black for nested simulation) MSG-SEVIRI channels during TF2 (00:00 UTC 2017-09-22 to 06:00 UTC 2017-09-23) and in domain 3.

While the modeled brightness temperatures are very similar for both simulations, their distributed are very different to the measured values (Fig. 1). The most similar is the WV_062 channel (Fig 1 a), which merely shows a shifted peak value of approximately 225 K for the measurements, as opposed to approximately 230 K for the simulations. For WV_073, the peaks are located in similar positions but the measurements have a larger tail to lower brightness temperatures (Fig 1 b). For IR_087, IR_108, IR_120 and IR_134 (Fig 1 c, e, f and g), the distributions for measurements are generally shifted to lower brightness temperatures, and the pronounced second mode at larger brightness temperatures seen for the simulations is missing.

The IR_097 distribution for the measurements (Fig 1 d) shows more pronounced peaks at lower brightness temperatures and less pronounced peaks at higher brightness temperatures than the simulations. Finally, the IR_039 distribution (Fig 1 h) has none of the pronounced peaks that the distributions for the simulations have. In the scope of this paper, we do not delve into the complexities of RTTOV simulations or examine possibilities for these large differences. For a discussion of these topics, we guide the interested reader to the RTTOV v14 technical report (Hocking et al., 2025) and to the NWP SAF website that lists all RTTOV-related publications (<https://nwp-saf.eumetsat.int/site/software/rttov/all-rttov-related-publications/>). In this paper, we instead aim to use three products derived from these channels and find adequate threshold values that allow for an objective (or as objective as possible) comparison between the simulations and the measurements.

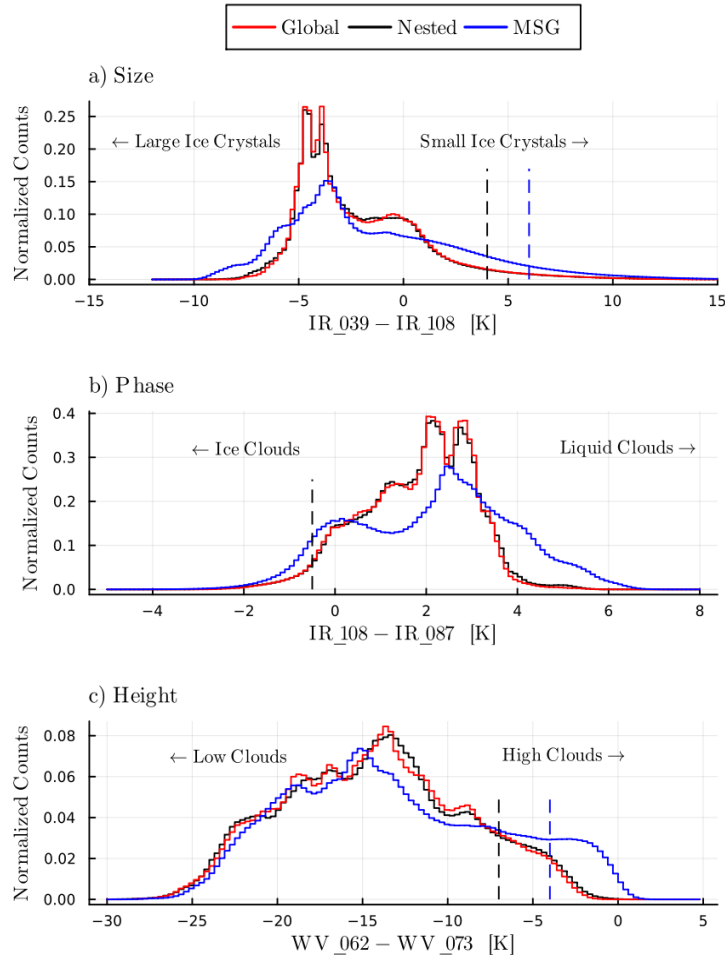


Figure 2. Normalized histograms of the measured (blue) and simulated (red for global simulation, black for nested simulation) MSG-SEVIRI products during TF2 and in domain 3 for size (a), phase (b) and height (c). Dashed vertical lines indicate the thresholds used for RTTOV simulations (black) and measurements (blue).

The three satellite products we define (size, phase and height, see Section 3.2) are very similar for the two RTTOV simulations (Fig. 2). Only for the height (Fig. 2 c) some modes are slightly more pronounced in the global than in the nested simulation. The distributions are again very different to the measured values, which have longer tails to the sides and shifted peak values. This is one of the reasons we chose only one threshold value per product and type (RTTOV and measurement). We chose the respective thresholds (Section 3.2) by examining the distributions (Fig. 2) and heatmaps

using contours to indicate threshold values (as in Fig. 9 for size, Fig. 3 for phase, and Fig. 6 for height. Note: there are two sub-figures for measurements, since we re-gridded them to the respective resolutions of the global and nested simulations). For the size product, we chose threshold values that would isolate objects where ice crystals are small; for the phase we chose them to isolate areas which are ice clouds; for the height, we chose them to isolate areas of high clouds. We note that the size product can only be used by night, since during the day the reflected sunlight saturates the image (Fig. 10 c and d). Therefore, the useful amount of data for this product is very limited.

When examining the geographical distributions (i.e heatmaps) of the three products at three different times during TF2 (Fig. 9, 10 and 11 for size; Fig. 3, 4 and 5 for phase; Fig. 6, 7 and 8 for height), we find that both the global and the nested simulations produce similar output. One notable exception is the height product during the main ascent phase at 07:30 UTC on 22 September 2017 (Fig. 7) a and b), where the nested simulation (Fig. 7) b) shows high-altitude cloud tops at the northern and north-eastern edge of the WCB that are mostly missing in the global simulation (Fig. 7) b), but which are present in the measurements (Fig. 7) c and d). Possibly, the convection parameterization is producing too little convection in this area, or alternatively, the parameterization fails to adequately represent cloud tops when triggered. With this difference in mind, it is possible to interpret the phase product at the same time (Fig. 4) similarly, since enclosed objects indicating the ice-phase are also slightly more abundant in the nested simulation at the northern edge of the WCB than in the global simulation.

The three products are similar for the nested and global simulation, and thus also show the same strengths and weaknesses when compared to observations. In the beginning of TF2 at 0:00 UTC on 22 September 2017 (Figs. 3, 6, 9), all three products are skillfully represented by both simulations, at least visually. However, this skill declines sharply with time. At 07:30 UTC on 22 September 2017, the simulations represent far fewer objects enclosed by the thresholds values for the phase (Fig. 4), and the height product becomes very dissimilar in the southern part of the WCB (Fig. 7). This latter issue is most prominent for all three products at 06:00 UTC on 23 September 2017 (Figs. 5, 8, 11), where both simulations grossly under-represent the amount and size of enclosed object in the progressing cold front. A full interpretation of the ICON model's shortcomings in this regard is not the aim of this paper, but we point out that this misrepresentation is very clear and also not fixed by an increase in model resolution. It is also a clear sign that our assumption, that the results from Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b) from the nested simulation represent reality better, should be treated with caution.

5 Fractal analysis and object tracking of satellite products

In domain 3, across TF2 and for all three products, the average fractal dimension α is higher in the global simulation than in the nested simulation and the MSG-SEVIRI measurements (Fig. 12 a, c and e), which indicates a lower degree of self-similarity or complexity of the shapes produced in the global simulation. Notably, this finding remains true even when comparing α to the MSG-SEVIRI measurements that are re-gridded to the low resolution of the global simulations. We only include objects in the analysis that include 10 grid cells or more, but any bias resulting from this should also be present in the re-gridded MSG data. Instead, this lower degree of complexity could be due to the fact that at any grid-resolution, multiple cells are required to fully resolve regions of the atmosphere, which leads to "smoothed" edges. For the nested simulation, α values are much more similar to MSG-SEVIRI measurements, and are only slightly larger for the phase (Fig. 12 c). This similarity suggests that the grid resolution of the nested simulation adequately represents the system's complexity at our observed scale.

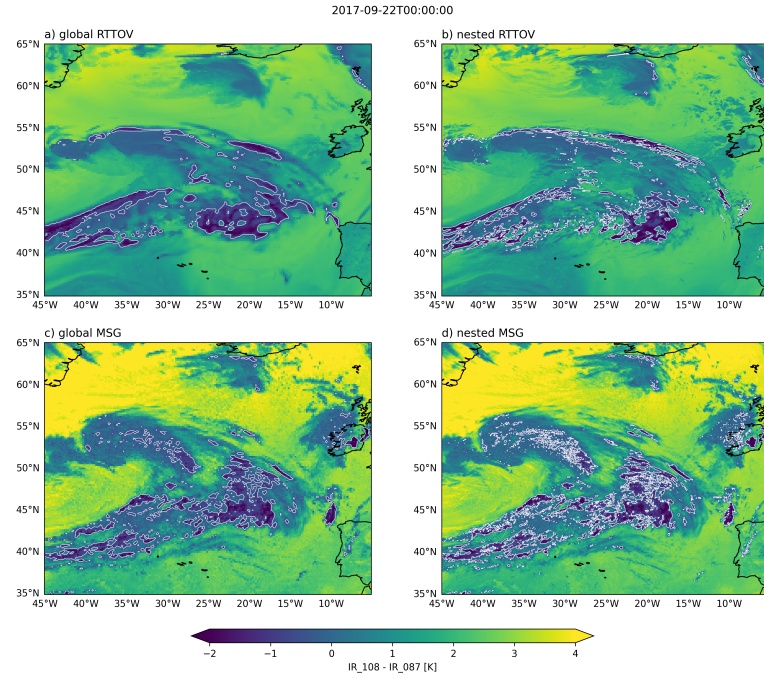


Figure 3. MSG-SEVIRI products for the phase (IR_108 - IR_087) on 00:00 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regridded from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in white along the threshold BTD value of -0.5 K for both measurements and RTTOV simulations.

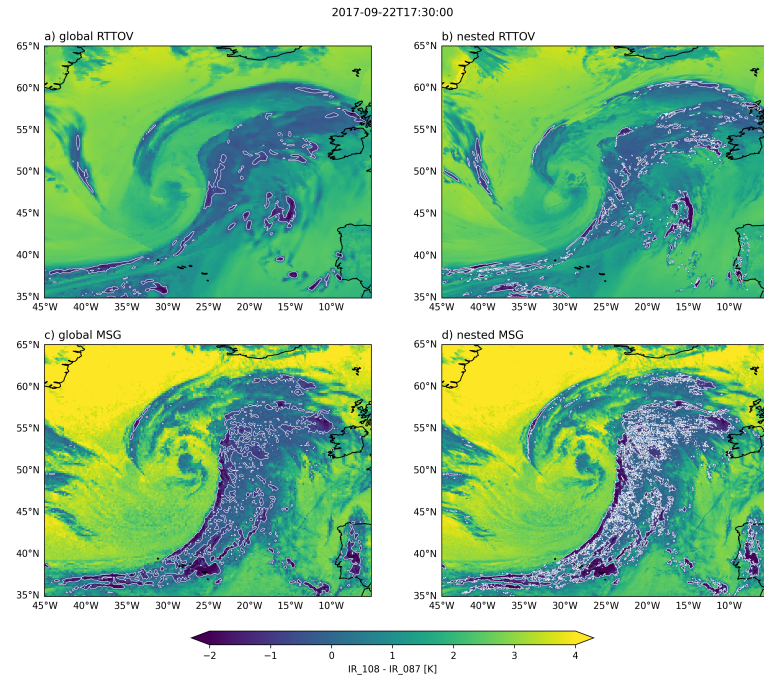


Figure 4. MSG-SEVIRI products for the phase (IR_108 - IR_087) on 17:30 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regridded from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in white along the threshold BTD value of -0.5 K for both measurements and RTTOV simulations.

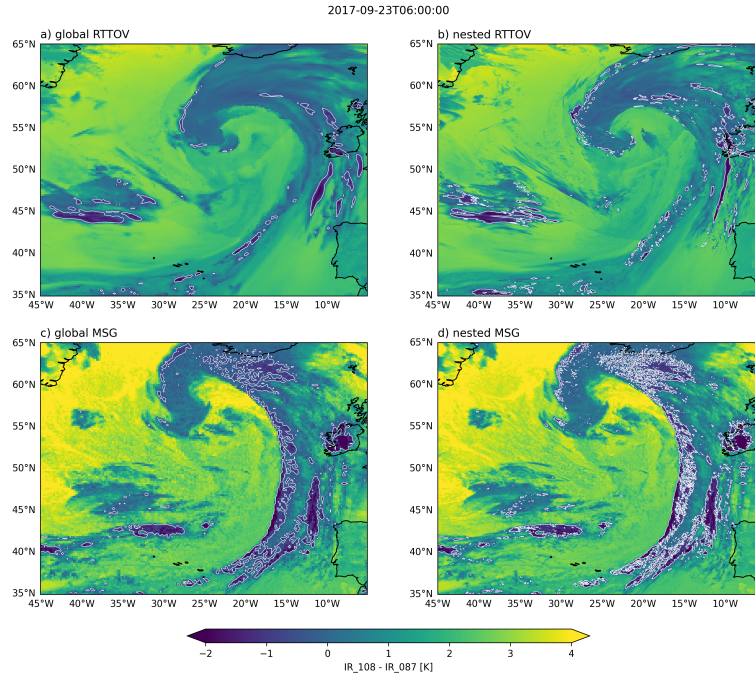


Figure 5. MSG-SEVIRI products for the phase (IR_108 - IR_087) on 06:00 UTC on 23 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regrided from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in white along the threshold BTD value of -0.5 K for both measurements and RTTOV simulations.

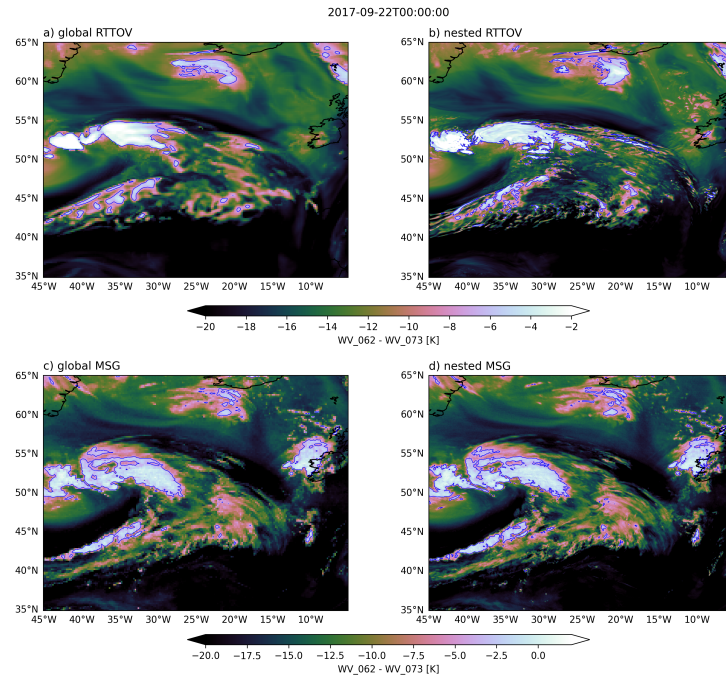


Figure 6. MSG-SEVIRI products for the height (WV_062 - WV_073) on 00:00 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regrided from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in blue along the threshold BTD values of -7 K for RTTOV simulations and -4 K for measurements.

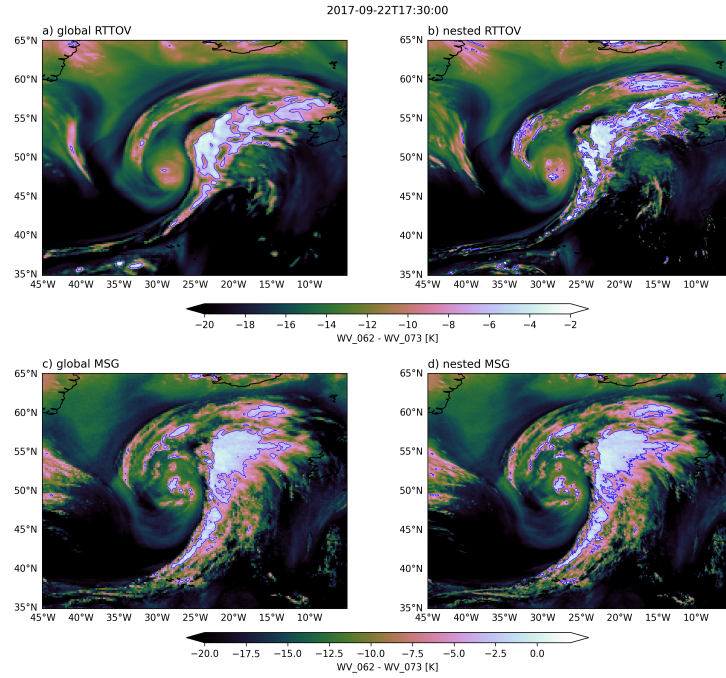


Figure 7. MSG-SEVIRI products for the height (WV_062 - WV_073) on 17:30 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regrided from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in blue along the threshold BTD values of -7 K for RTTOV simulations and -4 K for measurements.

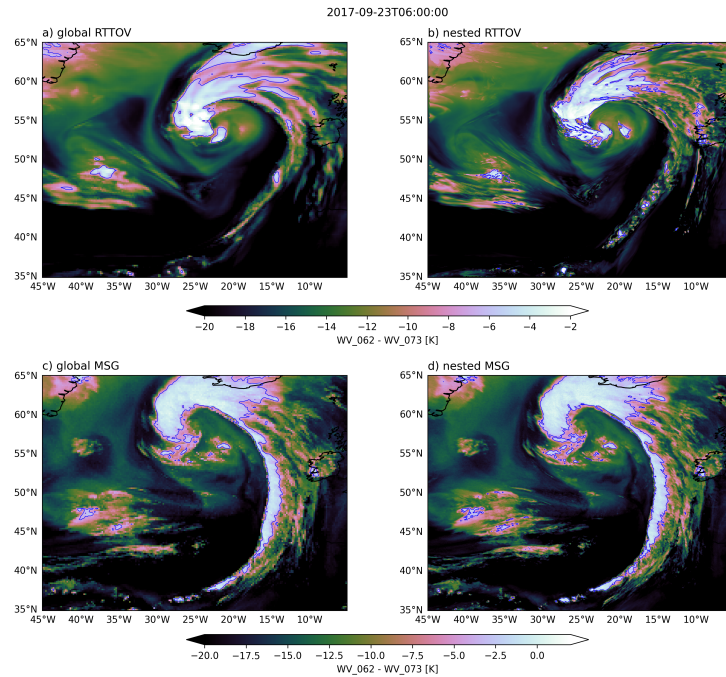


Figure 8. MSG-SEVIRI products for the height (WV_062 - WV_073) on 06:00 UTC on 23 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regrided from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in blue along the threshold BTD values of -7 K for RTTOV simulations and -4 K for measurements.

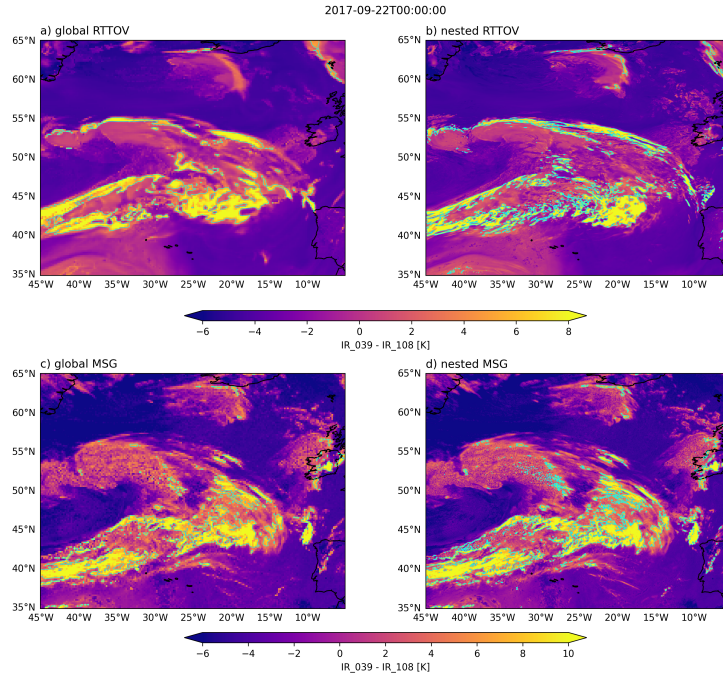


Figure 9. MSG-SEVIRI products for the size (IR₀₃₉ - IR₁₀₈) on 00:00 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regridded from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in cyan along the threshold BTD values of 4 K for RTTOV simulations and 6 K for measurements.

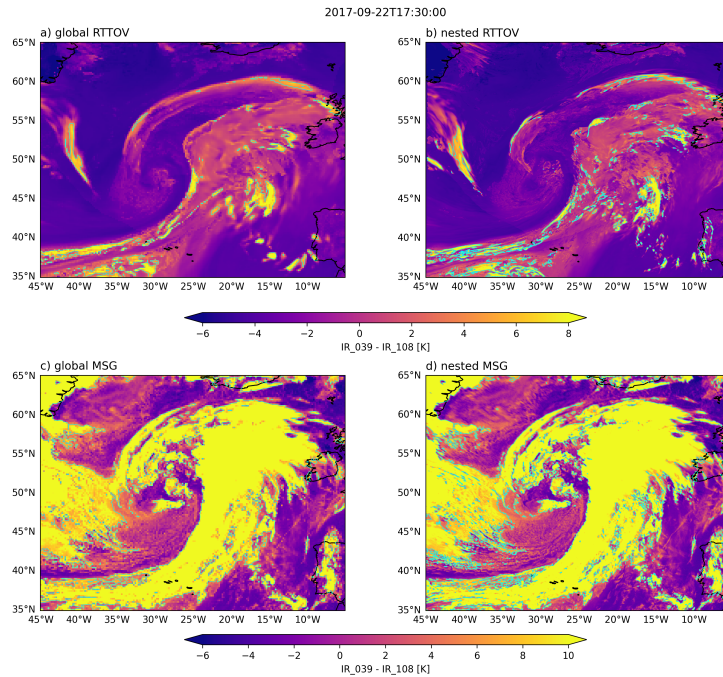


Figure 10. MSG-SEVIRI products for the size (IR₀₃₉ - IR₁₀₈) on 17:30 UTC on 22 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regridded from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in cyan along the threshold BTD values of 4 K for RTTOV simulations and 6 K for measurements.

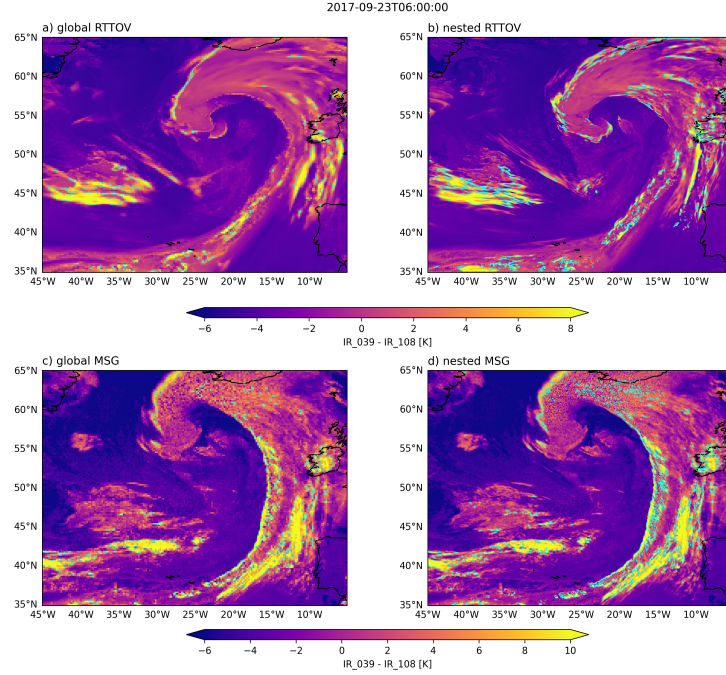


Figure 11. MSG-SEVIRI products for the size (IR_039 - IR_108) on 06:00 UTC on 23 September 2017, simulated using RTTOV for the global (a) and nested (b) simulation, and regridded from MSG observations to the global grid (c) and nested grid (d). Contours are plotted in cyan along the threshold BTD values of 4 K for RTTOV simulations and 6 K for measurements.

The temporal evolution of α is sporadic, with no clear patterns visible (Fig. 12 a, c and e). At the end of TF2, the average fractal dimension of the objects for all products is nearly identical for the nested simulation and the MSG-SEVIRI measurements, despite the poor representation of the cold front (Figs. 5, 8, 11). This suggests that, even when the nested simulation produces fewer objects for a given product, the fractal characteristics of the represented objects remain realistic. We can summarize that when analyzing the fractal characteristics of height, phase, and size products, the nested simulation produces objects indicating regions of small ice crystals and high-altitude cloud tops whose shapes are highly realistic. In contrast, these objects show too small a degree of self-similarity in the global simulation.

When examining the area of enclosed objects from the three products (during TF2 and in domain 3), we find that both the nested and global simulation produce more objects with a larger area than the MSG-SEVIRI observations (Fig. 12 b, d and f). For the height, the nested simulation produces objects with an median area of $17.1 \cdot 10^7 \text{ m}^2$, compared to $8.5 \cdot 10^7 \text{ m}^2$ for the re-gridded measurements. For the global simulation, this value is $15.8 \cdot 10^8 \text{ m}^2$ versus $5.7 \cdot 10^8 \text{ m}^2$. These differences are similar for the phase and size also. The distributions (Fig. 12 b, d and f) clearly indicate that the observations show more and smaller objects than the simulations. The data supports this: for the phase product, the nested simulation has $5.6 \cdot 10^4$ objects, versus $14.6 \cdot 10^4$ objects in the re-gridded MSG-SEVIRI data. For the global simulation, this difference is $9.1 \cdot 10^3$ versus $23.5 \cdot 10^3$ objects. This difference is once more a reflection of the fact that simulations need multiple grid cells to resolve the atmosphere and that the measurements do not suffer from this problem.

TOBAC was only applied to the phase and height products for object tracking, since the size product is only available at night. This means that tracks that last for many hours cannot be resolved. For the height and the phase, we find that in both the simulations and MSG-SEVIRI observations, most tracked objects have short lifetimes of around 15-30 minutes (Fig. 13). This reflects the dynamic

nature of convection in a WCB and also the abundance of small, isolated regions that live for a short time (large objects are only counted once). Just as there are more individual objects in the observational data than in the simulations, there are overall also more tracks. The exception is the height and the nested simulation (Fig. 13 b), which has more tracks than the re-gridded MSG-SEVIRI data. Outliers with track lifetimes exceeding six hours only appear in the simulations, suggesting that, in reality, the WCB is more temporally dynamic with respect to high-altitude cloud tops and regions dominated by ice than the simulations indicate. However, the general distribution of track lifetimes is similar for the simulations and observations. These results suggest that overall, the simulations effectively capture the temporal dynamics of high-altitude cloud tops and ice-phase cloud tops in the WCB.

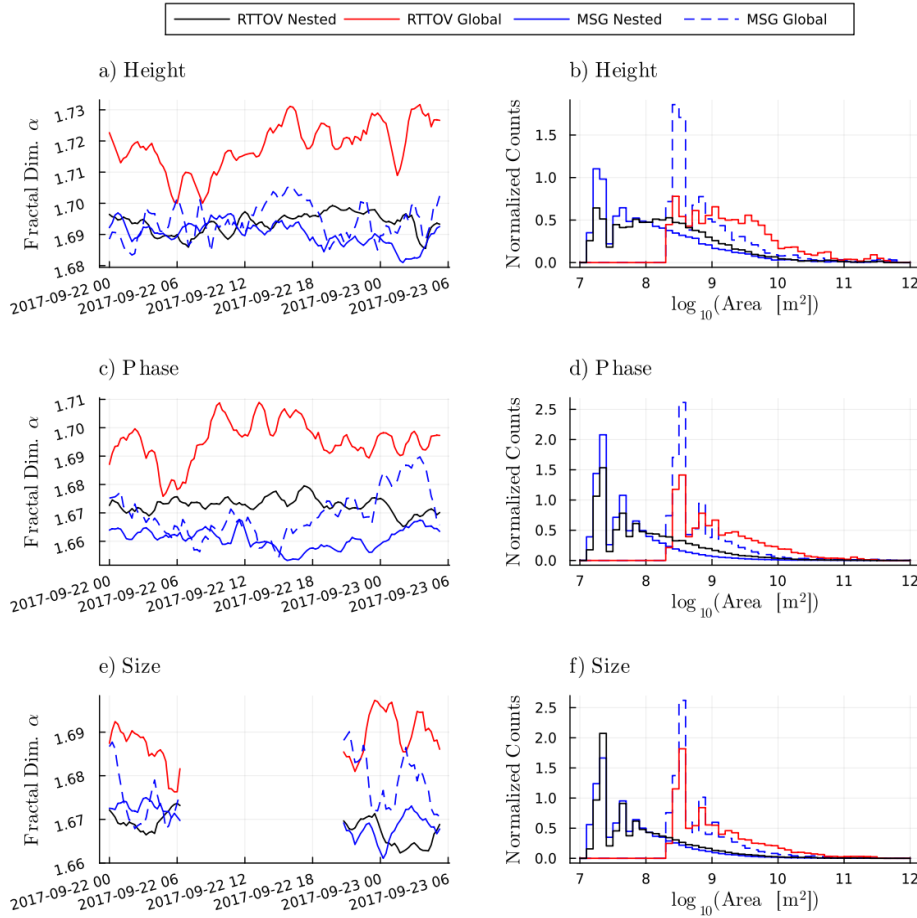


Figure 12. The average fractal dimensions (with a 45-minute running mean) of the enclosed objects are shown over time for the height product (a), phase (c), and size (e). The red lines indicate the global simulation, the black lines indicate the nested simulation, and the blue lines indicate the MSG-SEVIRI measurements regridded to the resolution of the nested simulation (solid line) and the global simulation (dashed line). Histograms showing the (logarithmic) area enclosed by each object across TF2 (with the exception of daytime for size) are presented for height (b), phase (d), and size (f), with the same color coding and line shading.

6 Conclusions and Discussion

This paper is the third part of a three-part study investigating the role of simulation scale in WCB moisture transport into the UTLS. While the first two parts (Schwenk and Miltenberger, 2025a, b) examined differences between simulations with different grid resolutions, this paper aimed to compare the simulations to observations. The goal was to develop a novel method for this task, apply it

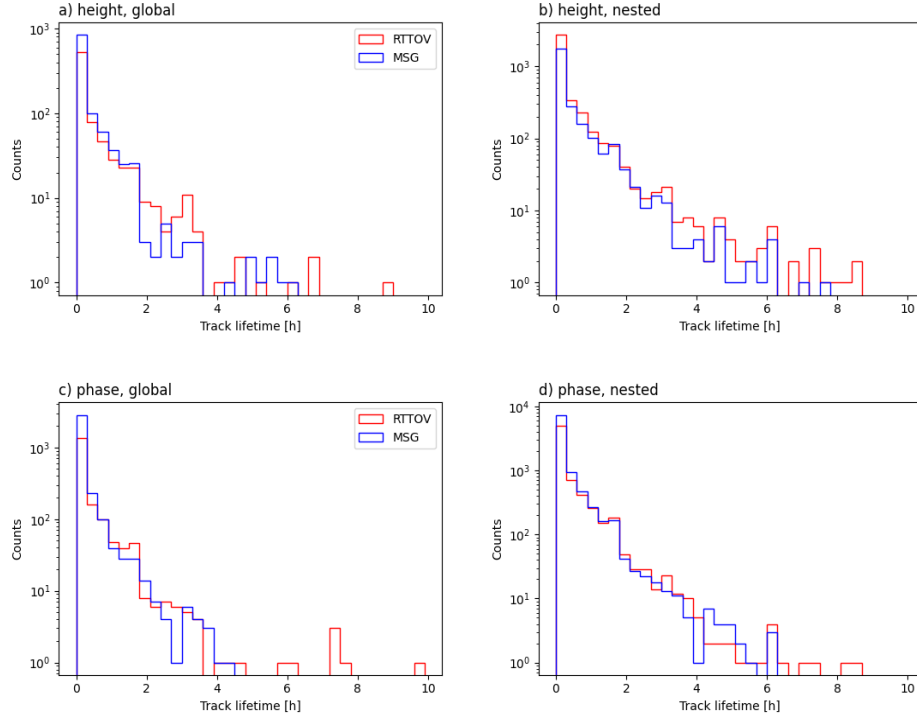


Figure 13. Histograms of track lifetimes for objects tracked by TOBAC in domain 3 during TF2 for the height (top row) and the phase (bottom row) products. Results for the global simulation and MSG-SEVIRI data re-gridded to the global simulations resolution are shown left, for the nested simulation right. Blue lines indicate MSG-SEVIRI data, red lines simulation data.

to our limited WCB case study, and determine if the higher-resolution simulation better represents reality than the lower-resolution simulation. If so, the conclusions from Schwenk and Miltenberger (2025a) and (Schwenk and Miltenberger, 2025b) would be more meaningful. Using the radiative transfer model for TOVS (RTTOV), we translated the simulation output from the higher resolution ("nested") and lower resolution ("global") ICON simulations into MSG-SEVIRI brightness temperatures. We then examined three products from these brightness temperatures, relating to the phase, size and height of high-level clouds, and compared them using fractal analysis and object tracking with the the Tracking and Object-Based Analysis of Clouds (TOBAC) Python package.

The different brightness temperatures simulated by RTTOV for the nested and the global simulation were found to be very similar, but noticeably different to observations (Fig. 1). The same was true for the phase, height and size products that we derived (Fig. 2). Nevertheless, we were able to determine threshold values for each product that could identify regions of ice-phased, high-altitude clouds with small ice crystals (Figs. 3 - 11). A qualitative analysis of the simulated products showed that the simulations were generally equally skilled in the beginning of the time-frame and in the region we examined. Small differences were found between simulations for the height and phase product in the northern regions of the WCB (Fig. 4 and 7), indicating that the convection parameterization does not adequately produce high-altitude, ice-phase cloud tops in this region. Towards the end of the time-frame, both simulations grossly misrepresented all three products in the region of the advancing cold front (Figs. 5, 8, 11). This clearly shows that the conclusions from Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b), as well as from Schwenk and Miltenberger (2024), should be read with the clear understanding that the high-resolution convection permitting simulation is not a perfect representation of reality, especially in the region of the advancing cold front.

We conducted a fractal analysis of objects masked according to the respective thresholds of the three products and determined that the nested simulation mostly accurately represents the observed fractal characteristics (Figs. 12 a, c and e). The global simulation mostly fails at this and produces fractal dimensions that are too large. The areas enclosed by the masked objects are larger in both simulations than in the observations (Figs. 12 b, d and f). This indicates that even the high-resolution simulation struggles to properly resolve smaller regions with high-altitude cloud tops. This finding is supported by the analysis of tracked object lifetimes, which show a larger peak at small lifetimes in the observations than in the simulations (Fig. 13). The fractal dimensions we determine in this paper are higher (indicating less complex structures) than in the studies conducted by Lovejoy and Mandelbrot (1985) and Batista-Tomás et al. (2015), who found values for α of approximately 1.4, compared to our values of approximately 1.6. This is presumably because of the scale at which we observe the WCB, which is more coarse than the respective studies.

The results of this paper show that, although both simulations misrepresent reality to some degree, the high-resolution, convection-permitting ICON simulation more realistically represents a WCB with respect to high-altitude and ice-phase clouds containing small ice crystals than the lower-resolution, convection-parameterizing simulation. We can therefore, cautiously, lend confidence to the conclusions from Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b), which identified large differences in the UTLS moisture content in the modeled outflow of a WCB and propose that the lower resolution simulation misrepresents the transport of moisture more strongly than the higher resolution simulation.

Since this paper is only a proof of concept based on a limited case study, there are many caveats to our conclusions that should be taken into account. Our analysis covers only a small domain and a time frame of one and a half days. Additionally, unlike in Schwenk and Miltenberger (2025a) and Schwenk and Miltenberger (2025b), we did not mask our data to isolate the WCB region. Therefore, the data and analysis presented here are very limited. We used the standard RTTOV installation and provided only the most necessary variables. A more sophisticated approach might produce brightness temperatures that align more closely with measurements. If this is accomplished, more products can be investigated and analyzed with many more tools. This could include using multiple threshold values, changing the minimum number of grid cells necessary to be included in the analysis, or conducting a full multifractal analysis similarly to Freischem et al. (2024). There are also more possibilities when using TOBAC than to just investigate track-lifetimes, such as object size a geographic pathways.

Code and data availability. Code and data will be made available after publication and will appear under the TPChange community on zenodo: <https://zenodo.org/communities/tpchange>. The code will also appear on my GitHub page <https://github.com/CornelisSchwenk/>.

Author contributions. CS and AM designed the experiment and conducted the numerical simulations. CS ran the RTTOV experiments wrote the post-processing code. CS and AM worked jointly on the interpretation of the results. CS drafted the final manuscript with contributions from AM.

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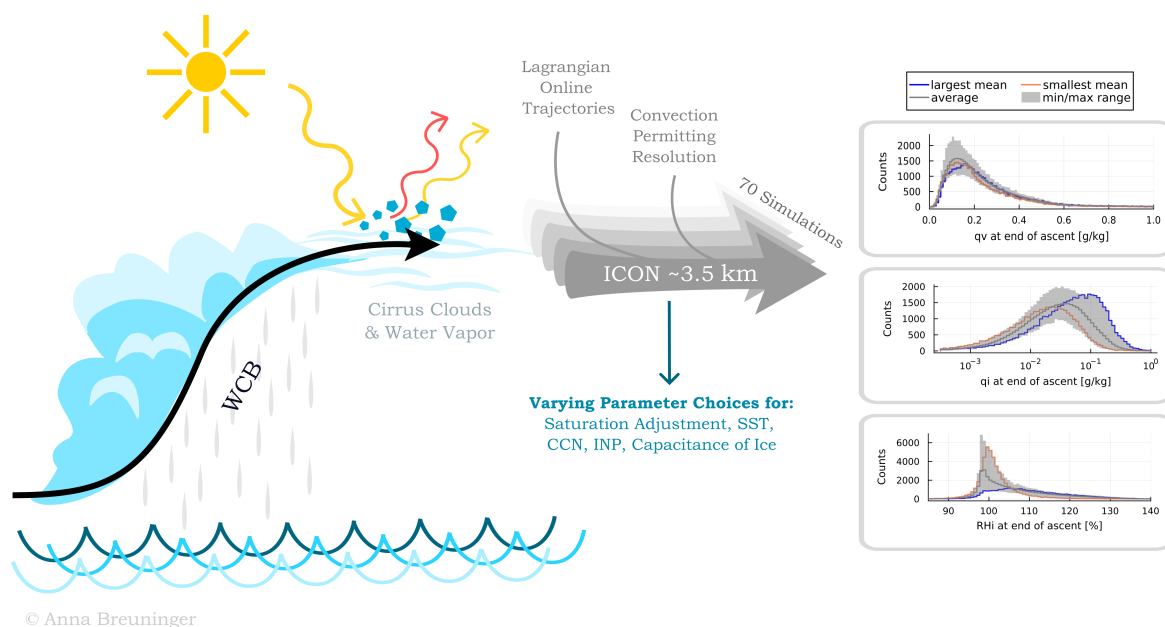
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Chapter 7

The Role of Parameter Perturbations

Research Question III

In what ways do perturbations to microphysical parameters affect modeled moisture conditions in the outflow of WCBs, and how does this effect vary with ascent timescale?



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Microphysical parameter choices modulate ice content and relative humidity in the outflow of a warm conveyor belt

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Abstract. Warm conveyor belts (WCBs) play a crucial role in Earth's climate by transporting water vapor and hydrometeors into the upper troposphere/lower stratosphere (UTLS), where they influence radiative forcing. However, a major source of uncertainty in numerical weather prediction (NWP) models and climate projections stems from the parameterization of microphysical processes and their impact on cloud radiative properties as well as the vertical re-distribution of water. In this study, we use Lagrangian data from a perturbed parameter ensemble (PPE) of a WCB case study to investigate how variations in microphysical parameterizations influence water transport into the UTLS and the outflow cirrus properties. We find that the thermodynamic conditions (pressure, temperature, specific humidity) at the end of the WCB ascent show little sensitivity to the explored parameter perturbations. In contrast, ice content and relative humidity exhibit substantial variability, primarily driven by the capacitance of ice (CAP) and the scaling of ice formation processes directly influenced by ice-nucleating particle (INP) concentrations. Different combinations of CAP and INP scaling yield vastly different ice and relative humidity distributions at the end of the ascent and in the subsequent hours. These differences are particularly pronounced in fast-ascending air parcels, where modifications to the saturation adjustment scheme (SAT) introduce small variations in pressure and temperature at the end of ascent. Our findings have potential implications for parameter choices in cloud models and considerations for geoengineering strategies. Future comparisons with high-quality observational data could help constrain the most realistic parameter choices, ultimately improving weather and climate forecasts.

1 Introduction

The most dominant greenhouse gas is water vapor (Schneider et al., 2010), and in the upper troposphere/lower stratosphere (UTLS) region, changes in its concentration lead to the most important positive feedback mechanism for climate change (e.g. Li et al., 2024; Held and Soden, 2000; Dessler et al., 2013). Even small changes in UTLS water vapor content can significantly alter the Earth's radiative budget (Wang et al., 2001; Hansen et al., 1984) and impact the mean and regional circulation patterns (e.g. Charlesworth et al., 2023; Ploeger et al., 2024). It is therefore crucial to quantify (i) the amount

and regional distribution of moisture in the UTLS (and how this has changed over time), (ii) the contributions of different transport pathways delivering moisture to the UTLS, and (iii) how these factors are likely to change in the future.

Although UTLS water vapor measurements have increased over the past 20 years (Zahn et al., 2014; Jeffery et al., 2022; Hurst et al., 2011; Tilmes et al., 2010; Konjari et al., 2024), the UTLS remains poorly characterized due to challenging measurement conditions (Jeffery et al., 2022) as well as instrument uncertainties, with calibrated measurements typically only accurate to within 5 %–10 % in super-saturated regimes (Petzold et al., 2020). In situ measure-

ments from aircraft and balloon-borne instruments provide high vertical resolution data on water vapor, but their limited regional and temporal coverage makes performing extensive climatologies difficult (Kunz et al., 2008; Zahn et al., 2014; Jeffery et al., 2022; Hurst et al., 2011). Satellite instruments offer long-term, global datasets, but they face challenges such as coarse vertical resolution and path length limitations (Hegglin et al., 2008, 2021). Therefore, accurate measurement and quantification of UTLS moisture, as well as changes over time, and quantification of its role in climate, remains a pressing and difficult challenge.

Moisture is transported to the UTLS through several pathways, with deep convection in the tropics being a major contributor. Both experimental (Corti et al., 2008; Danielsen, 1993; Lee et al., 2019; Gordon et al., 2024) and modeling studies (Ueyama et al., 2018, 2023; Dauhut et al., 2018; Hassim and Lane, 2010) have shown that UTLS moistening by convection occurs not only through the direct transport of water vapor but also via the sublimation of ice crystals detrained from convective clouds. Climate models predict that UTLS humidity will increase as the climate warms due to the warming of the tropopause (Gettelman et al., 2010), but Dessler et al. (2016) also determined that a large part of this can be attributed to the increase in sublimation of convective ice. It remains unclear how convective transport of moisture to the UTLS will change in the future, but a historical climatology by Jeske and Tost (2025) found that the height of convective outflow has shifted to lower pressures from 2011 to 2020 compared to the 1980s. While most of the aforementioned studies provide valuable insights into UTLS moisture transport and changes thereof, they focus primarily on the tropics.

Although more sparse, smaller in scope, and subject to large uncertainties, there have been studies investigating the transport of moisture in the extratropics. Homeyer et al. (2024) examined extratropical stratospheric hydration processes using observations over the continental United States, while Homeyer et al. (2014) and Homeyer (2015) conducted model analyses of overshooting convection in the same region. Heller et al. (2017) examined measurements from a case study of mountain waves over the Southern Alps in New Zealand and found evidence of vertical water vapor flux, suggesting that mountain waves have the potential to affect the water vapor distribution in the UTLS. Using satellite data, Sun et al. (2017) characterized seasonal and decadal changes in UTLS water vapor over the Tibetan Plateau, and found that in the summer, water vapor increased due to convective transport as determined by Fu et al. (2006). Globally, and covering the years from 2002 to 2012, Weigel et al. (2015) validated the SCIAMACHY satellite dataset to other satellite and in-situ measurement and found them to be reliable for altitudes between 15 and 23 km. However, and crucially, in the extratropics and for altitudes between 11 and 15 km, the mean percentage differences to other datasets exceeds 50 %. To determine the main transport pathways for UTLS

water vapor, Zahn et al. (2014) conducted a multi-year analysis of monthly UTLS water vapor measurements and identified warm conveyor belts (WCBs) as a key contributor to extratropical UTLS moisture. A recent climatological study by Guo and Miltenberger (2025) managed to quantify this contribution, and they determined that in December, January and February the moisture transported in WCBs by grid-scale advection accounts for up to 13.8 % (23.3 %) of the total water (condensate) transport into the upper troposphere above the North Atlantic and Pacific ocean basins.

Warm conveyor belts (WCBs) are large-scale, coherently ascending airstreams that develop near extratropical cyclones (ETCs) and generate the characteristic elongated cloud bands associated with ETCs (Madonna et al., 2014). Over approximately two days WCBs transport moist air from the boundary layer poleward and upward into the upper troposphere, while the air mass undergoes complex microphysical and dynamical processes. These processes contribute to precipitation formation, influence upper-level wave propagation, and inject water vapor and hydrometeors into the UTLS. As a result, WCBs play a crucial role by shaping mid-latitude weather systems, modulating storm development, and affecting Earth's radiative budget through cloud formation and moisture transport (Madonna et al., 2014). The ascending air produces significant amounts of precipitation, which can be hazardous (Pfahl et al., 2014); notably, WCBs account for more than 80 % of the total precipitation in Northern Hemispheric storm tracks (Pfahl et al., 2014; Eckhardt et al., 2004; Binder et al., 2016). In addition to producing precipitation, WCBs transport considerable amounts of energy as well as water vapor, hydrometeors, aerosols and trace gases across latitudes, and influence large-scale weather patterns (Joos et al., 2023; Madonna et al., 2014; Rodwell et al., 2018). The diabatic heating experienced by the ascending air masses affects cyclone strength and lifetime (Binder et al., 2016; Rossa et al., 2000; Binder et al., 2023) and produces potential vorticity (PV) anomalies (Oertel et al., 2020; Methven, 2014) that could influence atmospheric blocking (Pfahl et al., 2015; Wandel et al., 2024) and Rossby wave evolution (Pickl et al., 2023; Grams et al., 2011; Joos and Wernli, 2011; Wernli, 1997). Given that WCBs exert such wide-ranging influences, forecast skill is highly sensitive to how well models capture their path, vertical extent, diabatic-heating profile and mixed-phase cloud evolution, and any uncertainty in representing these aspects can quickly degrade predictions of cyclone intensity, heavy-precipitation placement, downstream wave development, and even heat waves (Pickl et al., 2023; Berman and Torn, 2019; Rodwell et al., 2018; Grams et al., 2018; Oertel et al., 2023).

WCBs influence Earth's radiative budget in multiple ways. During the early stages of ascent, when WCBs are usually located closer to the equator than during later stages, low-level clouds predominantly exert a cooling effect. As the WCB progresses northward, high-level frozen clouds have a warming or cooling effect depending on solar insolation, cir-

rus optical thickness, and potentially cloud overlap (Krämer et al., 2020; Joos, 2019; Johansson et al., 2019). In between, mixed-phase clouds have an uncertain net radiative effect. The outflow region of a WCB is also often accompanied by ice-supersaturated regions (Spichtinger et al., 2005) and an up to 3 km deep cirrus cloud shield (Binder et al., 2020), that has an net warming effect on average. Consequently, determining which overall effect dominates and how this balance may shift in the future remains a challenge (Joos, 2019). Additionally, it remains unclear how the transport of water vapor into the UTLS by WCBs will change in the future. For these reasons, accurately representing WCBs in NWP models is crucial for both weather and climate predictions.

However, accurately representing WCBs is challenging, partly because the ascending air undergoes a wide range of warm-phase, mixed-phase, and cold-phase microphysical processes (Binder et al., 2020; Forbes and Clark, 2003; Gehring et al., 2020), producing clouds with varying phases that add significant complexity to the model representation (Oertel et al., 2023; Schwenk and Miltenberger, 2024; Hieronymus et al., 2025). Additionally, the ascent occurs across different time-scales, which result in different dominating microphysical processes (Schwenk and Miltenberger, 2024). The precise composition of WCB clouds at different stages of the ascent is therefore difficult to determine and to represent in NWP models, and hence the parameterizations of the microphysical processes within them are a major source of uncertainty (Morrison et al., 2020; van Lier-Walqui et al., 2012; Posselt and Vukicevic, 2010; Oertel et al., 2025). Sensitivity experiments have demonstrated that the choice of microphysical parameterization schemes can significantly affect the characteristics of WCB ascent (Mazoyer et al., 2021, 2023), and that the diabatic processes occurring during WCB ascent are also sensitive to the specific choice of cloud microphysical parameters in a given parameterization scheme (Hieronymus et al., 2022; Neuhauser et al., 2023; Forbes and Clark, 2003). Classical sensitivity experiments, such as Monte Carlo simulations, could theoretically assess how specific parameterized processes contribute to uncertainties. However, applying these methods to NWP models across a large free parameter space, as typical for e.g. cloud microphysical parameterisations, is impractical due to their high computational cost and the vast amounts of data they produce. This has led to the employment of Perturbed Parameter Ensembles in atmospheric sciences (PPEs e.g., Lee et al., 2011; Collins et al., 2010; Johnson et al., 2015; Oertel et al., 2025), which are sensitivity experiments that sample the multi-parameter phase space spanned by the uncertain parameter values in a statistically optimal manner with a relatively small number of simulations.

In this paper we make use of a novel and unique PPE for a WCB case produced by Oertel et al. (2025), in which four cloud microphysical parameters and one parameter related to WCB inflow properties were modified. Oertel et al. (2025) specifically investigated (i) how the uncertainties in micro-

physical parameters influence the representation of WCB ascent and (ii) how the perturbed parameters affect the precipitation structure and larger-scale flow evolution. Our goal is to determine how microphysical parameter choices affect the transport of moisture and hydrometeors by ascending WCB air parcels into the UTLS. Schwenk and Miltenberger (2024) found that in the outflow of a WCB, the specific humidity (q_v) of Lagrangian WCB air parcels is predominantly constrained by the thermodynamic conditions at the end of ascent (i.e. the temperature and pressure in alignment with the Clausius-Clapeyron relationship). Remarkably, these results suggest that WCB air parcels are very efficient at converting water vapor into hydrometeors, otherwise, greater or more widespread supersaturation would be produced. This observation raises the question if this efficiency in vapor conversion reflects a physical reality, or if it is an artifact of the NWP model's potentially inadequate representation of vapor conversion in ascending air parcels?

One potentially inadequate representation could be the ICON model's saturation adjustment scheme, which forces 100 % saturation over water at every model time step as long as the parcel contains any liquid water (by either evaporating cloud condensate or condensing excess vapor). However, in reality supersaturation over water can form under conditions of high vertical velocity. Making the saturation adjustment scheme more realistic by allowing for some supersaturation over water when vertical velocities are high might change the vapor conditions in the outflow of a WCB. However, most WCB air parcels glaciate completely during their ascent, meaning that they spend a lot of time in a purely ice phase (Schwenk and Miltenberger, 2024). In this case other parameters – such as the capacitance of ice (CAP) or CCN as well as INP concentrations – likely play a more significant role in determining the efficiency of vapor conversion. For instance, an increase in the capacitance could accelerate ice particle growth, while higher concentrations of CCN can lead to more and smaller cloud droplets in the liquid phase, which can in turn increase the number of ice crystals in the frozen phase and enhance vapor conversion. Similarly, a greater concentration of INP might result in a faster transition to the ice phase, further affecting the overall conversion process. If the perturbations can significantly influence the ice content, ice number concentration or cirrus lifetime in the WCB outflow, they could also alter the radiative budget of the WCB, which would be important to consider when analyzing the future impact of WCBs on global warming. Such impacts would also be relevant for potential geoengineering strategies, such as marine cloud brightening (Latham et al., 2012) or cirrus cloud thinning (Lohmann and Gasparini, 2017; Villanueva et al., 2022; Gasparini et al., 2020), which aim to modify CCN or INP levels to change the radiative properties of clouds. The findings obtained from this paper might provide valuable insights for future geoengineering efforts seeking to manipulate cloud properties to achieve climate-modulating effects.

An additional consideration for moisture transport in WCBs is the role of convectively ascending air parcels. Schwenk and Miltenberger (2024) compared the transport of water for fast (convective) and slow ascending WCB air parcels and found that rapidly ascending WCB air parcels carry substantial amounts of ice into the UTLS. This could have notable implications for the radiative budget of WCBs, since high level clouds can contribute to global warming (Haslehner et al., 2024), and—as discussed above—sublimating ice could increase the UTLS vapor content (Dessler et al., 2016). This raises two additional questions: (i) how do different parameter choices affect the mass and number of ice particles (or snow) transported into the UTLS differently for fast and for slow ascending air parcels, and (ii) how does moisture transport change when we change the fraction of convectively ascending air parcels (by modulating the SST)?

In summary, this article addresses the following research questions by analyzing a PPE of a WCB case: (How) do perturbations in microphysical parameters and SST (i) influence moisture conditions in the WCB outflow, (ii) alter hydrometeor distributions, particularly for ice, in the WCB outflow, and (iii) how do these effects vary with ascent velocity and (iv) evolve in the hours following the end of ascent?

2 Methods

The WCB trajectory data we investigate in this study are taken from a perturbed parameter ensemble (PPE) which is described in detail in Oertel et al. (2025). In this section, we provide a brief description of the ICON model setup, the design of the PPE, and an overview of the chosen parameter perturbations (Sect. 2.1, 2.2), as described by Oertel et al. (2025). We then proceed to explain how we select WCB trajectories and define the beginning and end of ascent (Sect. 2.3). Finally, we explain the statistical methods used to analyze the impact of parameter choices on model output (Sect. 2.4).

2.1 ICON model Setup and Lagrangian data

A global simulation with the Icosahedral Nonhydrostatic (ICON) modelling framework (Zängl et al., 2014; version 2.6.2.2) with approximately 13 km grid spacing (R03B07 grid) was run for 72 h, initialized from the ECMWF analysis at 18:00 UTC on 3 October 2016, with a 120 s time step. Two nested domains with grid spacings of approximately 6.5 km (R03B08) and approximately 3.3 km (R03B09) and time steps of 60 and 30 s, respectively, cover the WCB ascent region and interact with the global simulation via two-way feedback. All domains have 90 vertical model levels, and in the WCB outflow region (~ 7 – 12 km) the vertical grid-spacing is between 200 and 400 m. The two moment cloud microphysics scheme from Seifert and Beheng (2006) is used, which contains six prognostic hydrometeor

types (cloud liquid (q_c), rain (q_r), ice (q_i), snow (q_s), graupel (q_g) and hail (q_h)). In the global domain convection is parameterised with the Tiedke-Bechtold convection scheme (Tiedtke, 1989); in the higher-resolution nests only shallow convection is parameterised. For all other parameterisations the standard schemes are used (see Oertel et al., 2025). The described set-up is used to perform 70 simulations with varying lower boundary conditions and cloud microphysics parameters as detailed in Sect. 2.2.

The online trajectory module by Miltenberger et al. (2020) and Oertel et al. (2023) is used to calculate air-mass trajectories (see Schwenk and Miltenberger, 2024; Oertel et al., 2025 for details). Trajectories are started from six vertical levels between 200 and 1500 m every 1 h during the simulation in a longitude and latitude region (55 to -15° E, 35 to 55° N) predefined by prior offline trajectory calculations with LAGRANTO (Sprenger and Wernli, 2015). Trajectory data is stored every 1 h. This study exclusively analyzes the Lagrangian data obtained from the PPE.

A detailed overview of the synoptic situation and trajectory data is given in Oertel et al. (2023) and Oertel et al. (2025). Here, we provide a short summary. The extratropical cyclone that produced the WCB cloud band examined in this paper originated at approximately 40° W and 50° N in the early hours of 4 October 2016. It propagated in the Eastern direction and by 4 October 2016 12:00 UTC, a clear front formed at approximately 30° W (Fig. 1a). After this time, the typical comma-shaped WCB cloud band formed and the cyclone propagated to the Northeast, reaching Iceland and the southeastern Greenland shore at 5 October 2016 00:00 UTC (Fig. 1b). The WCB progressed further Northeast (Fig. 1c) and at 6 October 2016 00:00 UTC (Fig. 1d) it began to dissipate over the arctic.

2.2 Perturbed parameter ensemble design

The PPE designed by Oertel et al. (2025) simultaneously perturbs the values of five parameters, which are kept constant throughout the simulation. These parameters are sea surface temperature (SST), the capacitance of ice and snow (CAP), the impact of ice nucleating particle (INP) number concentration on ice formation, cloud condensation nuclei (CCN) number concentration, and the saturation adjustment scheme (SAT). The specific parameter combinations for individual PPE members are chosen such that the PPE optimally explores the entire phase space spanned by the selected parameters using a Latin hypercube design. In total, 70 PPE members are available. For an overview of each parameter see Oertel et al. (2025); here we only provide a brief overview and present the chosen parameter ranges in Table 1.

CAP: The capacitance of frozen hydrometeors (CAP) influences the rate at which water vapor deposits onto ice surfaces, thereby influencing the growth of ice and snow, the phase state of the water substance in an air parcel, and the associated latent heat release. While theory predicts a nor-

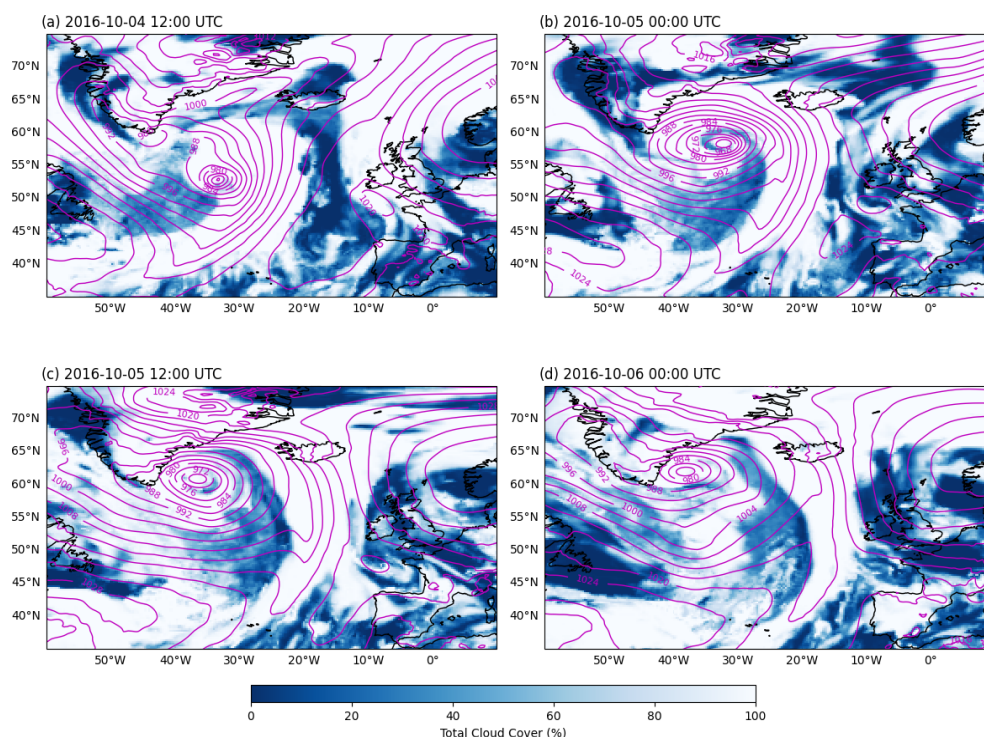


Figure 1. ERA5 data showing the total cloud cover and sea level pressure at 4 October 2016 12:00 UTC in (a), 5 October 2016 00:00 UTC in (b), 5 October 2016 12:00 UTC in (c) and 6 October 2016 00:00 UTC in (d).

Table 1. Overview of parameter perturbations (taken from Oertel et al., 2025).

Parameter	Min	Max	Unit	Spacing	Process
SST	−2	+2	K	linear	sea surface temperature/humidity/latitude
CCN	0.4	20	1	linear	cloud droplet activation
INP	0.01	20	1	logarithmic	ice nucleation
CAP	0.2	1	1	linear	growth of ice/snow by vapor deposition
SAT	0	0.1	1	linear	saturation adjustment scheme

malized CAP value of 0.5 for perfectly spherical particles, realistic ice and snow hydrometeors often exhibit considerably different values (Westbrook et al., 2008; Chiruta and Wang, 2003). Consequently, in the PPE, CAP for ice and snow is varied (simultaneously) by a scaling factoring ranging from 0.2 to 1, which explores a range for CAP from 0.1 to 0.5. CAP values for graupel and hail, which are assumed to be sufficiently spherical, are not changed.

INP: INP concentrations are important for cloud and precipitation formation, can alter cloud radiative forcing (CRF) (Schrod et al., 2020; Burrows et al., 2022), and their concentrations in the atmosphere vary over several orders of magnitude (Hande et al., 2015; DeMott et al., 2010). However, there are still large gaps in knowledge about their spatial, vertical and seasonal distributions and their representation in NWP models is highly uncertain (Wex et al., 2019, 2024;

Schrod et al., 2020; Li et al., 2022). In the PPE, Oertel et al. (2025) did not directly perturb the INP concentrations that are calculated by the ICON model, but instead scaled three processes which are known to be directly influenced by INP number concentrations-immersion freezing of cloud droplets, deposition nucleation of ice, and freezing of rain drops-by applying a logarithmically spaced factor ranging from 0.01 to 20 to emulate varying INP concentrations. This scaling is designed to emulate scaling the number of INPs. In ICON, the heterogeneous freezing parameterization includes immersion freezing of cloud droplets (Hande et al., 2015), deposition nucleation of ice (Hande et al., 2015) and freezing of rain drops (Bigg, 1953). In this scheme, immersion freezing is active below -12°C as long as cloud droplets are present. The number of activated INPs is a function of temperature, and the increase in ice-number concentration per

time-step depends on the current and previous number of activated INPs. Deposition nucleation takes place in a temperature range from -20 to -53 °C and the number of activated INPs additionally depends on supersaturation with respect to ice (Lüttmer et al., 2025). For example: at -15 °C and 110 % relative humidity over ice, the immersion freezing INP concentration amounts to $\approx 5 \text{ m}^{-3}$ for the unperturbed reference and varies between 1 and $\sim 100 \text{ m}^{-3}$. The rate of freezing of rain is temperature dependent; below -40 °C all rain drops instantly freeze. Ice-formation parameterizations in ICON that remain unchanged in the PPE are (i) the homogeneous nucleation, parameterized following Kärcher and Lohmann (2002) and Kärcher et al. (2006), where the freezing rate and critical supersaturation are functions of temperature; (ii) homogeneous freezing of cloud droplets (Jeffery and Austin, 1997), where below -50 °C, all cloud droplets freeze instantly; and (iii) secondary ice production (Hallet and Mossop rime splintering), active between approximately -8 and -3 °C (Hallet and Mossop, 1974; Seifert and Beheng, 2006).

CCN: The cloud droplet activation scheme implemented in ICON was originally designed for continental Germany by Hande et al. (2016) and assumes a vertical CCN profile that decreases as pressure decreases. This profile does not change over time nor across the globe, and droplet activation scales with the resolved vertical velocity. However, it is known that CCN number concentrations are not constant, but vary in space and time, depending on factors such as the season and air mass origin (Schmale et al., 2018; Rose et al., 2021). Over continents they can also differ substantially from concentrations over the open ocean. Therefore, Oertel et al. (2025) used a modified cloud droplet activation scheme designed by Oertel et al. (2023) to account for CCN concentrations over the northern Atlantic. The details are described in the latter publication, but we point out that, in the new scheme, cloud droplet activation continues to scale with pressure and vertical velocity, and the profile remains the same globally (i.e. it is a tailored adjustment for an open-ocean WCB). In the unperturbed reference, the CCN concentration is approximately $\sim 250 \text{ cm}^{-3}$ near the cloud-base (at $\sim 800 \text{ hPa}$) for large vertical velocities (see Fig. 2 in Oertel et al., 2023). In this PPE, the uncertainty in representing CCN number concentrations is accounted for by scaling the number of activated cloud droplets throughout the vertical profile at each model time step by a factor ranging from 0.4 to 20. This variation results in CCN concentrations varying from approximately 100 to 5000 cm^{-3} , which approximately represents the variability that has been observed for CCN concentrations (Hande et al., 2015, 2016; Genz et al., 2020; Wang et al., 2021).

SAT: As discussed in the introduction, when a cloud contains liquid water, the saturation adjustment scheme used in ICON instantaneously removes supersaturation or in-cloud sub-saturation by condensing or evaporating excess or deficient water vapor until a relative humidity of 100 % over water is achieved. This becomes increasingly unrealistic when

vertical velocities are high or CCN numbers very low (Lebo et al., 2012). In the PPE, the scheme is therefore modified to allow supersaturation over water to develop under conditions of strong vertical velocity. This modification is done by including a scaling factor f_{SATAD} , ranging from 0 to 0.1, which is multiplied with the vertical velocity (only when vertical velocities are larger than zero), added to one, and finally multiplied with the specific humidity from the default scheme:

$$q_{v,s}^*(T, p) = (1 + (f_{\text{SATAD}} \cdot w)) \cdot q_{v,s}(T, p) \mid w > 0. \quad (1)$$

A factor f_{SATAD} of 0 corresponds to the standard scheme, which inhibits the formation of supersaturated conditions. For a factor f_{SATAD} of 0.1, vertical velocities of 2 m s^{-1} can produce supersaturations of 20 %, which aligns with theoretical examinations of supersaturation in liquid clouds (Korolev and Mazin, 2003; Morrison and Grabowski, 2008).

SST: Tropospheric humidity in the inflow region of a WCB is crucial for the subsequent ascent (Schäfler and Harnisch, 2014; Christ et al., 2025; Berman and Torn, 2022, 2019) and formation of precipitation (Dacre et al., 2019). Yet NWP models can have substantial humidity errors (Schäfler et al., 2010, 2011). Since the moisture content of marine boundary-layer air is mostly controlled by temperature, the PPE modifies SST (which in ICON is kept constant throughout the simulation) within the range of $\pm 2 \text{ K}$ to analyze the sensitivity to both the temperature and moisture in the WCB inflow region. A difference in SST can also be seen as analogous to a shift in latitude or as an uncertainty in cyclone position or WCB inflow location.

2.3 Trajectory selection and ascent time characterisation

WCB trajectories must fulfill the criteria of ascending at least 600 hPa within 48 h. We use the widely used ascent time-scale τ_{600} , defined as the shortest time taken per trajectory to ascend at least 600 hPa, to characterise the ascent time of a WCB trajectory. As we are most interested in the characteristics of trajectories once they have completed their ascent and have entered the UTLS, we do not only consider the τ_{600} ascent-segment, but the entire time around this segment during which the ascent velocity remains above 8 hPa h^{-1} (using the algorithm described by Schwenk and Miltenberger, 2024). We refer to this time as τ_{WCB} . The end of ascent is defined as the last time-step in τ_{WCB} . Note that in contrast, Oertel et al. (2025) consider the τ_{600} segment.

2.4 Methods for quantifying parameter perturbation impacts

The impact of parameter perturbations are investigated for selected properties of WCB outflow (target variables): temperature, pressure, specific humidity, relative humidity over ice and hydrometeor content at (i) the end of WCB ascent (see Sect. 2.3), (ii) during the ascent and (iii) 5 h after the end

of ascent. The distribution of these values across all selected trajectories is considered as are summary statistics characterizing the distribution. The latter are particularly useful for further quantification of the parameter perturbation impact, for which we use Spearman correlation coefficients and random forest regression models. For each PPE member, the mean, median, standard deviation, as well as the 5th, 25th, 75th, and 95th percentiles over all trajectories (in that PPE member) are calculated for the target variables. Taking the specific humidity q_v as an example, we use the following notation for these statistics: q_v^{mean} , q_v^{median} , q_v^{std} , $q_v^{5\text{th}}$, and so on.

The values of the variables are usually taken at the end of ascent (e.g. ice content, temperature), but some are also taken during the ascent (e.g. temperature at 99 % glaciation, maximum hydrometeor content achieved during ascent) or after the ascent (e.g. ice content 5 h after completing τ_{WCB}). If the latter is the case this is explicitly stated in the text. In certain cases, trajectories from multiple PPE members are examined in detail (as opposed to looking only at the summary statistics), particularly to investigate how distributions change with changing parameters.

Spearman correlation. One metric for determining the impact of a parameter on the summary statistic of some target variable is the Spearman correlation coefficient (R_s). In contrast to the “regular” correlation coefficient, it ensures that also non-linear but monotonic correlations (e.g., an exponentially decreasing correlation) result in a high coefficient. If a parameter choice strongly influences a variable (monotonically), then R_s for the relevant summary statistic will be high. To see if the impact of the correlation is noticeable compared to the spread of values within an ensemble member, we use a scatter plot of mean values with shaded percentiles.

Random Forest Regression Models. If multiple parameters impact a variable in a complex and non-linear way, using simple correlation coefficients is not sufficient. We therefore also employ random forest (RF) regression models (Breiman, 2001), which can capture complex and non-linear interactions between the perturbed parameters and the target variables. To determine which PPE-parameters had the most influence on the RF-output, we examine the impurity-based feature importance (IBF) scores, which quantify the reduction in variance contributed by each parameter across all decision trees in the forest. Note: a high IBF importance score for a PPE-parameter does not necessarily imply a clear or strong correlation with the output variable. Instead, it indicates that the PPE-parameter contributes strongly to the RF-model’s predictions, possibly by affecting the RF-model’s decision-making structure only in combination with other PPE-parameters, and is only a meaningful metric when the RF-prediction is good.

We use partial dependence plots (PDPs) to visualize interactions between the target response X and the selected perturbed parameters (Friedman, 2001). To construct a PDP, the model prediction is calculated for a range of values of the fea-

ture of interest, while holding all other features constant. This process is repeated across all possible values of the chosen feature, and the average predicted response is plotted against the feature values. The resulting PDP visualizes the effect of the selected feature on the target response, providing insights into the feature’s influence and possible nonlinear relationships within the model. A two-dimensional PDP can help in identifying joint parameter impact on a target variable.

In summary, the purpose of our RF analysis is mostly to identify whether parameter changes have an impact, and if so, which parameters are the most important. We take the following approach:

- Train RF model for target variable X , called RF_X^1 (for example $\text{RF}_{\text{mean}(q_v)}^1$).
- Define most important perturbed parameters as those with an IBF score larger than 0.1.
- Train a new RF model using only these perturbed parameters with the same target variable (RF_X^2).
- Evaluate whether said parameters have a strong effect on target variable by looking at R^2 (linear correlation coefficient) and root mean square deviations (NRMSE) between RF_X^2 prediction and actual values (average over 100 model iterations).
- (Optional) construct and interpret PDPs using RF_X^2 .

It is important to clarify that we are *not* using the random forest model to make predictions in the conventional sense, especially as our analysis is conducted on the training data itself. Instead, if the model prediction is adequate, we assume that the model is able to find relationships between the output variables and the perturbed parameters. By examining feature importance and partial dependence plots, we aim to investigate these patterns, rather than to predict outcomes on new data. We selected model configurations that balance RF-model complexity with the need for generalization given the small dataset size (70 data points when using summary statistics and six perturbed parameters). Each model was configured with 100 trees to ensure sufficient coverage of feature space, we set the maximum tree depth to 3 to prevent overfitting and specified a minimum sample split of 10, which constrains each split to be based on a meaningful subset of the data.

3 Properties of WCB outflow at the end of ascent

In this section we provide an overview of the properties of WCB trajectories at the end of ascent, since this is where we focus most of our analysis. Spatially, the end of the ascent is spread across a large area, ranging in longitude from 60° W to more than 45° E (Fig. 2c). However, the core of end-of-ascent longitude positions are focused around 60 to 10° W

(Fig. 2a) with only very few extending further east. In latitude, trajectories finish their ascent between 35 and 85° N (Fig. 2c), with largest density at approximately 40, 50, 58 and 68° N (Fig. 2b), indicating the presence of large-scale organization in the ascent of air within the WCB. For the discussion of changes in the positions of WCB trajectories with parameter perturbations, the reader is directed to Oertel et al. (2025), since this is not the area of focus in this study.

At the end of ascent, trajectories across all PPE members ascend to pressures between approximately 200 and 400 hPa (Fig. 3c and d). Temperatures range from approximately −60 to −30 °C; the majority of trajectories (ca. 75 %) have temperatures below the homogeneous freezing temperature of −38 °C (Fig. 3a and b). Trajectories are very dry when they complete their ascent, with an average q_v^{mean} of 0.23 g kg^{−1} (Fig. 3e and f). However, the distribution of q_v in each PPE member is skewed towards higher values, meaning that although the median q_v at the end of ascent ranges from 0.17 to 0.20 g kg^{−1} for PPE members, all PPE members also have (few) trajectories with q_v larger than 0.6 g kg^{−1} at end of ascent. We discuss the effects of parameter perturbations on T , p and q_v in Sect. 3.1.

Almost all trajectories are fully glaciated by the end of their ascent and less than 1 % of all trajectories have cloud droplet mass mixing ratio (q_c) larger than zero (not shown). Trajectories at end of ascent also contain no graupel and no rain (not shown). The only relevant hydrometeor species are therefore ice and snow, with the former being the most predominant (with ice mass mixing ratio q_i^{mean} and ice number mixing ratio N_i^{mean} being on average 0.046 g kg^{−1} and 2.49×10^6 kg^{−1}, compared to q_s^{mean} and N_s^{mean} which are on average 0.001 g kg^{−1} and 540 kg^{−1}). We therefore focus our analysis on q_i at the end of ascent, which shows a peak at small values close to ~ 0.05 g kg^{−1} and a long tail to larger values of up to around 0.3 g kg^{−1} for the PPE member with the highest mean q_i (Fig. 3g). Hydrometeor contents are therefore roughly one order of magnitude smaller than q_v . In contrast to q_v , we find that q_i varies much more strongly between PPE members (Fig. 3g and h). The standard deviation of q_i^{mean} across all PPE members is 0.01 g kg^{−1} (vs. 0.004 g kg^{−1} for q_v^{mean}) and the highest q_i^{mean} (0.080 g kg^{−1}) is almost 3 times larger than the smallest q_i^{mean} (0.027 g kg^{−1}). This corroborates the visual impression from Fig. 3h of substantial changes in the q_i distribution in particular with respect to the occurrence of high q_i values. We assess the effects of parameter perturbations on the hydrometeor content at the end of ascent in Sect. 3.2.

The mean relative humidity over ice (RH_i) for all PPE members is larger than 100 % (Fig. 3j), but also here we find much stronger differences between PPE members (Fig. 3i and j) than for T , p and q_v , which is particularly interesting because the differences in q_v are small. The smallest $\text{RH}_i^{\text{mean}}$ of 101.2 % is not only much smaller than the largest (109.8 %), but the two distributions are also markedly different (Fig. 3i): The RH_i distribution from the PPE member with

Table 2. Impurity-based feature importance (IBF) scores of CAP, INP, CCN, SST and SAT derived from RF_X^1 for a selection of mean variables. Only scores ≥ 0.1 are included. The bottom rows show the R^2 and NRMSE value for the predictions of the second iteration of the forest regression model (RF_X^2) using only parameters with $\text{IBF} \geq 0.1$. Note: all of these values are slightly different each time a forest model regression is performed; the values shown here are averaged over 100 iterations.

	T^{mean}	p^{mean}	$\text{RH}_i^{\text{mean}}$	q_v^{mean}	q_i^{mean}	N_i^{mean}
CAP	0.295	0.143	0.319	0.115	0.486	0.434
INP	0.384	0.249	0.304	0.322	0.251	0.287
CCN		0.299	0.229		0.100	0.131
SST	0.136	0.132		0.382		
SAT	0.112	0.177				0.100
R^2	0.612	0.558	0.765	0.583	0.727	0.700
NRMSE	0.618	0.660	0.481	0.641	0.518	0.542

the smallest $\text{RH}_i^{\text{mean}}$ is relatively narrow with only few trajectories exceeding 110 % RH_i , while the distribution from the PPE member with the largest $\text{RH}_i^{\text{mean}}$ is very broad featuring RH_i values above 110 % (120 %) in 44 % (15 %) of the trajectories. We interpret the effects of parameter perturbations on the relative humidity at the end of ascent in Sect. 3.3.

3.1 Parameter effects on temperature, pressure and specific humidity at the end of ascent

As discussed above, T , p and q_v show little variation between PPE members (Fig. 3b, d and f), with the variability within each PPE member being significantly larger than the mean differences between PPE members. The distributions for PPE members with the highest and smallest mean values for T , p and q_v are also very similar (Fig. 3a, c and e). The relatively large spread of the distribution (min/max shaded areas in Fig. 3a, c, and e) is a result of the differing number of WCB trajectories per PPE member (which is primarily controlled by SST, see Oertel et al., 2025). The $\text{RF}_{\text{mean}(p)}^2$ and $\text{RF}_{\text{mean}(T)}^2$ model predictions (which we use to determine which parameters have the strongest influence on a variable, as long as the model prediction is strong) are also relatively weak, with mediocre R^2 and large NRMSE-values (Table 2). Therefore, the RF model does not find meaningful changes in T^{mean} , p^{mean} and q_v^{mean} with parameter perturbations. We interpret this inability of the parameters to change the pressure, temperature, and vapor conditions at the end of the WCB ascent as an indication that the thermodynamic conditions in the outflow of a WCB are largely constrained.

However, some relationships are worth noting. For example, $T^{95\text{th}}$ and $q_v^{95\text{th}}$ show an unmistakable correlation with SST perturbations (Fig. A1c in the Appendix) and to each other (Fig. A1a). Oertel et al. (2025) found that increasing SST leads to higher potential temperatures and a greater number of WCB trajectories, as enhanced latent heat re-

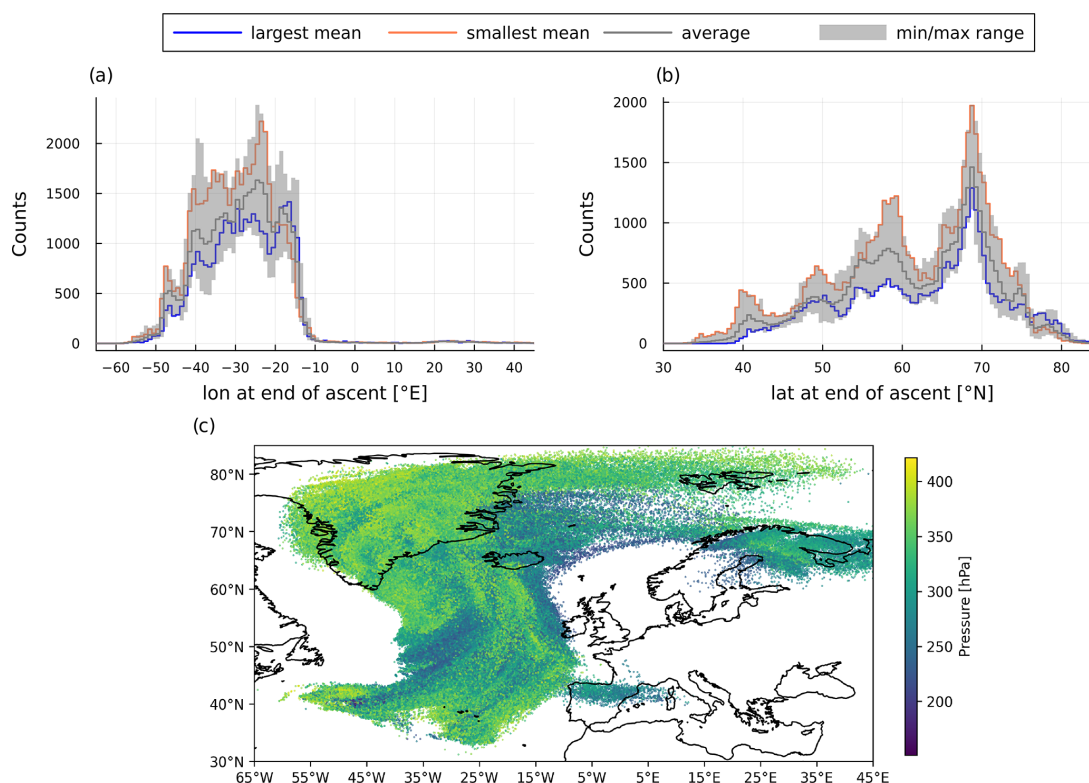


Figure 2. Histograms for (a) longitude and (b) latitude at the end of ascent, with each showing distributions for PPE members with the largest mean (blue) and smallest mean (orange), as well as the average distribution in grey and the min/max distributions per variable bin in the shaded area. Panel (c) shows the trajectory positions at the end of the ascent for all PPE members (note that trajectories finish their ascent at different times in the simulation), with pressure as the color scale.

lease facilitates more cross-isentropic ascent. Consequently, in PPE members with higher SST, more trajectories are able to meet the 600 hPa ascent threshold that would otherwise fall short at lower SST values. We therefore attribute the increase in $T^{95\text{th}}$ and $q_v^{95\text{th}}$ with SST to the larger number of WCB trajectories that end their ascent at relatively high pressures (~ 400 hPa) and lower latitudes ($\sim 40^\circ\text{N}$) in high-SST simulations (not shown). Trajectories that fall short of the 600 hPa threshold are more likely to be absent from the WCB trajectory set in PPE members with lower SST.

An important finding from a previous analysis of q_v in WCB outflow by Schwenk and Miltenberger (2024) was that q_v values are largely controlled by temperature. We find the same in the present PPE: the specific humidity for the ensemble members with the highest and lowest $q_v^{95\text{th}}$ values correlates strongly with the calculated saturation specific humidity (using parcel temperature T and pressure p , Fig. A1b; $R_s \approx 0.99$ for both). This strong correlation indicates that in both PPE members the specific humidity at the end of ascent is predominantly constrained by the thermodynamic conditions, i.e., changes in q_v are a result of changes in temperature.

In summary, the parameter perturbations in our PPE only have a limited effect on changing the distributions of T , p and q_v at the end of ascent, indicating that the conditions at the end of ascent experience strong thermodynamic constraints.

3.2 Effect of parameter perturbations on the hydrometeor content (during and at the end of the ascent)

At the end of ascent, virtually all hydrometeors are ice (Sect. 3). The summary statistics as well as distributions of mass mixing ratios q_i show large differences between PPE members (Fig. 3g and h) suggesting that the parameter perturbations influence the hydrometeor content in the outflow of the WCB. The IBF scores (Table 2) indicate that CAP and INP scaling factors are the most important parameters for the mean ice number (N_i) and mass (q_i) mixing ratio. The CCN scaling factor shows an IBF score of approximately 0.1 for these variables, indicating that it is less important in the ice-phase than it is in the liquid-phase and mixed-phase cloud regime (as is shown later in this section).

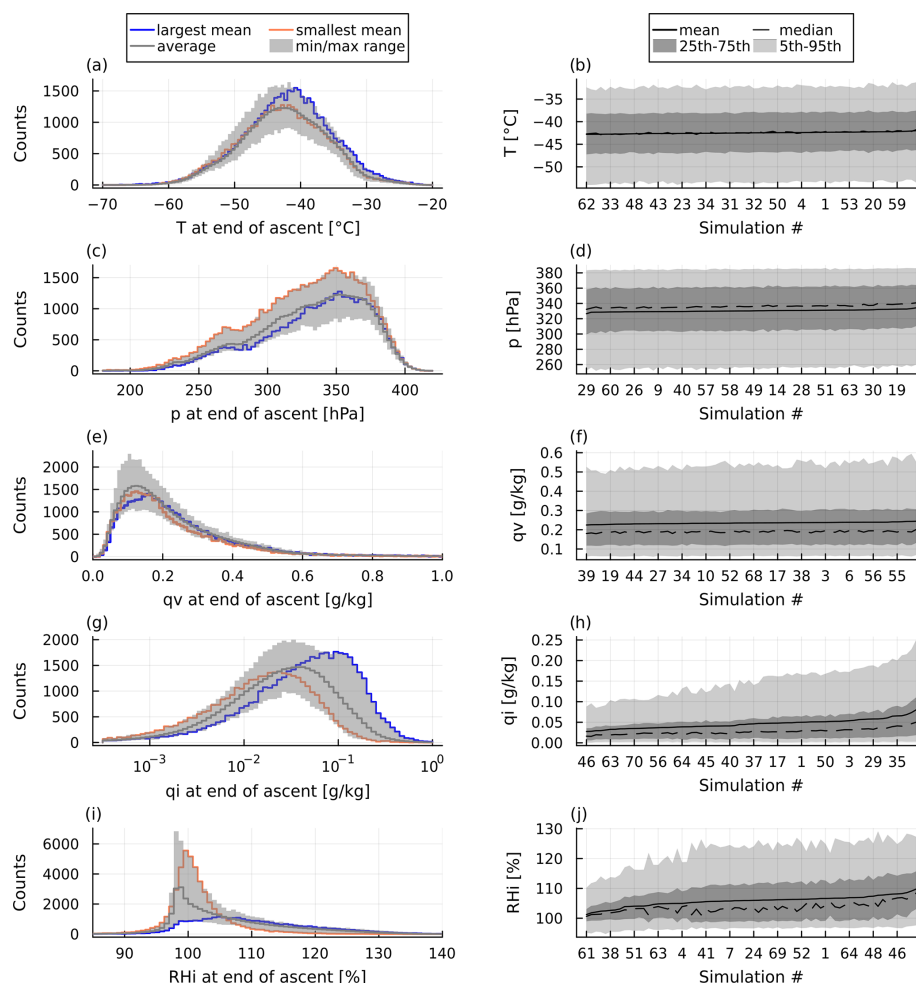


Figure 3. Histograms for (a) temperature T , (c) pressure p , (e) specific humidity q_v , (g) ice mass mixing ratio q_i (note the logarithmic x -axis) and (i) relative humidity over ice RH_i at the end of ascent. Each figure shows distributions for PPE members with the largest mean (blue) and smallest mean (orange) of the respective variable being depicted, as well as the average distribution in grey and the min/max distributions per variable bin in the shaded area. Panels (b), (d), (f), (h) and (j) show T^{mean} , p^{mean} , q_v^{mean} , q_i^{mean} and RH_i^{mean} per PPE member, respectively, sorted according to the mean, with 5th to 95th percentiles and inter-quartile range shaded in grey.

3.2.1 Impact of CAP on ice properties at end of ascent

Increasing CAP clearly decreases q_i^{median} at the end of the ascent while the spread in values remains largely unaltered (Fig. 4a, note logarithmic y -axis). The spread (in the percentiles) is mostly unchanged because the distribution and shape of q_i values remains similar for large and small CAP (Fig. 4d). Increasing CAP also strongly decreases N_i^{median} at the end of the ascent (Fig. 4b) and reduces the spread; the simulations with the largest and smallest values for CAP have standard deviations for $\log_{10}(N_i)$ of 0.54 and 1.01, respectively. This decrease in N_i and spread is a result of a large difference in the distribution shape of N_i values for PPE members with large and small CAP (Fig. 4e). When CAP is small (< 0.4) the distribution of N_i shows a large

spread and hints at bi-modality, with one peak (or mode) at approximately 10^5 kg^{-1} and a second mode at approximately $5 \times 10^6 \text{ kg}^{-1}$. When CAP is large (≥ 0.8) the bi-modality almost disappears, reducing the spread. Additionally, the peak of the distribution shifts to smaller N_i for PPE members with larger CAP. The radius r_i , on the other hand, increases clearly with CAP, as does the spread (Fig. 4c and f).

We interpret these findings as follows: An increase in CAP enhances the deposition rate, which, in the mixed-phase, can accelerate the Wegener–Bergeron–Findeisen (WBF) process. Consequently, when CAP is high, ice crystals grow more rapidly, depleting vapor more efficiently and accelerating the evaporation of cloud droplets. This leads to an earlier onset of full glaciation (see the Supplement, Fig. S1) – defined as the first time step when q_c reaches zero – at higher pres-

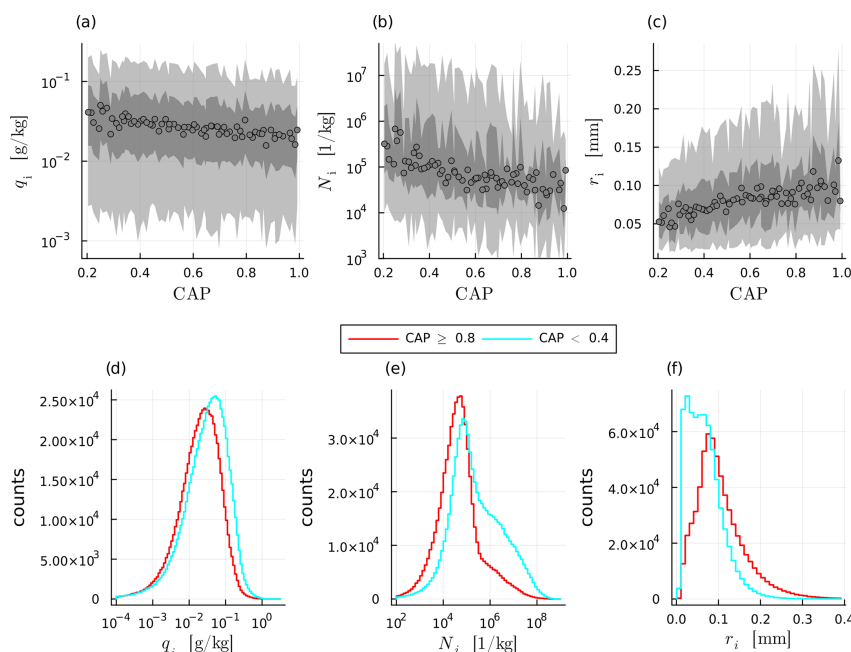


Figure 4. (a) q_i , (b) N_i and (c) ice radius (r_i) at the end of the ascent over CAP, with makers showing median values and shaded areas the 5th to 95th percentiles (light grey) and 25th to 75th percentiles (dark grey). Panels (d), (e), and (f) show q_i , N_i and r_i at the end of the ascent respectively, but as histograms for all PPE members with CAP values either larger than 0.8 (red; in total 16 PPE members and 633 400 trajectories) or smaller than 0.4 (cyan; in total 18 PPE members and 696 846 trajectories).

tures and temperatures. Thus, fewer cloud droplets reach the homogeneous-freezing level of -38°C , where they would otherwise freeze and generate numerous small ice crystals. This hypothesis is supported by the fact that the only physical process producing ice-crystal number mixing ratios in the order of 10^7 – 10^8 kg^{-1} is homogeneous cloud freezing. Heterogeneously produced ice-crystal number mixing ratios cannot exceed the number of INPs, which is on the order of 10^4 kg^{-1} . More homogeneously freezing cloud droplets when CAP is low explains why the distribution appears slightly bi-modal at low CAP values – indicating that a large proportion of trajectories contain small cloud droplets that reach the homogeneous-freezing level – but becomes uni-modal at high CAP values. Furthermore, an increased deposition rate leads to the formation of larger ice crystals (reflected in the increase of r_i with CAP), which have higher terminal velocities and precipitate more quickly. This precipitation process reduces the peak number mixing ratio of ice crystals N_i and, ultimately, the ice mass mixing ratio q_i .

3.2.2 Impact of INP on ice properties at end of ascent

The effects of varying the INP scaling factor exhibit similarities to those observed for CAP, but with some notable differences. As for CAP, an increase in the INP scaling factor leads to a decrease in q_i^{median} , albeit to a lesser extent than CAP (Fig. 5a). Similarly to CAP, low INP scaling factor val-

ues result in a distribution of N_i that is slightly bi-modal with a high spread, and high values result in a uni-modal distribution with a low spread (Fig. 5b). However, the key difference is that the peak of the N_i distribution shifts in the opposite direction, i.e. towards smaller N_i values with an increase in INP scaling factor (Figs. 5e, S2). Although the INP scaling factor has a stronger effect on the location of the peak N_i than CAP (Fig. S2), it shows no clear relationship with N_i^{median} values (Fig. 5b) because the spread of values is much larger and, due to the exponential nature of N_i -values, this means that the peak does not correspond to the mean. The peak for r_i shifts to smaller r_i values with a decrease in INP scaling factor, (Fig. 5f), but the spread is larger which is why this trend is less apparent in r_i^{median} (Fig. 5c).

We interpret these findings as follows: By design, increasing the INP scaling factor enhances the immersion freezing of cloud droplets, the deposition nucleation of ice, and the freezing of raindrops (Sect. 2), thereby accelerating the conversion of cloud droplets to ice. Similarly to CAP, a higher INP scaling factor leads to an earlier onset of full glaciation (see Fig. S1) with fewer cloud droplets reaching the homogeneous-freezing level, where they would otherwise form numerous small ice crystals. The explanation for the bi-modal distribution of N_i at low INP scaling factor values and the transition to a uni-modal distribution at higher values is therefore the same as for CAP. The only addition is that a smaller INP scaling factor shifts the location of the

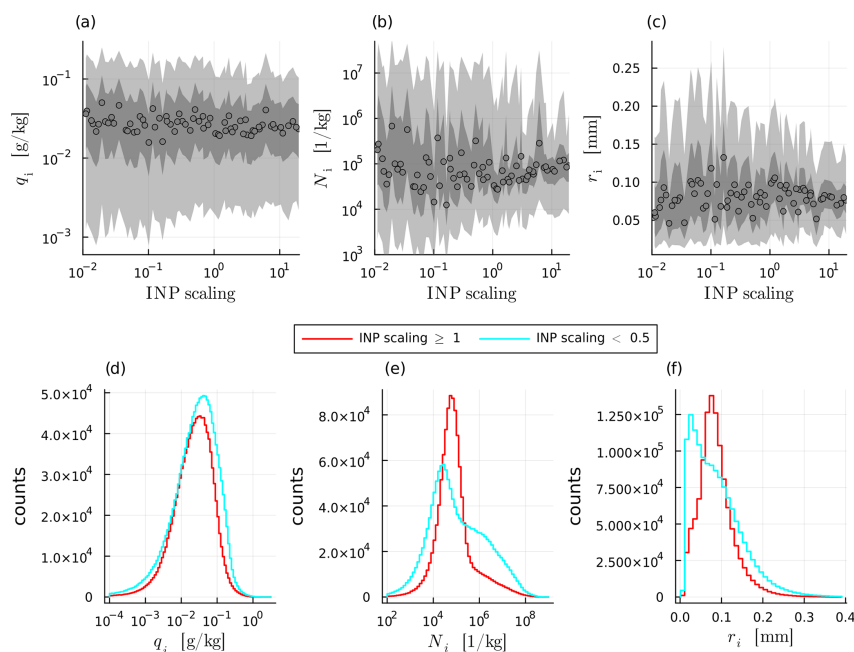


Figure 5. (a) q_i , (b) N_i and (c) ice radius (r_i), over INP, with markers showing median values and shaded areas the 5th to 95th percentiles (light grey) and 25th to 75th percentiles (dark grey). Panels (d), (e), and (f) show q_i , N_i and r_i respectively, but as histograms for all PPE members with INP scaling factors either larger than 1 (red; in total 29 PPE members and 1 123 557 trajectories) or smaller than 0.5 (cyan; in total 35 PPE members and 1 417 123 trajectories).

peak N_i to smaller values (a smaller CAP shifts it to larger values) and also exerts a stronger control on it (Fig. S2). If this peak is interpreted as those trajectories that encounter no (or less) homogeneous freezing of cloud droplets, then its location should primarily increase with INP scaling, which is what we observe. Furthermore, since hydrometeor growth by deposition is more efficient than by condensation, and a higher INP scaling factor prolongs the duration spent in the ice phase, a higher INP scaling factor ultimately promotes the formation of larger ice crystals, contributing to the observed increase in r_i and the slight decrease in q_i (heavier ice particles fall out more quickly). The tail towards larger r_i for small INP can be explained by the fact that when fewer ice crystals form during the ascent, there is less competition for available vapor, allowing for individual ice crystals to grow larger. This also enhances the conversion from ice to snow, which is supported by the fact that on average, q_s at the end of the ascent slightly increases when INP is small (not shown).

3.2.3 Impact of CCN on ice properties during the ascent

The third most important parameter (based on the IBF scores, Table 2) for the ice content at the end of the ascent is CCN. However, the CCN scaling factor primarily influences the hydrometeor content and mixed-phase microphysics during the WCB ascent as opposed to at the end. It correlates almost perfectly with the mean of maximum number mixing ratio

of hydrometeors achieved during the ascent ($R_s = 0.9994$, Fig. 6d). This clear dependency comes from the fact that an increase in the CCN scaling factor primarily increases the number of cloud droplets and reduces the number of rain drops during the ascent (Fig. 6e). With an increase in CCN, cloud droplets during the ascent become smaller; for simulations with the largest and smallest CCN scaling factor, the mean cloud droplet mass is approximately 72×10^{-9} and 9×10^{-9} g, respectively. Smaller cloud droplets delay the formation of rain and enhance the residence time of liquid condensate. This delay is the reason why air parcels begin their glaciation (defined as first time when $(q_c + q_r/q_{\text{tot}}) < 0.5$) at lower pressures and temperatures (i.e., later in the ascent) when the CCN scaling factor is larger (Fig. 6a and b). This in turn leads to a larger mean of the maximum mixing ratios of supercooled liquid water ($\text{max}(\text{SLW})$) achieved during the ascent with an increasing CCN scaling factor (Fig. 6c).

The mean of the maximum mixing ratio for snow ($\text{max}(q_s)$) during ascent also increases with CCN, while for graupel ($\text{max}(q_g)$) it decreases (Fig. 6e). We explain this as follows: increased CCN concentrations lead to more numerous but smaller cloud droplets, which slows the formation of raindrops. This, in turn, reduces the collision efficiency between ice or snow particles and cloud droplets, slowing the (continuous) conversion to graupel. As a result, snow remains more abundant and graupel less abundant. However, this large effect of the CCN scaling factor on the hydrome-

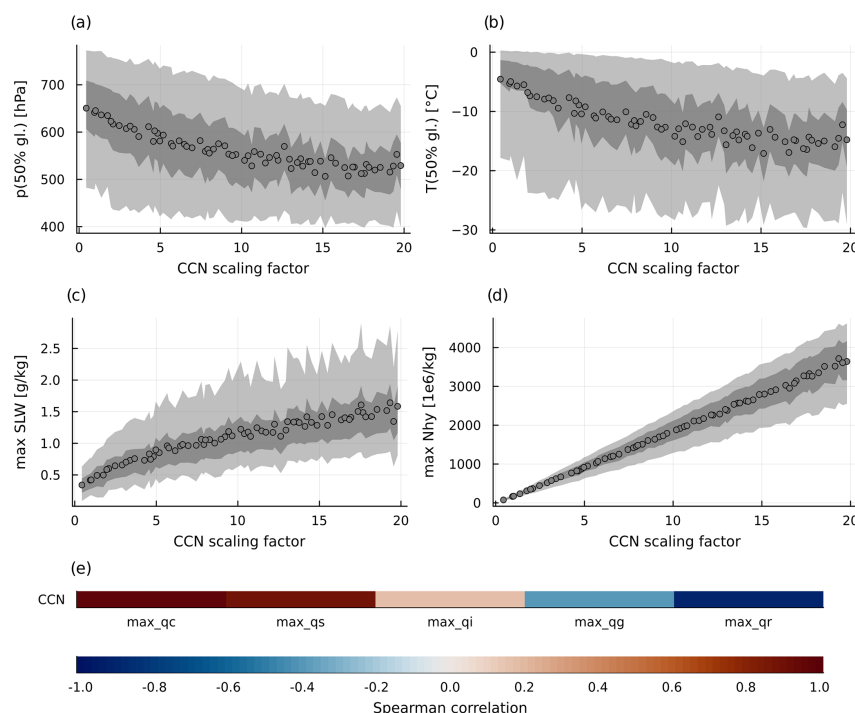


Figure 6. Scatter plots of (a) mean pressure at 50 % glaciation ($(q_i + q_s + q_g)/q_{\text{tot}} = 0.5$), (b) mean temperature at 50 % glaciation, (c) mean of maximum total hydrometeor mass mixing ratio during the ascent and (d) mean of maximum total hydrometeor number mixing ratio during the ascent over CCN scaling factors. Shaded areas show 5th to 95th percentiles (light grey) and 25th to 75th percentiles (dark grey). Spearman correlation coefficients for CCN scaling factors with individual mean of the maximum mass mixing ratios during the ascent for all hydrometeors are shown in (e).

teor population during the ascent is mostly lost by the end of the ascent, showing that once air parcels glaciate, CAP and INP dominate the ice-phase cloud microphysics.

3.2.4 (Non-)Impact of SST on ice properties

Another important result from the analysis of parameter effects on the hydrometeor population is that changes to SST, which substantially increase the amount of WCB trajectories that ascend quickly (Oertel et al., 2025), show no effect on distributions of N_i and q_i at the end of ascent. This absent effect is surprising, because Schwenk and Miltenberger (2024) found that fast ascending trajectories ($\tau_{600} < 5$ h) have a larger hydrometeor content at the end of the ascent than more slowly ascending trajectories ($\tau_{600} > 20$ h). Given that in the WCB case from this Paper, higher SSTs shorten ascent timescales and enhance convective activity (Oertel et al., 2025), we would have expected N_i and q_i to increase for simulations with higher SST. However, our findings indicate that the impact of CAP and INP dominates. A likely reason is that the WCB we examine is only weakly convective: only 5 % of trajectories have $\tau_{600} < 10$ h, and that fraction changes only slightly across the SST ensemble (4.5 % at -1 °C, 6.4 % at $+1$ °C). With so few strongly convective

parcels, SST-induced changes in vertical motion leave little imprint on N_i and q_i . We therefore hypothesize that, in a more convectively active WCB, an identical SST perturbation would produce a larger change in the convective fraction and, consequently, a clearer SST impact on N_i and q_i . Finally, we emphasize that changes in SST have no apparent effect on the temperature and pressure at which air parcels glaciate (see Fig. S7).

3.3 Effect of parameter perturbations on relative humidity over ice (RH_i) at end of ascent

Despite the strong temperature control on q_v values (Sect. 3.1), previous work suggests that substantial supersaturation over ice (up to 120 %) can exist in WCB outflow (Schwenk and Miltenberger, 2024; Spichtinger et al., 2005). In the PPE, we also find large RH_i values as well as a strong variability in RH_i distributions between PPE members (Fig. 3i and j). The parameters with the largest IBF scores (Table 2) for mean RH_i are CAP (0.319), INP (0.304) and CCN (0.229), and the $\text{RF}_{\text{mean}(\text{RH}_i)}^2$ model has a good predictive capability, with $R^2 = 0.765$ and $\text{NRMSE} = 0.481$. The PDPs show that increases in CAP, INP and CCN all work to decrease RH_i (see Fig. S3). The two-dimensional PDPs

(Fig. 7) additionally indicate that these parameters have a joint impact: an increase in CAP, INP and CCN is most effective in reducing RH_i^{mean} when the respective other two parameters are also increasing (lowest RH_i^{mean} values seen only in top-right corners of Fig. 7). Notably, these relationships seem non-linear when CAP, INP scaling and CCN scaling are particularly low or high, indicating that the distribution of these parameter values in the real-world are important for determining whether the overall sensitivity of RH_i to perturbations is large or small.

However, when considering the median RH_i values instead of the mean, the joint impacts of parameter perturbations turn out to be more nuanced. An increase in CAP values only reduces the median RH_i when the INP scaling factor is larger than one (red dots in Fig. 8a, $R_s = -0.72$). When the INP scaling factor is smaller than one, an increase in CAP slightly increases the median RH_i , but the correlation is weak (cyan dots in Fig. 8a, $R_s = 0.1$). The same is true the other way around: an increase in INP scaling factor increases the median RH_i when CAP is smaller than 0.5 (cyan dots in Fig. 8b, $R_s = 0.66$) and only shows a slight decrease in median RH_i when CAP is larger than 0.5 (red dots in Fig. 8b, $R_s = -0.28$). We also find that the spread of RH_i values reduces strongly with an increase in CAP and INP (shaded areas in Fig. 8a and b). The simulations with the smallest CAP and INP scaling have standard deviations in RH_i of 10.8 % and 9.8 %, and for the largest CAP and INP scaling this drops to 5.6 % and 6.8 %. This behavior is not seen for the CCN scaling factor, which reduces the median RH_i regardless of CAP and INP scaling and does not influence the spread (standard deviation of $RH_i \approx 9$ % for simulations with the largest and smallest CCN scaling; Fig. 8c). The fact that the mean, median and spread of RH_i values are all influenced differently by perturbations of the INP scaling factor and CAP is a strong indication that these parameters and their combination change the distribution of RH_i values between different PPE members.

Indeed, we find that RH_i distributions in PPE members with both large INP scaling factors and large CAP differ strongly from those with low values of both parameters (Fig. 9a). When accumulating PPE members where both $CAP > 0.8$ and the INP scaling factor > 1 (red histogram in Fig. 9a), we find a RH_i distribution that peaks at approximately $RH_i = 100$ % and is slightly skewed to larger values (tapers off at around $RH_i = 120$ %). However, the distribution of RH_i in PPE members with $CAP < 0.4$ and an INP scaling factor < 0.5 (cyan histogram in Fig. 9b) has a much sharper peak at RH_i slightly below 100 % and a much larger tail to large RH_i of 135 % and more. The mean and median RH_i for the first group (CAP and INP scaling large) are both approximately 103 %, whereas for the second group (CAP and INP scaling small), the mean RH_i of 106 % is larger than the median RH_i of 103 %. This explains both the differing behaviors of the mean and median RH_i with CAP and INP as well as the spread, which increases with the asymmetry

of the distribution. We note that the peak RH_i for small CAP and INP indicates that most trajectories in those PPE members are subsaturated at the end of the ascent. As we will see in Sect. 5, ice crystals sublimate in the outflow because of this subsaturation.

Before interpreting these findings, it is essential to note that in the ICON model, supersaturation with respect to water can only develop when an air parcel is fully glaciated, as this is the point at which the saturation adjustment scheme, which enforces relative humidity over water of 100 %, becomes inactive. Consequently, when examining the effects of parameter perturbations on RH_i at the end of ascent, it is important to focus on the timing of the air parcel's full glaciation, as well as the cloud microphysical evolution after this point, since only when there is no more liquid water will we see deviations from $RH_w = 100$ %. Note that $RH_w = 100$ % also constrains RH_i to a temperature dependent but otherwise fixed value (with $RH_i > 100$ %). For a fully glaciated air parcel, the relaxation timescale for supersaturation over ice (at a vertical velocity $w = 0 \text{ m s}^{-1}$) is inversely proportional to the product of ice number concentration and ice particle radius (e.g. Khvorostyanov, 1995; Khvorostyanov and Sassen, 1998; Khvorostyanov et al., 2001):

$$\tau_{\text{sat,ice}} = \frac{1}{4\pi \cdot CAP \cdot N_i \cdot r_i \cdot D_v} \quad (\text{with } [N_i] = \text{m}^{-3}), \quad (2)$$

with D_v the water vapor diffusion coefficient (taken as a constant value of $3 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$). Equation (2) shows that parameter perturbations that lead to higher N_i and r_i should lead to a faster reduction of supersaturation over ice, and therefore, given a similar vertical velocity distribution, on average smaller RH_i . This assumption is verified by the two-dimensional histogram of all RH_i and $\tau_{\text{sat,ice}}$ values (taken at the end of the ascent, because w must be approximately 0) from the PPE (Fig. 10). It reveals that when $\tau_{\text{sat,ice}}$ is smaller than approximately $5 \times 10^2 \text{ s}$, RH_i values are concentrated at or slightly below approximately 100 %. Above this threshold, the mean, median and spread of RH_i per $\tau_{\text{sat,ice}}$ -bin increase monotonically and plateau at $RH_i \approx 115$ % when $\tau_{\text{sat,ice}} > 10^4 \text{ s}$. The increase of RH_i past $\tau_{\text{sat,ice}} \approx 5 \times 10^2 \text{ s}$ shows that $\tau_{\text{sat,ice}}$ is the primary physical control of RH_i . RH_i at the end of the ascent also does not depend on RH_i at the moment of full glaciation, when the air parcels are still ascending (see Fig. S4) which solidifies that $\tau_{\text{sat,ice}}$ controls RH_i at the end of the ascent. Therefore, to interpret effect of parameter perturbations on RH_i , we must first understand their effects on $\tau_{\text{sat,ice}}$.

The median $\tau_{\text{sat,ice}}$ (Fig. 8d, e and f, note logarithmic y-axis) varies with CAP, INP and CCN parameter perturbations in a similar manner than RH_i (Fig. 8a, b and c): as indicated by the red and cyan lines in Fig. 8d and e, CAP slightly reduces $\tau_{\text{sat,ice}}$ when INP scaling is large ($R_s = -0.57$) and increases it when INP scaling is small ($R_s = 0.36$); the INP scaling factor slightly reduces $\tau_{\text{sat,ice}}$ when CAP is large ($R_s = -0.32$) and increases it when CAP is small ($R_s =$

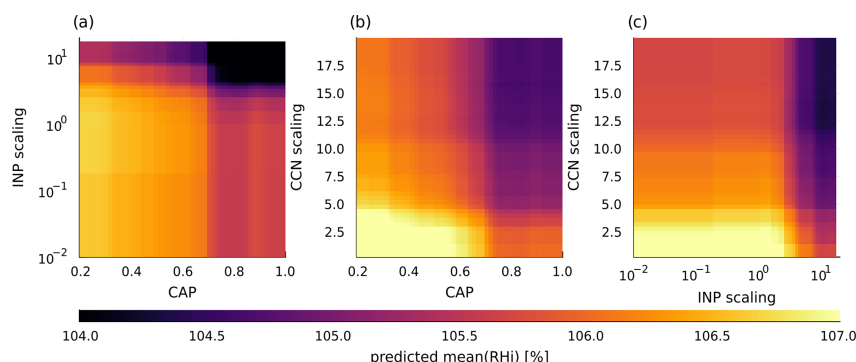


Figure 7. Two dimensional PDPs for $RF^2_{\text{mean}(RH_i)}$ for INP vs. CAP in (a), CCN vs. CAP in (b) and CCN vs. INP in (c).

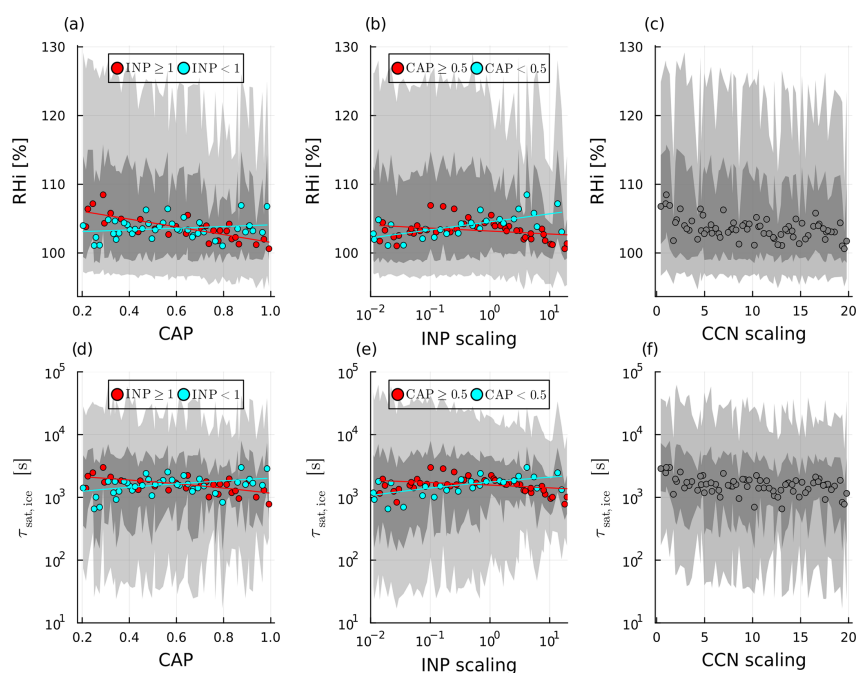


Figure 8. Scatter plots of RH_i^{median} over CAP (a), INP (b) and CCN (c) perturbations, with 25th–75th percentile range shaded in dark grey and 5th–95th percentile range shaded in light grey. Cyan and red points are masked according to INP in (a) and CAP in (b), with lines indicating a linear fit (can seem curved due to logarithmic axes) through the corresponding colored data points. The same in (d), (e) and (f) except for mean $\tau_{\text{sat,ice}}$.

0.62); and as indicated by Fig. 8f, the CCN scaling factor reduces $\tau_{\text{sat,ice}}$ regardless of CAP and INP scaling. As for RH_i , joint perturbations in the INP scaling factor and in CAP influence the distribution of $\tau_{\text{sat,ice}}$ (Fig. 9b). When $CAP > 0.8$ and INP scaling factor > 1 (red histogram in Fig. 9b, note the logarithmic x -axis and logarithmic spacing of bins), $\tau_{\text{sat,ice}}$ values peak at approximately 10^3 s and fall off symmetrically to either side. On the other hand, when $CAP < 0.4$ and $INP < 0.5$ (cyan histogram in Fig. 9b) the distribution is very broad and bi-modal, with the largest peak at a larger value of approximately 5×10^3 s and a second peak at a lower

value of approximately 5×10^2 s. As discussed above, $\tau_{\text{sat,ice}}$ -values below 5×10^2 s produce almost only RH_i -values of approximately 100 % (Fig. 10). The abundance of $\tau_{\text{sat,ice}}$ -values below this threshold for the group with small CAP and INP scaling (cyan histogram in Fig. 9b) therefore explains the sharp peak of the RH_i distribution (cyan histogram in Fig. 9a). The larger average value for $\tau_{\text{sat,ice}}$ and higher abundance of values above the threshold of 5×10^2 s for this same group of trajectories also explains the long tail to larger RH_i (cyan histogram in Fig. 9b), since RH_i increases strongly with $\tau_{\text{sat,ice}}$ beyond this threshold value (Fig. 10)

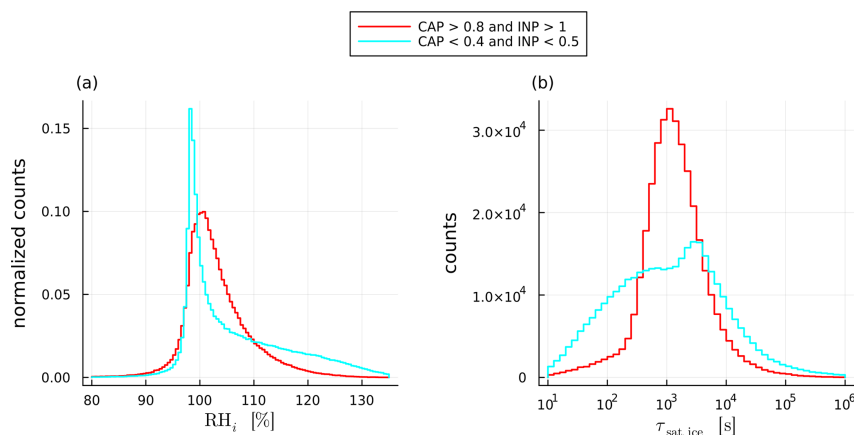


Figure 9. Histograms for (a) RH_i and (b) $\tau_{\text{sat,ice}}$ at the end of the ascent for all PPE members with $CAP > 0.8$ and INP scaling factor > 1 (red; in total 9 PPE members and 357 888 trajectories) or $CAP < 0.4$ and INP scaling factor < 0.5 (cyan; in total 9 PPE members and 363 512 trajectories).

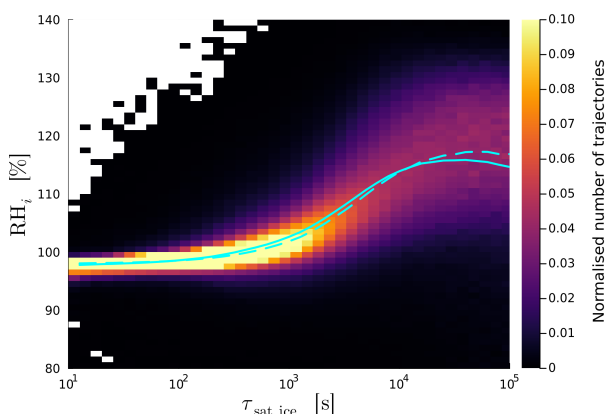


Figure 10. Two dimensional normalized histogram of RH_i over $\tau_{\text{sat,ice}}$ for all trajectories from all PPE members. Cyan lines show mean (solid) and median (dashed) values per $\tau_{\text{sat,ice}}$ -bin.

The effect that the INP scaling factor and CAP have on the distribution of $\tau_{\text{sat,ice}}$ (and thus RH_i) can be traced back to their effect on the hydrometeor populations at the end of the ascent (see Sect. 3.2 for physical interpretation), since $\tau_{\text{sat,ice}}$ is inversely proportional to N_i , r_i and CAP (Eq. 2). When the INP scaling factor or CAP is small, the distribution of N_i shows a large spread (Figs. 4b and e, 5b and e) and a bulge to larger N_i . This bulge in turn creates the bulge to smaller $\tau_{\text{sat,ice}}$ (Fig. 9b) and ultimately the sharp peak in RH_i (Fig. 9a). In addition, the larger spread in N_i also means there is a larger abundance of smaller N_i values, which together with the smaller r_i (Figs. 4c and f, 5c and f) explains the larger abundance of larger $\tau_{\text{sat,ice}}$ -values (Fig. 9b) and ultimately the long tail to larger RH_i (Fig. 9a). In contrast, when either the INP scaling factor or CAP are large, the N_i distribution becomes narrower and more symmetrical, which in

turn results in similarly shaped distributions for $\tau_{\text{sat,ice}}$ and ultimately RH_i .

The CCN scaling factor decreases the mean and median $\tau_{\text{sat,ice}}$ (Fig. 8f) but does not change the overall shape and spread of the distribution (see Fig. S5). The CCN scaling factor also has no notable effect on the hydrometeor populations at the end of the ascent (not shown), however the decrease in $\tau_{\text{sat,ice}}$ with increasing CCN scaling (which is weaker than the effect from INP scaling or CAP) must be due to a combined effect of CCN on N_i and r_i that does not appear in our analysis (possibly due to the time resolution of output data).

4 Impact of ascent timescale on parameter sensitivity

Schwenk and Miltenberger (2024) found that fast-ascending WCB trajectories transport significantly more hydrometeors into the UTLS and undergo distinctly different cloud microphysical processes (e.g. more riming, less deposition) compared to slower-ascending ones. Additionally, they demonstrated that precipitation formation pathways in convectively influenced WCB parcels differ markedly from those in slantwise-ascending parcels. Consequently, fast ascending trajectories likely play a key role in modulating moisture transport, which raises the question of whether the effects of parameter perturbations depend on the ascent timescale. In accordance with Oertel et al. (2025), we define trajectories with $\tau_{600} < 10$ h as fast ascending and repeat the analysis in the proceeding sections for this subset of WCB trajectories. We only briefly summarize how conditions at the end of the ascent change with ascent time, as they are consistent with findings by Schwenk and Miltenberger (2024).

Fast ascending trajectories ascend to lower temperatures and pressures, i.e., to higher altitudes: The average across all PPE members for the mean temperature and pressure are

$\sim -43^\circ\text{C}$ and $\sim 330\text{ hPa}$, respectively. If only fast ascending trajectories are considered, these values reduce to -47°C $\sim 285\text{ hPa}$. On average, q_i^{mean} and N_i^{mean} are also much larger for fast trajectories (0.12 g kg^{-1} and $1.6 \times 10^7\text{ kg}^{-1}$) than for all trajectories (0.04 g kg^{-1} and $0.2 \times 10^7\text{ kg}^{-1}$), and the mean $\text{RH}_i^{\text{mean}}$ at the end of ascent for fast trajectories is closer to saturation (100.5%) than for all trajectories (105.7%). The mean r_i^{mean} is smaller for fast trajectories (0.05 mm) than for all trajectories (0.09 mm). Fast trajectories also glaciate much later (at pressures and temperatures of $\sim 360\text{ hPa}$ and $\sim 241\text{ K}$) than all trajectories ($\sim 485\text{ hPa}$ and $\sim 253\text{ K}$, Fig. 11a and b), which is consistent with the findings from Schwenk and Miltenberger (2024).

The effects we observed for perturbations in the INP scaling factor (Fig. 5) and CAP (Fig. 4) for all trajectories are also seen for fast ascending trajectories, only that they are much more pronounced (Fig. 12 for fast trajectories vs. Figs. 4 and 5 for all trajectories; summary statistics as in Figs. 4 and 5a, b and c are not shown). The largest effect is seen for N_i : when the INP scaling factor and/or CAP are small, the second peak towards higher N_i in the bimodal distribution becomes the largest, effectively shifting the peak N_i by two orders of magnitude (from $\sim 10^5$ to $\sim 10^7\text{ kg}^{-1}$, Fig. 12b and e). This shift means that a larger proportion of fast trajectories contain homogeneously freezing cloud droplets compared to slower ascending trajectories. This proportion increases strongly when CAP and the INP scaling factor are small.

Additionally, r_i decreases (Fig. 12c and e) and q_i increases (Fig. 12a and d) more strongly with decreasing CAP and INP for fast ascending trajectories than for all trajectories (Figs. 4f and 5f). Even though r_i is smaller for fast trajectories (which should increase $\tau_{\text{sat,ice}}$), the much larger N_i -values (independent of the choice for INP scaling factor or CAP) lead to substantially smaller $\tau_{\text{sat,ice}}$ -values for fast ascending trajectories (compare Fig. 13b to Fig. 9b). The large shift in N_i values with INP scaling and CAP also induces a larger difference in $\tau_{\text{sat,ice}}$ -values for different INP scaling factors and CAP (Fig. 13b). However, the effect on the RH_i distribution (Fig. 13a) is small, because the peak $\tau_{\text{sat,ice}}$ for fast trajectories (Fig. 13b) is below the threshold of $5 \times 10^2\text{ s}$, below which RH_i stays mostly smaller than 100% (Fig. 10), regardless of INP scaling and CAP. Therefore, fast ascending trajectories reduce ice supersaturation much faster than all trajectories due to their high ice-number mixing ratio N_i .

We summarize that for fast ascending trajectories the INP scaling factor and CAP greatly influence the hydrometeor population at the end of the ascent, but that this has only a small effect on RH_i because large very large N_i -values lead to small $\tau_{\text{sat,ice}}$ in almost all cases. We also note that for fast ascending trajectories, N_i and r_i are more sensitive to CAP and the INP scaling factor than for all trajectories, suggesting that overall sensitivities to these parameters might increase for higher resolution simulations, where a greater fraction of trajectories ascend quickly.

SAT and the CCN scaling factor have very little impact on the outflow properties of the bulk of WCB trajectories, but become important when considering only the sub-set of fast ascending trajectories. SAT becomes important for the pressure, temperature and specific humidity at the end of the ascent (Fig. 14). While the distributions of these variables do not change visibly with SAT (not shown), the mean and median, as well as the upper percentiles all decrease linearly with an increase in SAT. This decrease indicates that modifications to the saturation adjustment scheme, which permit supersaturation under sufficiently high vertical velocities, can influence the thermodynamic conditions at the end of ascent when vertical velocities during the ascent are high enough. However, this effect is also a reflection of the fact that the fast ascending trajectories glaciate much later (Fig. 11), meaning that the saturation adjustment scheme is active for a longer portion of the ascent. Therefore, fast ascending trajectories spend more of their ascent at $\text{RH}_w = 100\%$, or slightly above that value if the SAT factor is large. Thus, more vapor, that would otherwise have condensed, is transported higher into the atmosphere, resulting in a delayed latent heat release that provides some additional buoyancy later in the ascent. However, even for fast trajectories SAT still does not influence RH_i , because fast ascending trajectories are also mostly glaciated by the end of the ascent. Additionally, $\tau_{\text{sat,ice}}$ -values are so small, that supersaturation is quickly reduced regardless of SAT. Therefore, SAT can change the pressure and temperature at the end of the ascent for fast ascending trajectories, but is unable to modify the specific humidity values which remain constrained by the thermodynamic conditions.

For fast ascending trajectories, the CCN scaling factor becomes relevant for r_i , N_i , q_i , $\tau_{\text{sat,ice}}$ and RH_i at the end of the ascent (Fig. 15). However, the shape of distributions is only changed for N_i . The effects during the ascent are similar to those discussed in Sect. 3.2 (not shown). With an increase in CCN, the amount of cloud droplets during the ascent increases (Fig. 6d), and fast ascending trajectories have more cloud droplets that freeze homogeneously (Fig. 12b and e). Therefore, when the CCN scaling factor is large and ascent velocities are high, even more cloud droplets freeze homogeneously, decreasing r_i at the end of the ascent (Fig. 15c) and increasing N_i (Fig. 15b) as well as q_i (Fig. 15a). The orders of magnitude larger N_i and only slightly different r_i implies smaller $\tau_{\text{sat,ice}}$ (Eq. 2; Fig. 15c and d) and therefore smaller RH_i (Fig. 15e), which is on average 102% when CCN scaling < 5 and 99% when CCN scaling > 15 . Overall, RH_i changes only slightly across PPE members for fast ascending trajectories, but among these changes, CCN scaling factor perturbations produce the largest effect. This is because for small CCN, N_i is reduced enough to produce $\tau_{\text{sat,ice}}$ values slightly above the threshold value of $5 \times 10^2\text{ s}$.

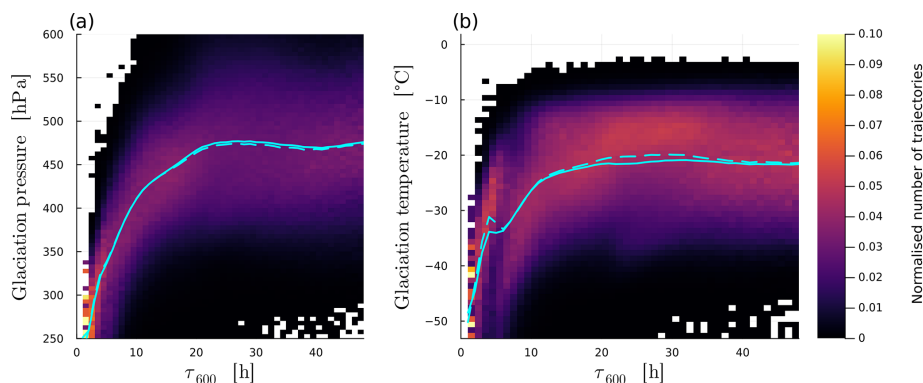


Figure 11. 2-dimensional histograms of pressure in (a) and temperature in (b) at glaciation (defined as the first 60 min time step for which q_c is zero; trajectories with non-zero q_c are excluded) over the fastest 600 hPa-ascent time τ_{600} . Mean and median are shown by the solid and dashed line, respectively.

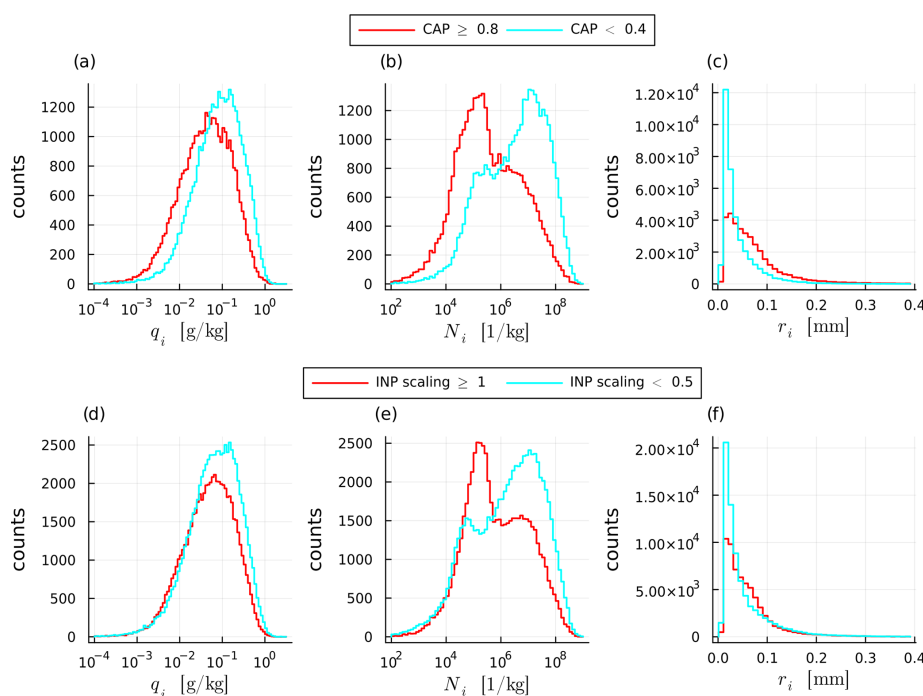


Figure 12. Histograms of q_i (a, d), N_i (b, e) and r_i (c, f) only for fast ascending trajectories and for subsets of PPE members sorted according to CAP (a–c) and the INP scaling factor (d–f).

5 Effects of parameter perturbations on cirrus properties 5 h after end of ascent

The conditions at the end of the ascent of a parcel likely influence but are not necessarily representative for those in the hours afterwards. Hydrometeors may precipitate out of the parcel and/or grow to reduce supersaturation (or evaporate/sublimate in case of sub-saturated conditions). RH_i values 5 h after the end of the ascent have a large spread (not shown) due to the subsidence of a large portion of trajec-

tories (37 % of trajectories across all PPE members have a pressure at least 10 hPa higher than at the end of the ascent). We are interested in those that could retain large RH_i and therefore consider only those that have no net-descent after 5 h (~ 50 % of all WCB trajectories across all PPE members). For this subset of trajectories, we consider the impact of the parameter perturbations on q_i^{mean} , N_i^{mean} and RH_i^{mean} 5 h after the end of ascent. The two most important parameters according to the IBF score are still CAP and the INP scaling factor (Table 3) and the RF_X^2 -predictions are very good,

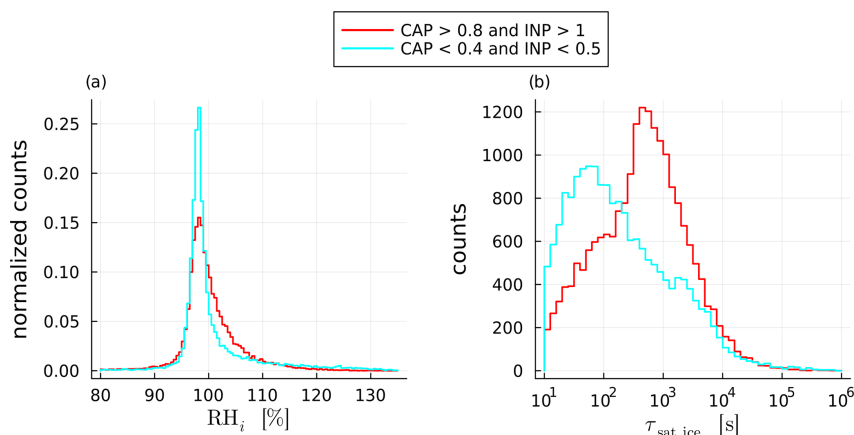


Figure 13. Histograms for (a) RH_i and (b) $\tau_{\text{sat,ice}}$ at the end of the ascent for only the fast ascending trajectories and PPE members with both $CAP > 0.8$ and INP scaling factor > 1 (red) or $CAP < 0.4$ and INP scaling factor < 0.5 (cyan).

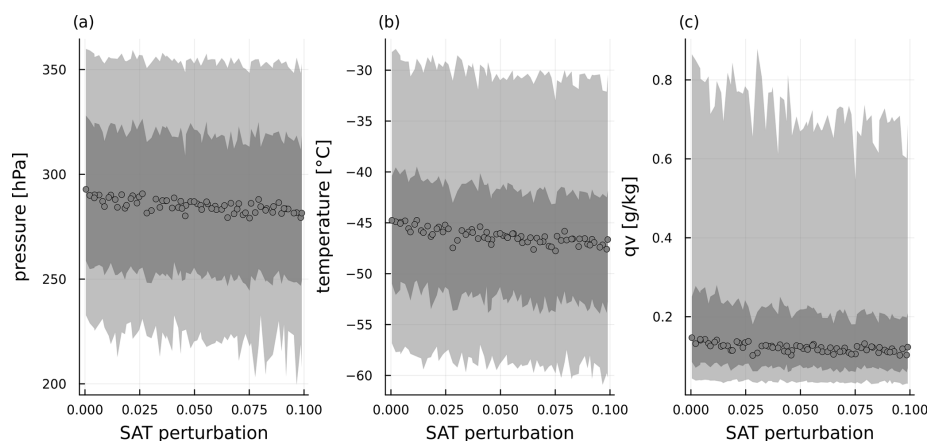


Figure 14. Pressure in (a), temperature in (b) and specific humidity q_v in (c) at the end of the ascent for only the fast ascending trajectories over the perturbation of the saturation adjustment scheme (SAT). For plots (a) and (b) the dots show the mean per PPE member, in plot (c) the dots show the median. Shaded areas indicate the inter-quartile range (dark grey) and 5th to 95th percentile (light grey).

with R^2 and NRMSE of approximately 0.8–0.9 and 0.3–0.5, which supports this statement.

Apart from the deposition nucleation, the INP scaling factor only effects processes in the mixed-phase. Therefore, as long as there is a negligible amount of deposition nucleation, any effect seen from it after the ascent is a reflection of the conditions at the end of the ascent. N_i 5 h after the ascent increases with the INP scaling factor (Fig. 16a), which is probably a reflection of the larger N_i found at the end of the ascent (Fig. 5e). However, at the end of the ascent and for small INP scaling factors, N_i has a bimodal distribution (Fig. 5e), which is no longer seen 5 h afterwards (Fig. 16a). One would presume that trajectories with large N_i and small r_i at the end of the ascent retain the ice for a long time due to the lower sedimentation velocity of smaller ice crystals. However, the absence of higher median N_i 5 h after the ascent suggests that these small ice crystals sublimate. The alternative ex-

planation to sublimation is collisional growth, which is proportional to the particle number mixing ratio ($\sim 10^5 \text{ kg}^{-1}$), which is much smaller than in a typical mixed-phase cloud ($\sim 10^8 \text{ kg}^{-1}$), and the sticking efficiency, which is temperature dependent and very small at these low temperatures (Seifert and Beheng, 2006, 2001). Therefore, we find sublimation the most likely explanation, given the large fraction of trajectories that are sub-saturated with respect to ice during this time (Fig. 16c). Indeed, only the subset of trajectories with RH_i close to 100 % 5 h after the ascent still have the bimodal distribution in N_i that supports N_i values larger than 10^6 kg^{-1} (see Fig. S6). Some small ice crystals also persist when CAP is small (increased number of high N_i values for cyan line Fig. 16b), which would explain why CAP is more important for N_i^{mean} 5 h after the ascent even though the peak N_i is not shifted.

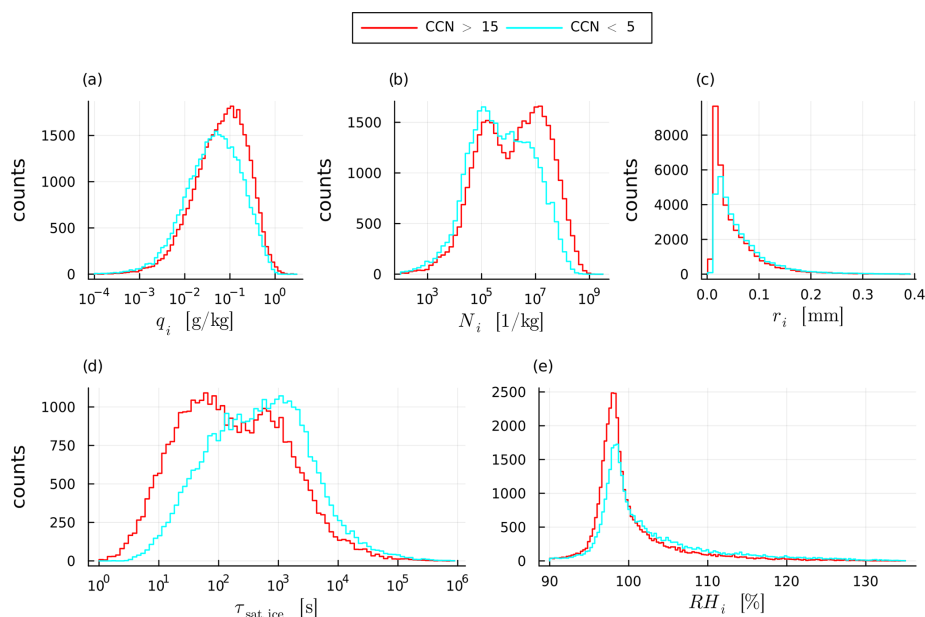


Figure 15. Histograms for q_i in (a), N_i in (b), r_i in (c), $\tau_{\text{sat,ice}}$ in (d) and RH_i in (e) at the end of the ascent for only the fast ascending trajectories and PPE members with a CCN scaling factor larger than 15 (red; 17 PPE members and 686 947 trajectories) or smaller than 5 (cyan; 17 PPE members and 627 912 trajectories).

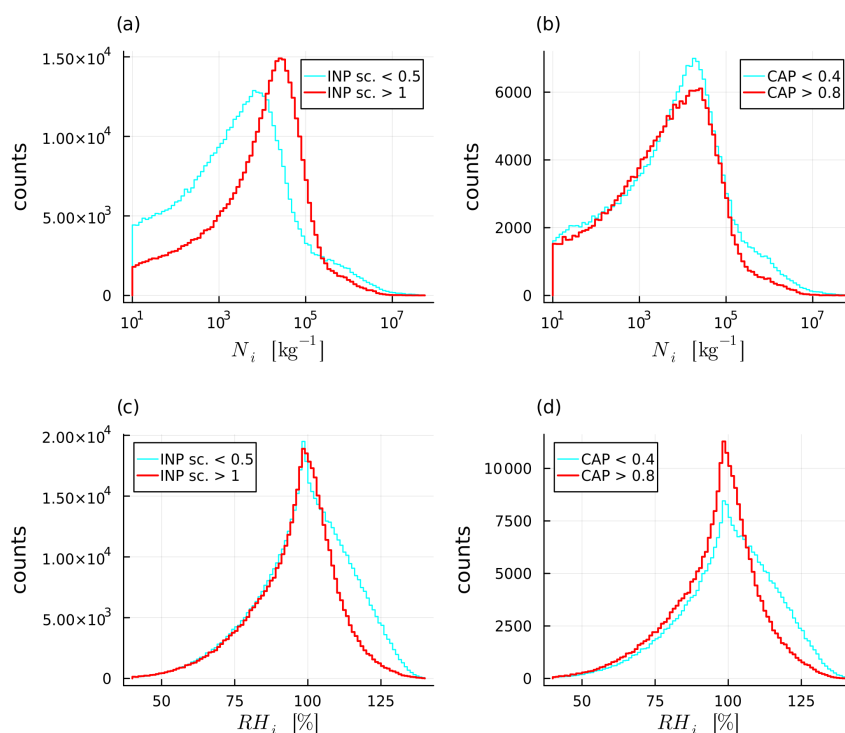


Figure 16. Histograms for N_i 5 h after the end of ascent, split according to the INP scaling factor in (a) and according to CAP in (b). The same for RH_i in (c) and (d). Only trajectories that do not descend after the end of the ascent are included.

Table 3. Impurity-based feature (IBF) importance scores of CAP, INP and CCN derived from RF_X^1 for $\text{RH}_i^{\text{mean}}$, q_i^{mean} and N_i^{mean} 5 h after the end of ascent. Only scores ≥ 0.1 are included. The bottom rows show the R^2 and NRMSE value for the predictions of the second iteration of the forest regression model (RF_X^2) using only parameters with IBF ≥ 0.1 . Note: all of these values are slightly different each time a forest model regression is performed; the values shown here are averaged over 100 iterations.

	mean(RH_i)	mean(q_i)	mean(N_i)
CAP	0.291	0.540	0.524
INP	0.461	0.206	0.266
CCN			
SST	0.109	0.103	
SAT			
R^2	0.849	0.833	0.758
NRMSE	0.385	0.406	0.488

For RH_i , CAP and the INP scaling factor are almost equally important (Table 3). They also have a very similar effect of increasing the tail of RH_i -values 5 h after the ascent to larger values while leaving the peak at $\sim 100\%$ unchanged (Fig. 16). For the INP scaling factor, this effect is a reflection of different N_i 5 h after the ascent (smaller N_i means slower reduction of supersaturation) and for CAP the signal simply occurs because larger CAP values reduce supersaturation faster.

6 Conclusions and Discussion

Warm-conveyor belts (WCBs) are a key feature for extra-tropical transport of humidity from the lower troposphere to the tropopause region. Due to the strong ascent and associated cloud formation this transport is modulated by cloud microphysical processes, the representation of which suffers from well-documented uncertainties in state-of-the-art weather and climate models. In this paper we therefore analyze Lagrangian trajectory data from a perturbed parameter ensemble (PPE) to answer the following research questions: (How) do perturbations in microphysical parameters and SST (i) influence thermodynamic and moisture conditions in fresh WCB outflow, (ii) alter hydrometeor distributions in fresh WCB outflow, (iii) impact the moisture and ice properties in aged WCB outflow, and (iv) how these effects vary with ascent velocity?

The distributions of T , p and q_v at the end of the ascent (research question (RQ) i) hardly vary across PPE members (Fig. 3a–f). This rigidity indicates that the thermodynamic conditions at the end of the ascent are constrained by the larger-scale environment and that parameter perturbations do not substantially influence their magnitude or distribution. In contrast, the RH_i distribution and its mean value varies quite strongly with INP scaling factor and CAP: When the INP

scaling factor and CAP are small, the RH_i distribution has a sharp peak below 100% and a long tail to larger values (Fig. 9a). If the INP scaling factor and CAP are large, the distribution is more Gaussian-shaped with a peak at about 102% . We have shown that the differences in the RH_i distribution are caused by the impact of INP scaling factor and CAP on the ice crystal population, which is connected to RH_i by the timescale for super-saturation depletion $\tau_{\text{sat,ice}}$ (Eq. 2, Fig. 10).

The hydrometeor population at the end of ascent (RQ ii) is dominated by ice particles. The largest variation in ice mass and number mixing ratio are induced by the INP scaling factor and CAP: For a small CAP and INP scaling factor, the N_i distribution hints at bimodality with a second mode at large N_i values, that vanishes for larger CAP and INP scaling factors. This behavior can be explained by the more efficient cloud glaciation with large CAP values and INP scaling factor, which strongly reduces the number supercooled droplets reaching the homogeneous freezing level. For enhanced INP scaling factors we additionally detect a small shift for smaller N_i values, which correlates directly with the number of heterogeneously formed ice crystals. The ice crystal mass mixing ratio decreases with increasing CAP (and to a lesser extent with increasing INP scaling factor) due to the larger size of ice crystals and more efficient sedimentation.

Interestingly we find no significant impact of CCN scaling on the ice crystal population at the end of ascent, although it substantially modulates the hydrometeor population and onset of substantial glaciation during the ascent. However, this signature does not propagate to the fully glaciated state at the end of ascent. Also we do not find an impact of SST or SAT perturbations on the outflow ice crystal population. This absence of an effect is somewhat surprising, as Oertel et al. (2025) found altered fractions of fast ascending trajectories and Schwenk and Miltenberger (2024) showed that fast ascending trajectories contain more ice at end of ascent. However, the fraction of fast ascending trajectories is overall very small and therefore the impact on the overall distribution of q_i and N_i in the outflow are small.

Considering only fast-ascending trajectories (RQ iv), CAP and the INP scaling factor remain the most important parameters for hydrometeor content and RH_i at the end of the ascent (Figs. 12b and e, 13b), albeit with the a less pronounced impact on the RH_i distribution than for all trajectories. This small impact on RH_i is likely a consequence of the fast ascent and short timescale for glaciation, with some super-cooled liquid reaching the homogeneous freezing temperature irrespective of the choice of CAP and INP scaling factor. N_i and r_i are more sensitive to CAP and the INP scaling factor for fast ascending trajectories than for all trajectories.

This sensitivity indicates that simulation scale might also influence N_i and r_i , since Choudhary and Voigt (2022) found that a greater fraction of WCB trajectories ascend quickly when model resolution is increased. In contrast to the consideration of all trajectories, perturbations of the CCN scal-

ing factor have small yet noticeable impact on q_i , N_i and RH_i distributions, with large CCN scaling factors shifting those distributions to smaller values. We also find a small impact of SAT perturbations on the thermodynamic conditions (p , T) and specific humidity q_v at the end of ascent (Fig. 14). The larger impact of SAT for fast-ascending trajectories is due to more substantial super-saturation being possible at larger vertical velocities and the larger fraction of the ascent spend in mixed-phase conditions, where water vapor-condensate partitioning is controlled by saturation adjustment. Therefore, altering SAT impacts buoyancy more strongly and enables a deeper ascent with an increasing SAT perturbation.

The ice particle population 5 h after the end of the ascent (RQ iii) is impacted mostly by the choice of CAP and the INP scaling factor. In comparison to end of ascent (RQ ii), the large N_i values (associated with small r_i) are not present anymore and crystals from the small N_i mode dominate. Consequently the INP scaling factor is more important than CAP 5 h after the end of ascent, while they were of similar importance at the end of ascent.

The analysis presented here is based on the PPE from Oertel et al. (2025), which focused on the dynamical impact of parameter perturbations, i.e. on the role for diabatically induced potential vorticity anomalies. Some earlier studies have also provided insight into the role of uncertainty in the microphysical parametrization for WCBs, but in contrast to our paper have predominantly focused on the impact on diabatically induced PV anomalies, upper-level flow and its predictability. In particular, Joos and Forbes (2016) investigated differences between two versions of the ECMWF global model (IFS) dominated by changes in the cloud microphysics scheme. Similarly, Mazoyer et al. (2021, 2023) explored the impact of using different cloud microphysics schemes available in the French Mesoscale Non-Hydrostatic model (MesoNH) in convection-permitting simulations. Pickl et al. (2022) investigated the effect of Stochastically Perturbed Parameterization Tendencies (SPPT) on WCBs using the IFS model at 100 km resolution, running simulations over multiple model years. In addition, recent studies have proposed a concept for objectively selecting impactful parameter perturbations (Hieronymus et al., 2022; Neuhauser et al., 2023). While their results partly align with the subjective parameter selection in our PPE, guidance from their results could be used to expand the investigation in future studies.

Our results suggest that in ICON's two-moment cloud microphysics scheme, the hydrometeor populations and RH_i -conditions in WCB outflow are sensitive to the parameter choices in the cloud microphysics scheme, specifically those relevant for microphysical processes in the ice phase (CAP and INP scaling factor). However, this study consists of only one case study using one microphysics scheme, and it isn't clear how representative our findings are of WCBs (or other model configurations and microphysics schemes) in general. Additionally, PPEs sample the uncertain multi-

parameter phase space but assume constant parameter values throughout the simulation for some parameters that are not constant in reality. Therefore, sensitivities can be overestimated relative to a potentially more realistic stochastic framework in which parameters vary during the simulation (e.g. Stanford et al., 2019). Furthermore, the microphysics scheme employed in the ICON simulation only considers rime splintering; it does not account for raindrop fragmentation upon freezing or ice collisional breakup for secondary ice production. This could result in biases in mixed-phase ice concentrations. Finally, there are likely additional important parameters that influence the variables examined in this study, but that were not perturbed (see Hieronymus et al., 2025). Nevertheless, the findings from this study lead to the following implications and practical recommendations for modeling, observation, and future research:

- The effects identified in this study could inform decisions for microphysical parameter choices in numerical models or parameters to be perturbed in ensemble prediction systems.
- Comparison of the distribution of N_i , q_i , r_i and RH_i documented here and distributions from large observational data-sets may help in selecting appropriate values for CAP and the INP scaling factor for WCB flow configurations.
- Future modeling studies on WCB moisture transport should keep CAP and the INP scaling factor, but reserve additional resources to investigate other potentially important factors such as the representation of graupel and ice sedimentation velocity or collision efficiencies of ice-phase particles.
- Our findings may have implications for geoengineering strategies such as cirrus cloud thinning (Lohmann and Gasparini, 2017), which aim to modify processes similar to those affected by the INP scaling factor. Since N_i 5 h after the ascent increases with the INP scaling factor 5 h after the ascent, accelerating these processes could pose a risk of “overseeding”, as discussed in Gasparini et al. (2020).
- Following this, future studies could investigate how the different UTLS moisture conditions produced by the PPE members influence the radiative effect of WCBs.
- The vertical water transport in WCBs is an important contribution to the water content and supersaturation distribution in the extra-tropical UTLS. Future studies should examine the impact of the changes found here for the occurrence of extra-tropical ice supersaturated regions, which impact downstream in-situ cirrus formation including aviation-induced cirrus.

Appendix A: Additional Figures

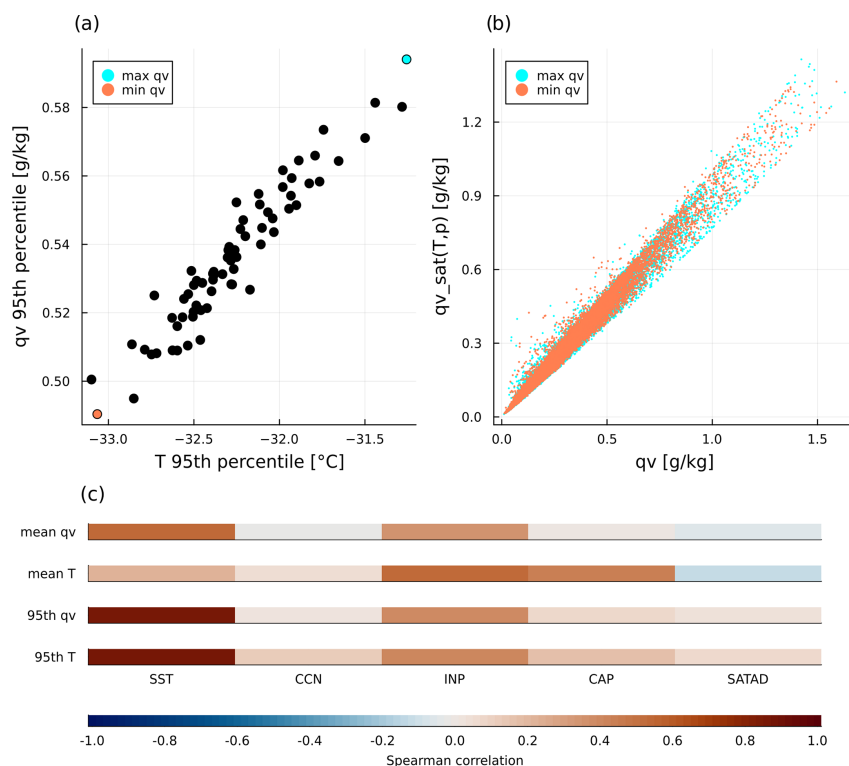


Figure A1. (a) Scatter-plot of 95th percentiles of q_v over T per ensemble member (dots). The ensemble members with the highest (lowest) 95th percentile of q_v values are plotted in cyan (orange). (b) Scatter-plot of saturation specific humidity (calculated using temperature and pressure of parcel) against specific humidity q_v for these two ensemble members. Panel (c) shows the Spearman correlation of mean and 95th percentile for q_v and T with all perturbed parameters.

Code and data availability. The ICON source code is distributed under an institutional licence issued by the German Weather Service (DWD). The code and data necessary to recreate the results shown here are available from <https://doi.org/10.5281/zenodo.17179532> (Schwenk et al., 2025). The data repository for Oertel (2025) is available in RADAR4KIT from <https://doi.org/10.35097/ecgs4f56mp3ymjmt> (Oertel, 2025). Further data that support the findings of this study are available from the authors upon reasonable request.

Supplement. The supplement related to this article is available online at <https://doi.org/10.5194/acp-25-11333-2025-supplement>.

Author contributions. AO set up and conducted the simulations for the PPE and provided the pre-selected Lagrangian WCB trajectory data. AM and AO developed the concept of this study. CS conducted/performed the post-processing calculations, the statistical analysis, and visualized the results. All three authors were involved in the discussion and interpretation of the results.

Competing interests. At least one of the (co-)authors is a member of the editorial board of *Atmospheric Chemistry and Physics*. The peer-review process was guided by an independent editor, and the authors also have no other competing interests to declare.

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Microphysical parameter choices modulate ice content and relative humidity in the outflow of a warm conveyor belt

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S1 Additional Plots

Here we present some additional plots that were left out of the publication for the sake of brevity.

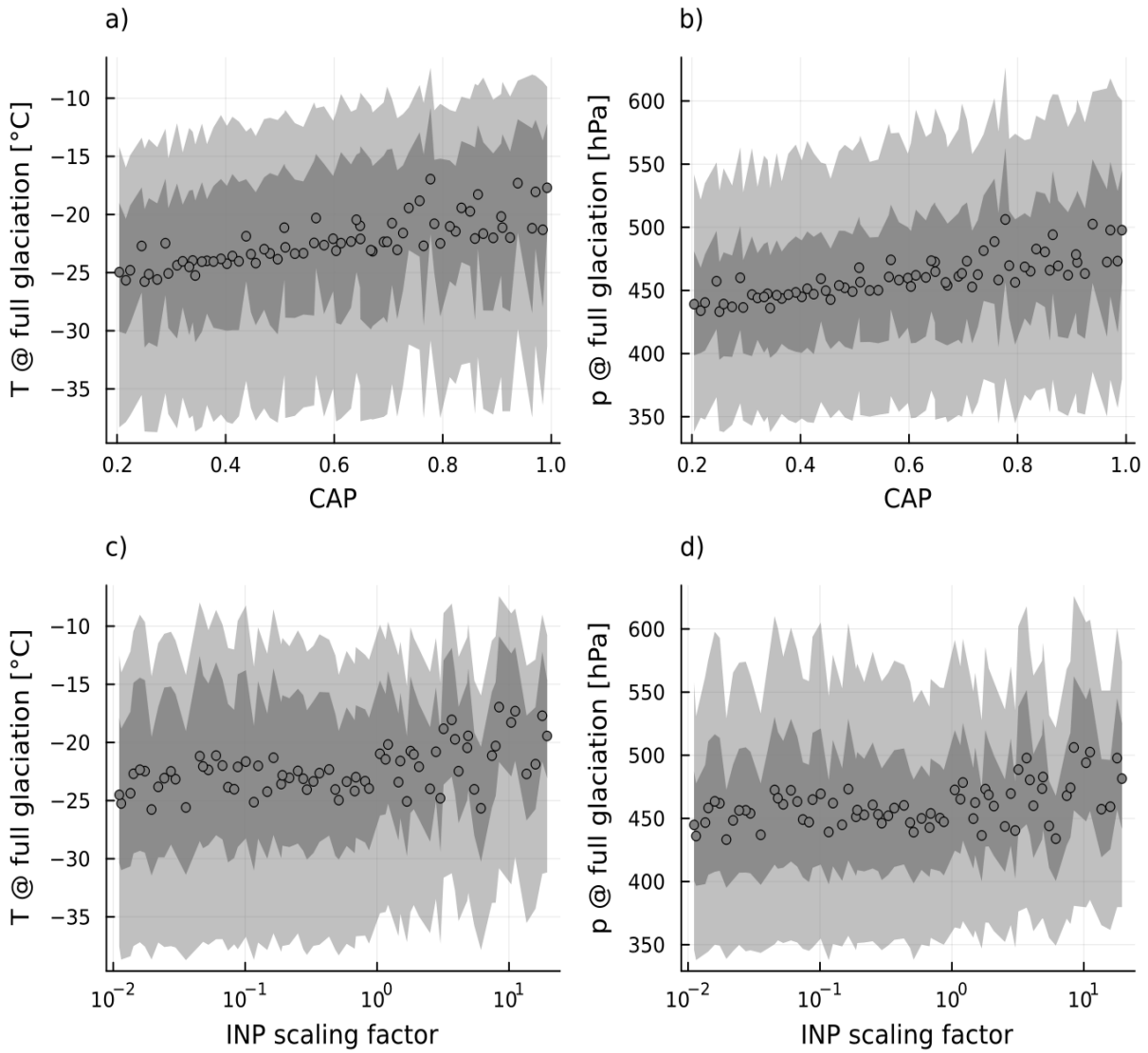


Figure S11. Scatter plots of the mean T (a and c) and p (b and d) at the first full glaciation over CAP (a and b) and INP scaling factor (c and d). Shaded regions show 5th to 95th percentile range (light grey) and interquartile range (dark grey).

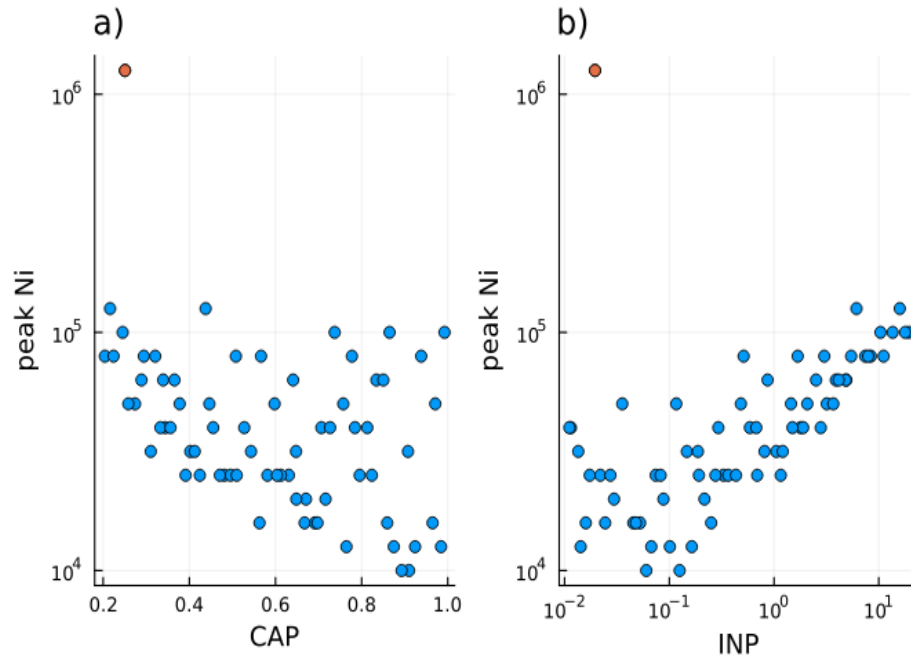


Figure SI2. Scatter plots of the location of the peak of the N_i -distribution at the end of the ascent over CAP (a) and INP (b). For the orange outlier at both low CAP and INP the second higher concentration mode in the bimodal distribution is the highest.

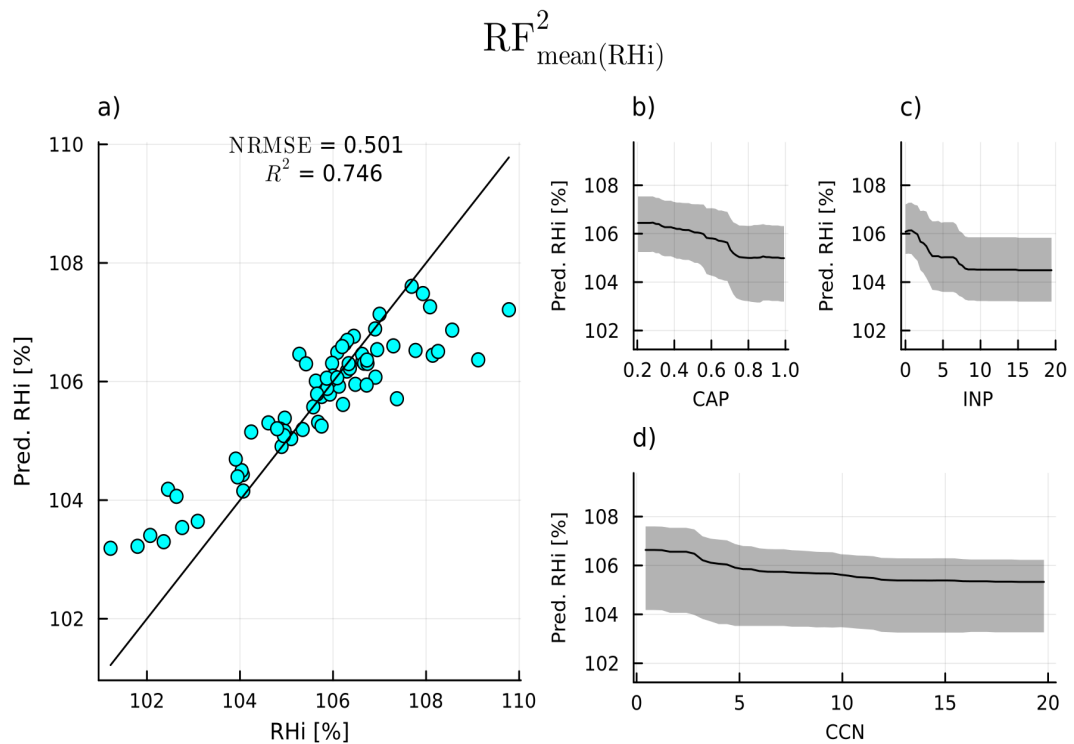


Figure SI3. a) prediction of RF2 mean(RH_i) over actual mean(RH_i), b), c) and d) show PDPs for CAP, INP and CCN with shaded areas indicating spread of predictions.

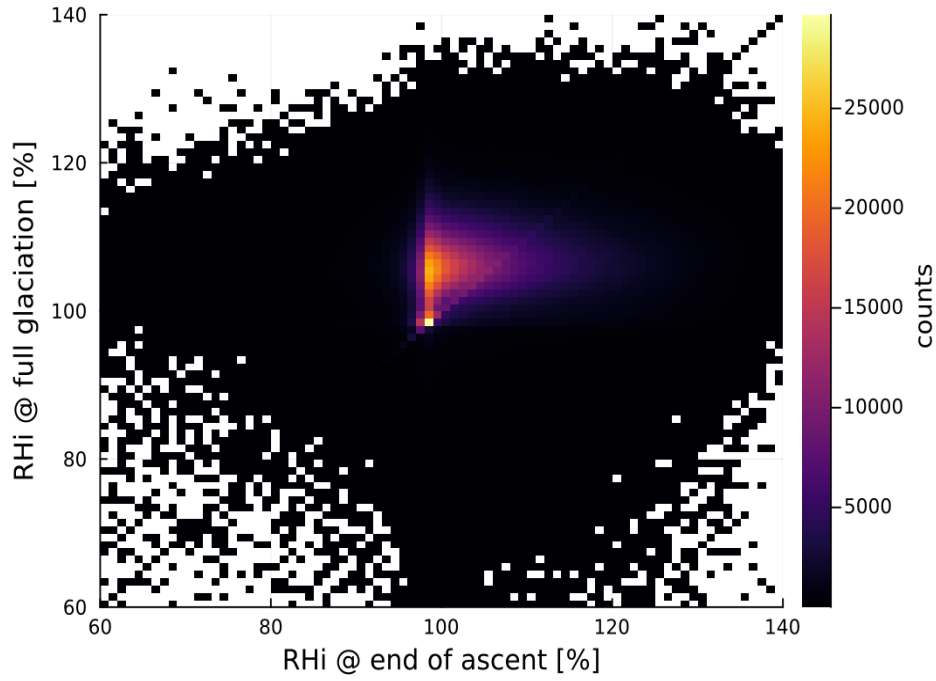


Figure SI4. 2 dimensional histogram of RH_i at the moment of full glaciation over RH_i at end the of the ascent showing no correlation.

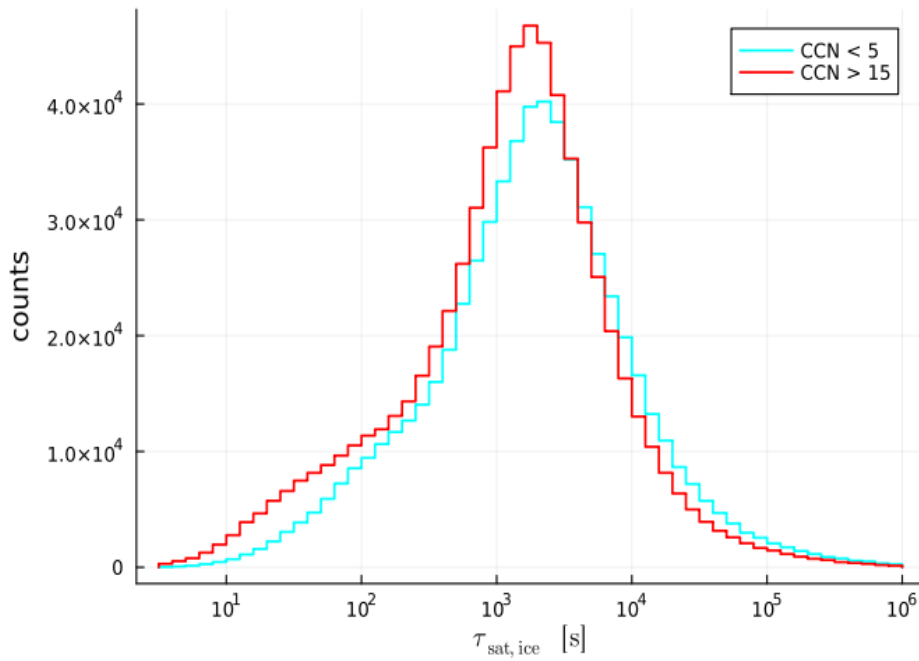


Figure SI5. Histogram of $\tau_{sat,ice}$ values for two subsets of trajectories, one with a CCN scaling factor larger than 15 (red) and one smaller than 5 (cyan).

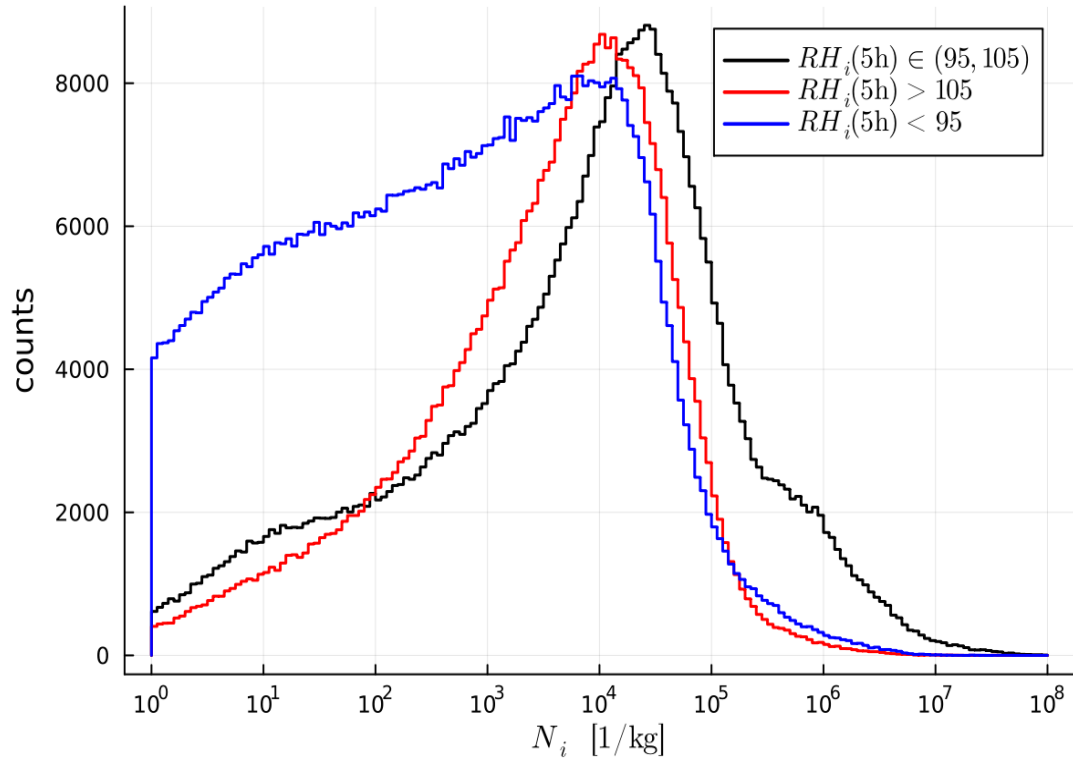


Figure SI6. Histogram of N_i values 5 hours after the end of the ascent for trajectories across all PPE members sorted according to their RH_i 5 hours after the ascent.

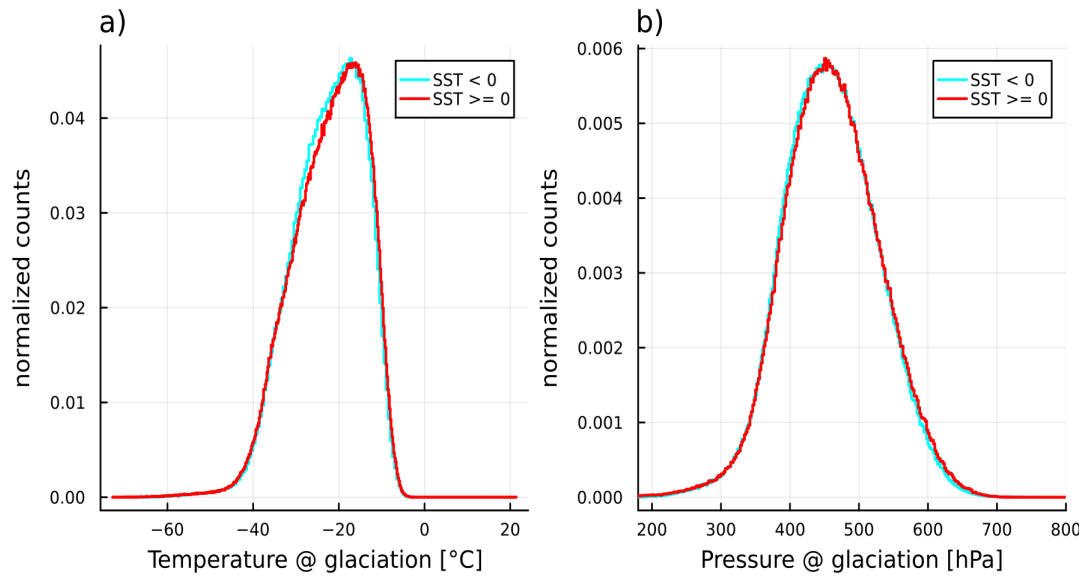


Figure SI7. Histogram of temperature (a) and pressure (b) at glaciation for all trajectories in PPE members with $SST < 0$ (cyan) and $SST \geq 0$ (red).

Chapter 8

Conclusions and Outlook

The work presented in this thesis addresses long-standing uncertainties and unknowns regarding the physical processes that govern the transport of moisture into the upper troposphere and lower stratosphere (UTLS) via warm conveyor belts (WCBs). The methods, all publicly available code, and theoretical frameworks developed in the scope of this thesis provide the scientific community with advanced tools for more thorough WCB investigations. Our findings provide new insights that will inform further investigations into the sensitivity of WCB moisture transport to simulation scale, convection intensity, and microphysical parameter choices.

Research Question I

Which microphysical processes control WCB moisture transport into the UTLS, and how do these processes differ between regions of convective and non-convective activity?

In Chapter 3, we analyzed a convection-permitting ICON simulation of a WCB case study, mainly using Lagrangian online trajectory data. The results demonstrate that convective activity affects the microphysical processes that air parcels undergo as they ascend within a WCB, as well as the ice and water vapor content in the WCB outflow. Fast ascending air experiences more hydrometeor growth by riming and produces more frozen precipitation, whereas slowly ascending air produces more liquid precipitation. Due to the production of many small cloud droplets that reach a supercooled state before freezing, fast ascending air glaciates later (at lower pressures and temperatures) than slowly ascending air. After exceeding the homogeneous freezing temperature, these cloud droplets form many small ice crystals that remain in the parcel when it reaches the WCB outflow, leading to much higher number mixing ratios of ice for fast ascending air.

Because of this large number of ice crystals, supersaturation over ice can be efficiently reduced, which leads to a relative humidity over ice distribution that peaks below 100%. On the other hand, slowly ascending air experiences a longer ascent phase, which means that there is more time for ice crystals to grow. Smaller ascent velocities also mean fewer ice crystals are retained until the end of the ascent, which leads to a larger relaxation timescale for supersaturation over ice. Therefore, the relative humidity over ice distribution in the WCB outflow for slow ascending air peaks above 100%. However, in a first-order approximation for both fast and slowly ascending air parcels, the vapor content is controlled by temperature and pressure.

Based on these results, future studies could simulate multiple WCB cases with varying levels of convective activity to determine if noticeable changes occur in the moisture conditions

of the WCB outflow. Large differences would underscore the importance of resolving convection and help us understand how WCBs might influence the Earth's climate system in the future when warmer sea surface temperatures are expected to increase convection. To validate the results from Chapter 3, it might be possible to evaluate measurements of microphysical cloud properties within WCBs that are able to distinguish regions according to the prevalence of convection, either using in-situ measurements from flights or satellite measurements.

Research Question II a

How do the modeled moisture transport, the governing microphysical processes, and the radiative effects of WCBs differ when transitioning from a convection-parameterizing simulation to a convection-permitting one?

Research Question II b

Are convection-permitting simulations better at simulating UTLS moisture in the outflow of a WCB than convection-parameterizing simulations?

In Chapters 4, 5, and 6, we conducted a thorough analysis of the role of simulation scale for WCB moisture transport to the UTLS. By comparing the high-resolution, convection-permitting ICON simulation of the WCB case study from Chapter 3 to a lower resolution, convection-parameterizing simulation of the same case using both Lagrangian and Eulerian simulation data, we were able to show that large differences exist in the modeled moisture content in the WCB outflow and the microphysical processes that occur during the ascent phase. In the higher resolution simulation, larger vertical velocities lead to higher number and mass mixing ratios of ice crystals and lower relative humidities in the WCB outflow, in accordance with the results from Chapter 3. The microphysical processes experienced by ascending air parcels also shift to more frozen-phase precipitation and enhanced riming when the resolution is increased.

We were able to identify that the convection parameterization scheme produced upper-level humidity that is too high, and ice crystal number mixing ratios that are too low, when compared to the higher resolution simulation. The amount of outgoing longwave radiation at the top of the atmosphere was found to be higher in the high resolution simulation, and was plausibly attributed to the lower vapor content in the WCB outflow. Comparisons of both simulations to satellite data using fractal analysis and object tracking solidified the assumption that the higher resolution simulation produces more realistic output (Chapter 6). However, the cloud band in the cold front progressing south of the main WCB area during the later stage of the WCBs' lifetime is poorly represented in both simulations. This emphasizes the importance of exercising caution when interpreting model output as reality in modeling studies.

These results clearly show that there are possible shortcomings in simulations that parameterize convection when modeling the transport of moisture by WCBs. Global climate models use resolutions that require convection parameterization schemes, and might therefore overestimate upper level vapor and ice crystal content in the outflow of WCBs, as well as the phase and intensity of precipitation. Similarly to the outlook presented for the study from Chapter 3, investigating multiple case studies with varying degrees of convective activity might provide insight as to how strong the biases introduced from the convection parameterization scheme are, and might offer ways to calibrate the scheme to align more closely with output from simulations that resolve convection explicitly. This may be a necessary step in improving

global climate forecasts, the accuracy of which depends heavily on correctly representing WCBs and upper level moisture. The methods introduced in Chapter 6 for comparing model output for the upper troposphere with satellite imagery could be useful for the scientific community in terms of model comparisons or validation.

Research Question III

In what ways do perturbations to microphysical parameters affect modeled moisture conditions in the outflow of WCBs, and how does this effect vary with ascent timescale?

Finally, in Chapter 7 we analyzed a perturbed parameter ensemble of 70 convection-permitting ICON simulations of a WCB case study and identified important sensitivities of WCB moisture transport to microphysical parameter choices. Specifically, we found that the ice number and mass mixing ratios, as well as the relative humidity over ice, are highly sensitive to changes in the capacitance of ice and to processes influenced by INP concentrations. CCN concentrations control the number and size of hydrometeor during the ascent but exert little influence on the conditions in the WCB outflow. Temperature, pressure and vapor content were mostly not sensitive to parameter perturbations, indicating that conditions in the outflow of a WCB are physically constrained. However, when looking only at fast ascending air parcels, changes to the saturation adjustment scheme were able to influence the pressure, temperature and vapor content. This indicates that when vertical velocities are very high, the standard saturation adjustment scheme (which does not allow for the production of supersaturated conditions over water) may introduce biases to the modeled WCB outflow.

The results from the analysis of the perturbed parameter ensemble reveal important sensitivities in the model representation of WCBs to different parameter choices. This can inform future studies that aim to quantify these sensitivities further, or help identify parameters to be perturbed in ensemble prediction systems. Additionally, our findings suggest that changes to CCN and INP concentrations, as proposed by geoengineering, could alter the composition of hydrometeors in WCB clouds and the moisture conditions in WCB outflow.

Concluding Remarks

In summary, this thesis tackled the issue of WCBs concerning the physical processes that govern their role in transporting moisture to the UTLS from multiple complementary angles. The results demonstrate that the vertical velocity structure in WCBs plays a key role in determining the upward transport of moisture and the resulting properties of water vapor and hydrometeors in the outflow. By revealing how the ascent time scale, model resolution, and microphysical parameter choices modulate these processes, this work provides the scientific community with new and useful knowledge and opens the door to many possible future studies.

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