



Evaluation of the vertical microphysical properties of fog as simulated by Meso-NH during the SOFOG3D experiment

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Abstract. This study evaluates the representation of fog microphysics in high-resolution simulations from the Meso-NH model using the two-moment LIMA microphysical scheme, based on data from the SOFOG3D field campaign. This campaign combines remote sensing and vertical microphysical observations from a tethered balloon. Two fog events were simulated in order to assess the model's ability to reproduce their life cycle and identify any missing physical processes. The analysis focuses on the vertical microphysical structure, observed consistently by the various instruments, and on the simulated processes during the different phases of fog development (i.e. before, during, and after the transition from thin to thick fog). The model realistically reproduces the thermodynamic and dynamic evolution of fog, resulting in a satisfactory simulation of its development stages. A comprehensive analysis of microphysical processes is conducted throughout the entire height of the fog, based on a comparison with observations and a budget of modelled processes. While microphysics is generally well represented, certain systematic errors emerge: excessive liquid water content values during the thin-to-thick transition and adiabatic phases, due to excessive condensation; an inaccurate representation of the droplet size distribution, with the absence of the largest droplets; and an inability to capture the droplet concentration vertical gradient, with values that are too high near the ground. While some of these shortcomings can be explained by dynamic biases, and more cases are needed to confirm our results, various recommendations are proposed. These include assessing the impact of drizzle and representations that could benefit all warm clouds.

1 Introduction

Modeling fog is a significant challenge to forecasters. A key step towards achieving this is to understand the complex, interrelated physical processes that influence the fog life cycle. Fog results from the intricate interactions of local airflow, surface processes, radiation, turbulence, aerosol loading, and microphysics.

Numerous numerical studies have investigated the relationships between fog microphysics and other physical processes (see, for example, Zhang et al., 2014; Mazoyer et al., 2017; Boutle et al., 2021; Contreras Osorio et al., 2022). These studies emphasize the critical role of accurately representing the Droplet Size Distribution (DSD) for sedimen-

tation, radiative, and dynamic effects throughout the fog life cycle. However, even bin models, which are among the most accurate microphysical models, have struggled to capture this complexity (Boutle et al., 2021). In particular, the transition from thin fog to well-mixed thick fog remains poorly understood (Price, 2011; Bergot and Lestringant, 2019) and may be linked to microphysical processes. This transition occurs as the surface becomes radiatively shielded by the forming fog, causing the radiation inversion to shift upwards and thereby initiating a convective regime (Roach et al., 1976). In modeling scenarios, this shift tends to be too abrupt, often due to excessive fog droplet number concentration and liquid water content (Poku et al., 2021; Boutle et al., 2021). A nu-

merical study by Boutle et al. (2018) suggested that aerosol growth and activation are crucial in triggering adiabatic fog.

Modeling these microphysical processes remains challenging (Boutle et al., 2018; Schwenkel and Maronga, 2019; Poku et al., 2021; Contreras Osorio et al., 2022; Vié et al., 2024), which has motivated the study of fog microphysics in field campaigns for many years (Gerber, 1991; Wendisch et al., 1998; García-García et al., 2002; Zhao et al., 2013; Price et al., 2018; Mazoyer et al., 2022; Ghude et al., 2023; Nurowska et al., 2025). Experimental field studies are essential to improve our understanding of fog microphysical processes and their modeling. Historically, measurements were primarily made near the ground for practical reasons. Data from numerous field campaigns indicate that, for diameters of 2–50 μm , surface droplet concentrations (N_c) near the ground range from 10 to 300 cm^{-3} , with Liquid Water Contents (LWC) varying between 0.001 and 0.4 g m^{-3} . The DSD has been documented as having undergone four evolutionary phases (formation, development from thin fog to well-mixed thick, mature and dissipation), driven by a combination of the following processes: activation, condensation, evaporation and sedimentation, while collision-coalescence or Ostwald ripening (when larger droplets grow at the expense of smaller ones) were only suspected. It has been shown that transitions between phases are linked to thermodynamics and radiation evolution (Mazoyer et al., 2022). However, little is known about microphysical properties in the fog layer as vertical measurements of fog microphysics using tethered balloons have a short history.

One of the aims of the Local and Nonlocal Fog Experiment (LANFEX) campaign (Price et al., 2018) was to investigate the vertical microphysical structure of fog using tethered balloon observations between September 2014 and March 2016. However, the scarcity of the profiles collected during LANFEX made documentation difficult. A few years later, the SOuth West FOGs 3Dimensions (SOFOG3D) campaign (Burnet et al., 2020) investigated the vertical microphysical structure of fog during the winter of 2019–2020 in South-West France using tethered balloon observations. The campaign employed a combination of in situ instrumentation and remote sensing to offer a thorough evaluation of fog events. SOFOG3D provided a valuable opportunity to analyze the vertical and temporal evolution of various fog layers.

So far, initial vertical measurements could not investigate the fog life cycle as they corresponded to isolated profiles within the fog layer, and they produced apparently contradictory results. While Goodman (1977) and Pinnick et al. (1978) showed DSD broadening with height, other studies from Okita (1962), Pilié et al. (1975), and Egli et al. (2015) reported DSD narrowing with height. These results may be explained by the timing of the measurements (before, during, or after the transition from thin to well-mixed thick fog), as demonstrated by Dione et al. (2023) and Costabloy et al. (2025) for LWC using SOFOG3D measurements. By analyzing 140 vertical profiles of LWC measured during this cam-

paign, Costabloy et al. (2025) showed the following trend; before the transition, high LWC values are concentrated near the bottom; during the transition, higher amounts are found near the top as the temperature profile becomes unstable; and in the mature adiabatic phase after the transition, high LWC values are typically found near the fog top. However, these strong values at the fog top can be disrupted by entrainment-mixing processes, which reduce fog adiabaticity. The authors demonstrated that the transition is strongly linked to an increase in fog optical thickness, which is related to DSD. According to these authors, collision-coalescence and sedimentation may redistribute the LWC through the fog layer from the top to the ground, thereby balancing the described trend.

Modelling can provide additional insight into the microphysical processes that occur during the fog life cycle. Accurate prediction of the fog life cycle is also challenging. This requires the use of a two-moment microphysical scheme that considers prognostic droplet concentration in addition to mixing ratio, as demonstrated by Ducongé et al. (2020), Fathalli et al. (2022), Antoine et al. (2023), and Vié et al. (2024) among others. However, these schemes still exhibit inaccuracies and therefore require specific validation for fog. To improve fog representation, recent modeling studies have consequently focused on aerosol activation schemes, which parameterize the droplet number concentration based on a change in supersaturation at a given time. However, initial supersaturation calculations, inherited from convection studies, only consider adiabatic lifting and many impose a minimum vertical velocity (e.g. 0.1 m s^{-1}) to account for the unresolved subgrid ascent. According to Poku et al. (2021), fog's optical thickness is overestimated due to excessive aerosol activation during formation whereas at this time, supersaturation should be driven by non-adiabatic processes such as radiative cooling. This leads to overly intense convective regimes and a feedback loop of high activation and LWC production. They show that reducing the minimum vertical velocity and including radiative cooling in the supersaturation calculation slows the transition from thin to well-mixed thick fog. However, they also state that physical processes should be missing. In addition to Poku et al. (2021), Vié et al. (2024) also accounted for condensation on pre-existing cloud droplets as a sink term in the supersaturation equation. They demonstrated that fog development was slower, and that the supersaturation values were very close to those obtained using the reference but costly semi-prognostic scheme of Thouron et al. (2012). However, these two studies were based on the LANFEX campaign, which observed a limited number of microphysical vertical profiles.

The analysis of DSD evolution can also provide valuable insights into the microphysical processes at work. In a bulk scheme, the DSD is given by a distribution law, whose parameters depend on LWC and N_c for a two-moment microphysical scheme. Additionally, LWC and N_c represent the first- and third-order moments of the DSD, respectively. But, the evolution of the observed DSD is far more com-

plex. The evolution of each DSD bin in the commonly observed diameters of 2–50 μm may result from different processes. Following Mazoyer et al. (2019, 2022), the growth of the 10 μm small-diameter mode results from the activation of aerosol particles into droplets. Its condensation growth would feed small- to medium-sized droplets (up to 30 μm in diameter). For larger diameters, as stated by Yau and Rogers (1996), their growth must result from other processes, as the timescale for condensation growth is longer than their appearance in fog (see also Mazoyer et al., 2022). Several hypotheses can explain the presence of droplets larger than 30 μm in diameter, including collision-coalescence (Beard and Ochs, 1993; Pinsky et al., 2001), which can be gravitational or due to turbulence (Grabowski and Wang, 2013; Vaillancourt and Yau, 2000) especially at the fog top. Further investigations based on both observation and modelling are still needed to better understand the microphysical processes responsible for the evolution of the DSD.

This study aims to evaluate the two-moment microphysical scheme Liquid Ice Multiple Aerosols (LIMA) (Vié et al., 2016) with the supersaturation equation developed in Vié et al. (2024) within the Meso-NH model (Lac et al., 2018), using high-resolution mesoscale simulations and the SOFOG3D observations to evaluate fog microphysics. Notably, Meso-NH is the research laboratory for the French operational model AROME (Seity et al., 2011; Brousseau et al., 2011, 2016), which may integrate the LIMA microphysical scheme in the near future. This study will focus on radiation/advection fog during Intense Observation Periods (IOP) 6 and 11 of the SOFOG3D campaign, which have been well documented in Dione et al. (2023) and Costabloy et al. (2025). Microphysical assessments will be conducted according to the four evolutionary phases (formation, transition from thin to thick, mature adiabatic phase and dissipation) outlined in Mazoyer et al. (2022) and Costabloy et al. (2025). During IOP 6, 42 tethered-balloon profiles will be analyzed and 16 during IOP 11. A budget analysis of simulated LWC and N_c , combined with an analysis of DSD evolution, will be performed to identify the primary microphysical processes involved in these phases, and to evaluate their representation in LIMA.

To the authors' knowledge, this is the first study to evaluate simulated vertical microphysical profiles alongside observations throughout the fog life cycle. The study addresses the following questions:

- To what extent is the fog life cycle, especially the vertical microphysical profiles (LWC, droplet concentration, DSD, and reflectivity), accurately represented using the two-moment microphysical scheme LIMA, particularly in light of the recent improvements to the treatment of supersaturation?
- Which microphysical processes, if any, are still missing from the representation, and how could the model be improved?

The paper is structured as follows. Section 2 describes the instrumental setup, observed cases, and model characteristics. Section 3 evaluates the simulations by comparing them with thermodynamic and dynamic measurements. Section 4 outlines the methodology for determining the transition from thin to well-mixed thick fog and different fog phases. Section 5 compares model simulations with tethered balloon data regarding microphysics and analyzes the main simulated microphysical processes. Sections 6 and 7 provide discussion and conclusions, respectively.

2 Case description and methodology

2.1 SOFOG3D Measurements set-up

The SOFOG3D field experiment (Burnet et al., 2020) took place in southwestern France, as illustrated in Fig. 1, which shows the geographical location of the campaign at the edge of the Landes forest, an area characterized by a heterogeneous orography between two river valleys. Conducted during the autumn and winter of 2019–2020, the campaign included 15 IOPs, and its dataset has been extensively analyzed in previous studies (Martinet et al., 2022; Antoine et al., 2023; Dione et al., 2023; Thornton et al., 2023; Costabloy et al., 2025; Price, 2025).

Measurements were collected at 17 sites over an area of $30 \times 50 \text{ km}^2$ (larger rectangle of Fig. 1a), but this paper focuses on data from a denser $10 \times 10 \text{ km}^2$ region surrounding the supersite, which includes Jachère and Charbonnière – the most instrumented sites (Fig. 1b). Both sites were equipped with ground-based meteorological stations that provided standard measurements, including temperature, wind speed and direction, and longwave and shortwave radiation. The Jachère site also provided flux measurements, particularly turbulent kinetic energy and vertical wind variance.

A distinctive feature of the SOFOG3D campaign was its profiling capabilities, which combined remote sensing and in situ measurements. At the Charbonnière site, the 95 GHz Bistatic rAdar SysTem for Atmospheric studies (BASTA) vertical cloud radar (Delanoë et al., 2016) measured droplet reflectivity and Doppler velocity up to 18 km, with a vertical resolution of 12.5 m. The Cloud Top Height (CTH) was derived from the reflectivity data. In addition, a Humidity And Temperature PROfiler (HATPRO) 1/10 Hz radiometer was deployed at Charbonnière, using neural network inversion (Martinet et al., 2022) to measure vertical profiles of temperature and humidity up to 2.5 km, with a vertical resolution of 25 m at lower altitudes and 30 m at higher altitudes. Note that Martinet et al. (2022) estimated a 0.5 K cold bias below 2 km and a 1.4 g m^{-3} dry bias below 1.7 km when comparing HATPRO measurements with radiosounding (RS) data. The radiometer also retrieved the Liquid Water Path (LWP) for the entire atmospheric layer with errors between 1 and 14 g m^{-2} according to Martinet et al. (2022). A WindCube V2 lidar manufactured by Leosphere provided

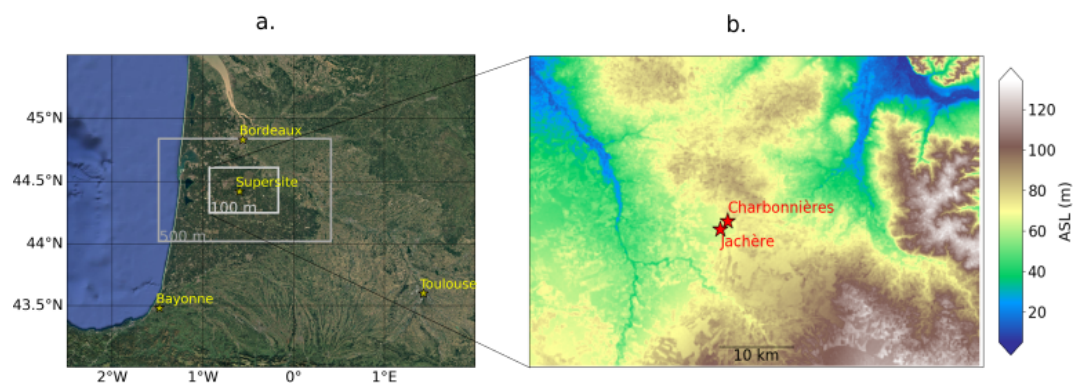


Figure 1. (a) Geographical location of the SOFOG3D supersite (Imagery © 2025 NASA, Map data © 2025 Google) and domains used for the 500 m simulation (dark grey rectangle) and the 100 m simulation (light grey rectangle). (b) Orography within the nested 100 m domain centred on the supersite station. The red stars correspond to the two measurement sites that make up the supersite.

the wind measurements at 10 levels from 40 m to 220 m a.g.l. at the Charbonnière site. The measurements, made at a frequency of 1 Hz and a vertical resolution of 20 m, allow for the estimation of turbulence parameters such as Turbulent Kinetic Energy (TKE). In the presence of fog or low stratus, the vertical range of the lidar decreases. WindCube lidar data are presented in more details in Dione et al. (2023). Furthermore, a tethered balloon equipped for microphysical measurements was deployed at the Charbonnière site, using a modified Cloud Droplet Probe (CDP) to collect data. The CDP measures DSDs ranging from 2 to 50 μm in diameter. To operate under a tethered balloon, a wind vane was used to align the sampling section perpendicular to the wind and a small fan fixed just to the rear of the laser beam sucked in the air flow. The air speed in the sampling section was therefore equal to the wind speed plus 5 m s^{-1} . More details on the CDP under the tethered balloon can be found in Fathalli et al. (2022), Dominutti et al. (2022), and Costabloy et al. (2025).

Instrumentation at the Jachère site was set up to assess the aerosol load and its activity as Cloud Condensation Nuclei (CCN). This included a continuous flow streamwise thermal gradient CCN chamber (CCNC) (Roberts and Nenes, 2005), which measured CCN number concentration at five different supersaturations ranging from 0.06 % to 0.28 %. A Scanning Mobility Particle Sizer (SMPS) provided the number size spectrum of dry aerosol particles. The SMPS consisted of a differential mobility analyzer (DMA; TSI 3071), which selected particles from 10.6 to 496 nm, paired with a Condensation Particle Counter (TSI CPC 3022).

A summary of the instruments used in this study is presented in Table 1.

2.2 SOFOG3D case studies

This study examines two fog events observed during IOPs at the SOFOG3D supersite.

During the IOP 6, the fog formed on 5 January 2020 at 20:37 UTC and dissipated on 6 January 2020 at 09:28 UTC under high-pressure conditions associated with a strong anticyclone over central Europe, and lasted 12 h and 51 min. Satellite images (not shown) indicated low-level cloud advection at 20:00 UTC: the fog was classified as radiative by the Tardif and Rasmussen (2007) algorithm but as a radiative-advective fog according to Costabloy et al. (2025).

In contrast, IOP 11 was a radiative fog (Costabloy et al., 2025) that formed at 20:38 UTC on 8 February 2020 and dissipated at 03:49 UTC on 9 February 2020, lasting 7 h and 11 min. This fog developed after the passage of a cold front earlier in the day, but its persistence was disrupted by the arrival of high clouds at 04:00 UTC on 9 February, leading to its dissipation before sunrise (satellite images not shown).

The Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT, Draxler and Rolph, 2010) model suggests that the air masses during IOP 6 were of continental origin, whereas those during IOP 11 were of maritime origin (data not shown). Due to technical issues, aerosol measurements were not available for these dates but for other cases of the campaign which had allowed to retrieve an average activation spectra. A sensitivity study to aerosol loading is consequently described in the discussion.

2.3 Model set-up

The non-hydrostatic research model Meso-NH (Lac et al., 2018) is used in a 3D configuration with hectometre-scale resolution, which Cuxart (2015) describes as high-resolution, and has proven effective for fog simulations (Ducongé et al., 2020; Fathalli et al., 2022). Simulations were performed over a 48 h period, starting on 5 January at 00:00 UTC for IOP 6 and on 8 February at 00:00 UTC for IOP 11. Fields from the AROME model (Seity et al., 2011; Brousseau et al., 2011), with a 1.3 km horizontal grid, were used as initial and boundary conditions with the configuration identified as the most satisfactory: AROME analyses for IOP 6, and AROME-IFS

Table 1. List of instruments used in this study.

Instruments	Vertical position	Measured Variable	Unit	Uncertainty	Temporal Resolution
Kipp and Sonnen Spectrometer	1 m	LWD/LWU	W m^{-2}	10 %	60 s
METEK Sonic Anemometer	3 m	TKE	$\text{m}^2 \text{s}^{-2}$	20 %	30 min
Vaisala PWD22	3 m	Horizontal Visibility	m	15 %	15 s
BASTA Cloud radar	profile	Reflectivity (Z) CBH/CTH	dBZ m	0.5–2 dB	3 s
HATPRO Radiometer	profile	T LWP	K g m^{-2}	1.5 K RMSE 20 g m^{-2}	10 min 1 s
XSENS Sensor	profile	Altitude	m	5 m	10 s
Cloud Droplet Probe	profile	LWC, N_c $D = [2\text{--}50] \mu\text{m}$	g m^{-3} , cm^{-3}	30 %, 20 %	1 s
Vaisala RS41 Radiosounding	profile	P , T HU, V	hPa, $^{\circ}\text{C}$ $\%$, m s^{-1} °	1 hPa, 0.4°C 4 %, 0.15 m s^{-1} 2°	1 s
TSI SMPS	3 m	Dry aerosols particles distribution $D = [10.6\text{--}493] \text{ nm}$	cm^{-3}	20 %	5 min
CCNC DMT	3 m	CCN concentration $\text{SS} = [0.06\text{--}0.26] \%$	cm^{-3}	10 %	20 min

forecasts for IOP 11. AROME-IFS refers to an AROME configuration in dynamical adaptation from Integrated Forecasting System (IFS), i.e., initialized with IFS analyses and coupled with IFS forecasts.

The parent domain has a 500 m horizontal resolution, covering a $150 \text{ km} \times 90 \text{ km}$ area (Fig. 1a). Within this domain, a nested 100 m horizontal grid model spans $60 \text{ km} \times 40 \text{ km}$ and is centered on the main observation site. Both nested domains use 138 vertical levels, extending from the ground to the model top at 10 km, with vertical grid spacing ranging from 2 m near the ground to 500 m at the model ceiling. A nudging technique is applied to large-scale dynamical and thermodynamic fields above 9000 m a.s.l., using an absorbing layer to relax forecast variables toward large-scale conditions.

For momentum variables, a fourth-order centered advection scheme is employed, along with a fourth-order Runge–Kutta time integration. Scalar variables are transported using the piecewise parabolic scheme (Colella and Woodward, 1984).

The atmospheric model is coupled to the soil-biosphere-atmosphere interaction surface scheme (Noilhan and Planton, 1989) through the externalized surface model SURFEX (Masson et al., 2013). Orography data come from the Shuttle Radar Topography Mission, providing a resolution of 90 m for the 500 m model and 30 m for the 100 m model. Land cover and surface parameters are derived from the ECOCLIMAP database at 1 km resolution. The turbu-

lence scheme is based on a turbulent kinetic energy equation (Cuxart et al., 2000), implemented in one-dimensional mode with the Bougeault and Lacarrere (1989) mixing length for the 500 m grid, and in three-dimensional mode with the Deardorff (1980) mixing length for the 100 m grid, both of which are well-suited to these resolutions. At 500 m grid length, subgrid condensation is activated. It depends on the parameters computed by the turbulence scheme and a climatological variance (Rooy et al., 2010; de Rooy et al., 2022) associated with a critical relative humidity of 96 %. At the 100 m grid length, no subgrid condensation scheme has been used; sensitivity tests could be performed in future studies, as in Smith et al. (2021). In the following parts, we will only consider the 100 m simulated fields.

The microphysical scheme used is the two-moment LIMA scheme (Vié et al., 2016), which prognoses the mixing ratios of five hydrometeor species (cloud droplets, cloud ice, snow, rain, and graupel) and the concentrations of droplets, raindrops, and ice crystals. LIMA also accounts for the evolution of the aerosol population, providing a detailed representation of aerosol–cloud interactions to address competition between multiple CCN modes. Each CCN mode is characterized by its chemical composition, size distribution, and nucleation properties. The parameterization of CCN activation, based on the maximum supersaturation diagnosis (Cohard and Pinty, 2000), considers vertical velocity (including a contribution from TKE), cloud droplet growth by condensation, and radiative cooling. Recent improvements in the CCN activation

parameterization in LIMA, as developed in Vié et al. (2024), exclude the advection contribution to the temperature cooling term and take into account the growth of existing cloud droplets by condensation, thereby mitigating the overestimation of cloud droplet activation. The rain formation rate in LIMA is determined by the cloud droplet size distribution following Berry and Reinhardt (1974) and Cohard and Pinty (2000). Droplet sedimentation is calculated using the Stokes law for cloud droplets and is active till the ground, while fog deposition is activated with a constant velocity of 2 cm s^{-1} according to Mazoyer et al. (2017) to represent the aerodynamic capture of drops. LIMA assumes that the cloud droplet size distribution $nc(D)$ follows a generalized gamma law:

$$nc(D) = N_c \frac{\alpha}{\Gamma(\nu)} \lambda^{\alpha\nu} D^{\alpha\nu-1} \exp(-(\lambda D)^\alpha), \quad (1)$$

where λ is the slope parameter, depending on the prognostic variables r_c (cloud mixing ratio, kg / kg) and N_c (cloud droplet concentration, cm^{-3}):

$$\lambda = \left(\frac{\pi}{6} \rho_w \frac{\Gamma(\nu + 3/\alpha)}{\Gamma(\nu)} \frac{N_c}{\rho_a r_c} \right)^{1/3} \quad (2)$$

ρ_a and ρ_w are the density of dry air and water, and α and ν are the parameters of the gamma law. These parameters used for both microphysics and radiation are set to $\alpha = 3$ and $\nu = 1$ within LIMA.

To specify aerosol loading, the characteristics of the activation spectra Cohard and Pinty (2000) were calculated according to the κ -Köhler theory (Petters and Kreidenweis, 2007) implemented in LIMA, using all data collected during the campaign with the SMPS (aerosol size distribution from 10.6 to 496 nm) and the CCNC (activation spectra from 0.06 % to 0.28 %), following the methodology in Mazoyer et al. (2019). The mean activation spectra obtained during SOFOG3D were found to be consistent with the Paris-Fog data (Mazoyer et al., 2019), with values of 80 cm^{-3} at 0.05 %, 250 cm^{-3} at 0.1 % and 1000 cm^{-3} at 0.3 % (not shown).

To compute model reflectivity offline, this study employs the radar forward operator developed by Augros et al. (2016). This forward operator accounts for hydrometeor and water vapour attenuation and is based on the T-matrix scattering theory (Mackowski and Mishchenko, 1996). Its implementation within Meso-NH has been previously described in Mazoyer et al. (2023).

The radiation code used in this study is ecRad (Hogan and Bozzo, 2018), with the Rapid Radiation Transfer Model (RRTM, Mlawer et al. (1997)) as the longwave (LW) radiation scheme, and 14 bands for the SW radiation. Cloud optical properties are computed according to Savijärvi et al. (1997) with the LIMA scheme by taking into account the prognostic cloud droplet concentration. Note that the radiative effect of aerosols is only considered through a monthly climatology in these simulations. Although this is beyond the scope of the present study, future research should assess

the direct and semi-direct effects of aerosols on fog lifetime, as has been done in Ding et al. (2019) and Al Asmar et al. (2022).

2.4 Observational limits and impact on comparisons

This section analyzes the effect of observational constraints regarding microphysics and their influence on comparisons. Table 1 presents the instrument uncertainties and measurement ranges; refer to the “Measured Variable” column. It is important to note that the CDP measures only within the diameter range $[2\text{--}50] \mu\text{m}$, a range widely used in fog modeling comparison studies (Price et al., 2018; Gultepe et al., 2021; Wagh et al., 2023). In the model, the LIMA microphysical scheme distinguishes between droplet and rain categories based on Cohard and Pinty (2000) and assumes a separation threshold at $82 \mu\text{m}$ in diameter. However, since LIMA is a bulk scheme, only the first moment (number concentration) and the third moment (mass) of the distribution are predicted. Therefore, a thorough evaluation of the model’s cloud droplet spectra compared to observational spectra is essential before any comparison. This ensures that equivalent diameter ranges are compared, which will be addressed in the remainder of this study (see Sect. 5.3). Further insights into cloud characteristics are offered by BASTA, capable of identifying particles larger than $50 \mu\text{m}$. Comparisons between the reflectivity measured by BASTA and the reflectivity derived from the 6th-moment of the distribution measured by the CDP may consequently reveal the existence of particles exceeding $50 \mu\text{m}$, that will be explored in a specific subsection of this paper (see Sect. 5.1). Nevertheless, the model calculates reflectivity by accounting for all particles, enabling a precise comparison with the observed reflectivity.

3 Thermodynamic and dynamic evaluation by comparison to measurements

This section evaluates the simulated fog event by comparing it with thermodynamic and dynamic measurements. The objective is to assess the consistency of the subsequent detailed microphysical comparisons.

In order to evaluate the fog life cycle and its physical characteristics, we first compare observed and simulated vertical reflectivities. Next, we analyze the vertical profiles of the thermodynamic fields using radiosoundings (RS), alongside data from the HATPRO radiometer and wind lidar for the dynamic fields.

3.1 RADAR evaluation

Figure 2 compares the simulated radar reflectivity and LWP with the CDP observations under the tethered balloon and the BASTA radar observations and LWP from HATPRO at Charbonnières for IOPs 6 and 11, using a vertical logarithmic scale. Please note that comparisons of reflectivity de-

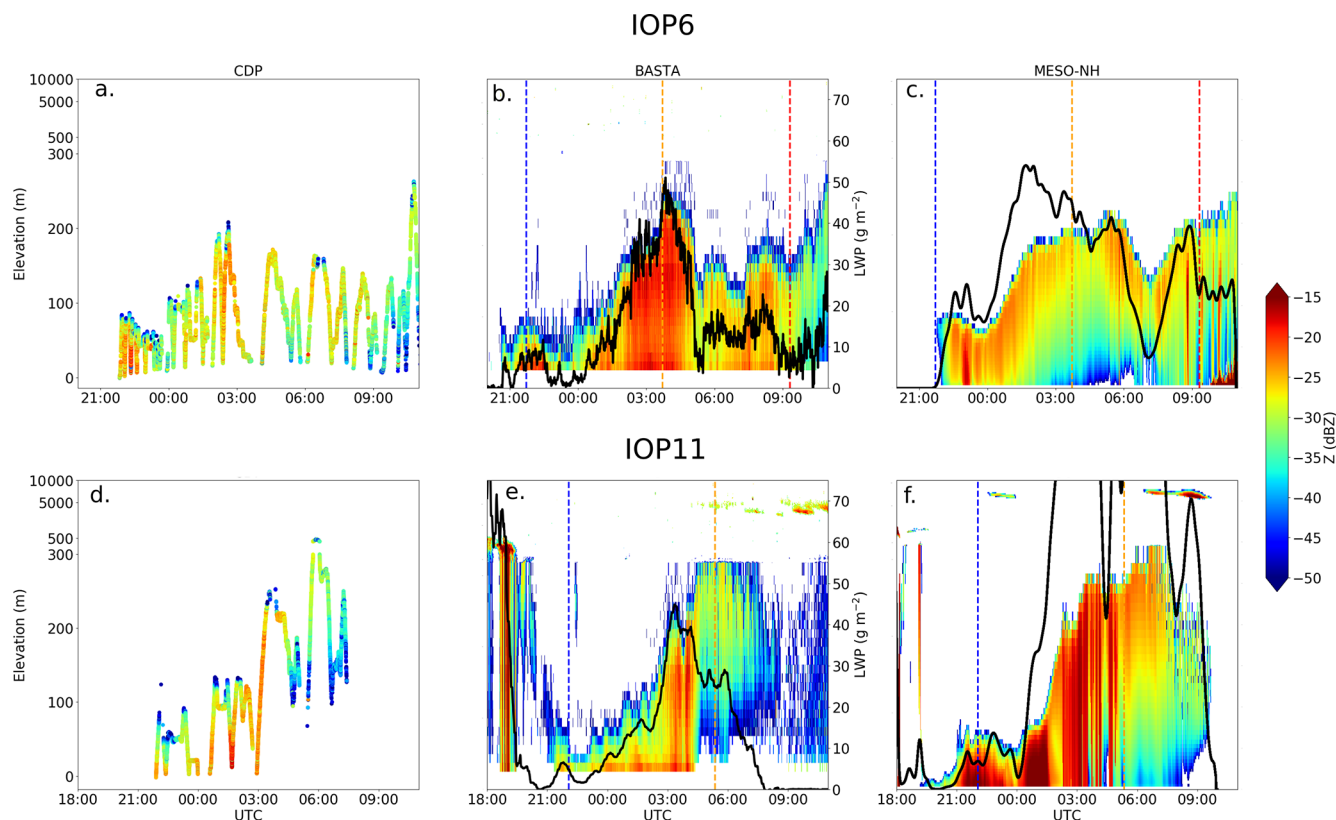


Figure 2. Temporal evolution of vertical profiles of radar reflectivity for IOP 6 (a, c) and IOP 11 (d, f), derived from CDP observations under the tethered balloon (a, d), from BASTA radar (b, e) and the 100 m simulation (c, f). Black lines represent the observed LWP from HATPRO (b, e) and the 100 m simulation (c, f). Vertical coloured dashed lines refer to the RS start time presented in Fig. 3. The vertical axis follows a logarithmic scale above 500 m.

rived from CDP and measured by the BASTA radar will be treated in Sect. 5.

For IOP 6, the observed fog from radar formed at 20:38 UTC and rapidly developed vertically, reaching 100 m by 21:30 UTC. The fog top subsequently lowered until 00:00 UTC due to the wind shifting from southwest to southeast, and then continued to develop progressively until 04:00 UTC, finally dissipating as stratus by 09:00 UTC. High reflectivity values (around -20 dBZ) were recorded between 02:00 and 05:00 UTC with the BASTA. While the observed BASTA reflectivity values varied over time, they remained fairly homogeneous vertically. Costablos et al. (2025) classified this episode as radiative-advective fog, because advection from the west maritime area favored the formation of fog in addition to radiative cooling near the ground. The simulated fog for IOP 6 exhibited a similar fog development, although discrepancies are noted. Notably, the simulation did not capture a thin fog layer before 22:00 UTC. The advective contribution to fog formation in the simulation appeared to be dominant over local formation, as fog first appeared at a height of 100 m at 22:00 UTC, before lowering abruptly. After formation, the simulated fog developed vertically to 250 m before lowering again and evolving into a stratus layer

by 09:00 UTC, which is in close agreement with the observation. However, lower reflectivity values were simulated when the fog vertically extended, and the simulated vertical reflectivity gradient did not match the BASTA observations well. It is important to note that the strong radar reflectivities in the surface boundary layer after 09:00 UTC corresponds to minimal rain formation below the stratus. The simulated LWP values were generally higher, however both observed and simulated values reached up to 50 g m^{-2} , but at 04:00 UTC in observations and 02:00 UTC in the simulation. The lower reflectivity despite higher simulated LWP could be attributed to potential inaccuracies in the simulated DSD with an underestimation of large droplets, as the reflectivity is proportional to the sixth order moment of the DSD. This point will be explored in more detail later on.

For IOP 11, the observed fog from radar formed at 20:38 UTC and was classified as a radiative fog by Costablos et al. (2025), considering only its formation, which was primarily driven by significant radiative cooling near the ground. The fog then developed progressively until 04:00 UTC, when the first high clouds, indicative of an approaching warm front, appeared at 10 000 m. As the advection became a major process during its life cycle, Dione

et al. (2023) classified IOP 11 as an advection-radiation fog. A reverse vertical reflectivity gradient, with the highest values near the ground, was observed with the BASTA until 03:00 UTC, suggesting the presence of large droplets near the bottom and a coalescence signature (Kumjian and Prat, 2014). From 03:00 UTC, the highest reflectivity values were recorded near the fog top, with the observed LWP reaching up to 45 g m^{-2} . The simulation for IOP 11 was in good agreement with the BASTA observations in terms of fog development until 04:00 UTC. However, when examining the vertical reflectivity profile, simulated reflectivity values were significantly higher than observed, correlating with higher simulated LWP. After 04:00 UTC, high clouds were also simulated, but they were likely too thin, and the fog too thick, leading to a delayed fog dissipation in the simulation – occurring only at 09:00 instead of 04:10 UTC as observed.

Overall, despite discrepancies in reflectivity intensity and vertical profile, the fog evolution as inferred from reflectivity is generally consistent with observations for both IOP 6 and IOP 11 in the period after formation for IOP 6 and up to 04:00 UTC for IOP 11. We will focus on these periods in the following sections.

3.2 Thermodynamics and dynamic evaluation

3.2.1 Radiosoundings

Before making any meaningful comparison of the micro-physical properties, the consistency between the model and observations in terms of thermodynamics must first be verified. Figure 3 compares the RS profiles of potential temperature and vapour mixing ratio (dotted lines), launched at Charbonnières, with the vertical profile simulated by Meso-NH (solid lines) at the same time. In addition, we include the vertical profiles from the coupling model (dashed lines), using temporally interpolated analyses from AROME for IOP 6 and forecasts from AROME-IFS for IOP 11, to assess the influence of the Meso-NH simulations. The vertical dashed lines in Fig. 2 indicate the launch times of the RS and the simulated profiles, allowing us to examine the coincident vertical evolution of the fog.

For IOP 6, at 18:00 UTC, the stable boundary layer and surface cooling show that Meso-NH and AROME were not cold and dry enough in the surface layer compared to the RS data. By 21:43 UTC (blue lines in Figs. 2 and 3), when both the observed and modeled fog layers became adiabatic, the coupling model closely aligned with the RS, although Meso-NH remained slightly warmer and wetter, retaining a higher water vapour content. This is consistent with the delayed fog formation in Meso-NH, which occurred at 22:00 UTC, later than in the observations (Fig. 2a, b). At 03:43 UTC, when the fog reached its maximum vertical development, the adiabatic potential temperature profiles were quite similar, although always slightly colder in the observations. In the same way, the observed fog layer was higher, consistent with Fig. 2a, b. By

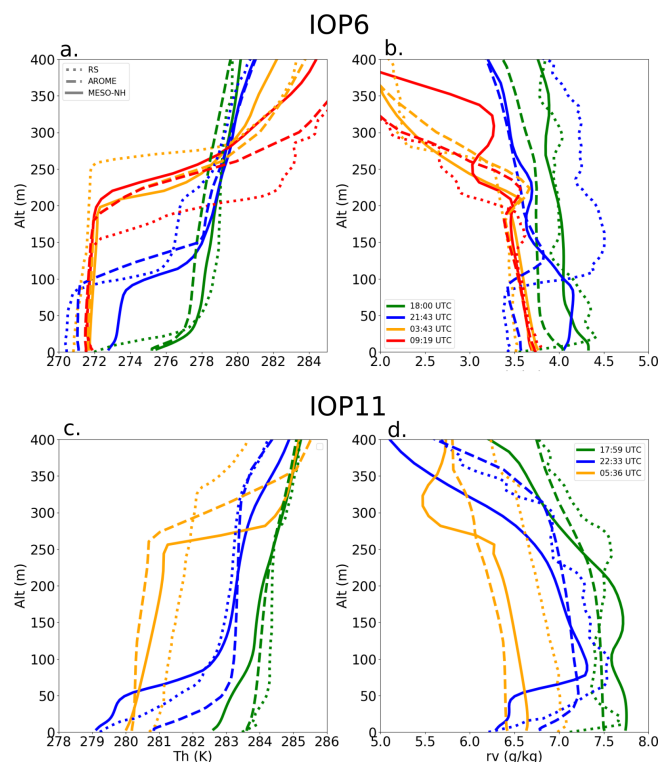


Figure 3. Vertical profiles of potential temperature (a, c) and vapour mixing ratio (b, d) at Charbonnières from radiosounding (dotted lines) and corresponding profile from AROME (dashed lines) and Meso-NH (plain lines) for different launch times 18:00, 21:43, 03:43, and 09:19 UTC for IOP 6 (a, b) and 17:59, 22:33, and 05:36 UTC for IOP 11 (c, d). AROME fields correspond to temporally interpolated AROME analyses for IOP 6 and AROME-IFS forecasts for IOP 11. The launch times are reported in Fig. 2.

09:19 UTC, the model simulated a more developed fog than observed, but the profiles remained very similar beneath the inversion.

For IOP 11, small differences between the observations, Meso-NH, and AROME-IFS are apparent from 18:00 UTC but they are quite small, and explain why fog formation was almost correctly reproduced, unlike in IOP 6. These differences persisted throughout the simulation. Meso-NH appeared cooler and drier, while AROME-IFS was warmer and wetter than the observations. The discrepancies observed at 22:33 UTC between the RS and Meso-NH may be due to the more developed simulated fog. By 05:36 UTC, after the first disturbances from the approaching warm front arrived, the profiles show similar trends. Notably, for this IOP, stable profiles were observed and modeled at 17:59 and 22:33 UTC, whereas a quasi-adiabatic profile was observed and modeled only at 05:36 UTC, indicating a more progressive thin-to-thick transition than during IOP 6.

Overall, Meso-NH exhibits some discrepancies relative to its coupling model AROME, highlighting the specific influences of its configuration (such as the vertical grid) and phys-

ical parametrizations, in addition to the different dynamics. A precise analysis of the causes of the differences is beyond the scope of this study. Nevertheless, the simulated vertical profiles of potential temperature and vapour mixing ratio appear to be consistent with the observed RS data, in terms of absolute values and inversion level throughout the fog life cycle, for both IOP 6 and IOP 11, lending confidence to the thermodynamic representation of these fog events. It should be noted, however, that the simulation biases differ between the two IOPs: the model is generally too warm and too moist for IOP 6, while it is too cold and too dry for IOP 11.

3.2.2 Radiometer and lidar

Figures 4 (IOP 6) and 5 (IOP 11) show the temporal and vertical temperature profiles measured by the HATPRO radiometer, alongside wind profiles and turbulent kinetic energy (TKE) recorded by lidar. Radiosounding measurements are also included for temperature, wind and vapour mixing ratio. The simulated fields from Meso-NH are compared, with the simulated TKE field including resolved and sub-grid energy. It is important to note that lidar wind and TKE measurements are subject to strong attenuation due to fog droplets, from bottom to top, which limits the vertical measurement.

For IOP 6, HATPRO observations suggest temperatures that are slightly warmer than those indicated by RS data within the fog layer. The simulated temperatures are slightly cooler than those from HATPRO and fall between the RS and HATPRO temperature measurements. The fog top is also more pronounced in the model than in HATPRO observations, primarily due to the lower vertical resolution of the measurements. Simulated winds below 150 m show fairly good agreement with lidar wind measurements, supporting their use for analyzing the dynamic characteristics of the fog layer. The simulated profiles of temperature, water vapour mixing ratio, and wind highlight a warm and dry low-level southerly jet, starting after 00:00 and intensifying above 200 m from 06:00 UTC onward. This jet likely contributes to the slight lowering of the fog top at 06:00 UTC. Observed TKE values are significant within the fog layer and are also captured by the model. However, Meso-NH overestimates TKE near the ground after 01:30, as confirmed by additional measurements from a 3-m anemometer (shown later in Fig. 6e). The simulation favours a bottom-driven boundary layer, with higher TKE values near the surface, whereas the observed vertical TKE profiles, although limited by droplet attenuation, are more homogeneous, with some profiles showing slightly higher TKE values near the fog top and therefore being more consistent with a top-driven boundary layer. These discrepancies are likely related to the overestimated fog-top inversion and excessively strong surface fluxes. They may also be linked to limitations of the turbulence scheme in accurately representing thermal turbulence near the top of the boundary layer. An LES configuration,

in which turbulence is largely resolved rather than parameterised, such as that used by Mazoyer et al. (2017), provides a more realistic representation of the dynamics near the top of the boundary layer.

Overestimation of TKE in fog simulations is a known issue (Zilitinkevich et al., 2008). According to Shao et al. (2025), improving the surface flux parameterization could be one way to improve the TKE representation by weakening the wind shear contributions. However, this is beyond the scope of the present study.

For IOP 11, Meso-NH's temperature and wind profiles are in fairly good agreement with observations until 04:00 UTC. Around 03:00 UTC, as the warm front approaches, the wind shifts southward and experiences a significant increase in TKE, which the model reproduces to some extent. Beyond 03:00 UTC, however, while the increase in wind intensity and TKE is reproduced, discrepancies become more pronounced. Consequently, this study focuses specifically on the fog formation and development of IOP 11 up to 03:00 UTC.

Overall, Meso-NH quite accurately reproduces the thermodynamic (temperature and vapour mixing ratio) and dynamical (wind intensity and direction) profiles, while overestimating TKE, especially near the surface, for both IOPs 6 and 11. This validation is crucial for the subsequent microphysical analysis, as highlighted by Costabloy et al. (2025), which shows that the microphysical profile is closely related to the thermodynamic state of the fog.

The next step is therefore to categorize the fog into different phases based on its adiabatic state, using the methodology developed by Costabloy et al. (2025) and Dione et al. (2023), which integrates ground instruments and remote sensing data.

4 Fog phases determination

Fog analysis is categorized according to the phases of fog evolution (Dione et al., 2023; Costabloy, 2024). The initial development phase begins when the horizontal visibility falls below 1000 m, indicating the formation of a thin fog layer on the ground, associated with stable conditions. This is followed by a transition phase, during which the fog layer becomes thick enough to radiate sufficiently in the longwave spectrum, warming the surface. The fog then evolves into a well-mixed layer. The third phase, called the adiabatic phase, corresponds to the life cycle of this well-mixed fog until approximately 30 min before dissipation, followed by the final dissipation phase, which lasts until approximately 30 min after the fog has lifted.

To characterize the thin-to-thick transition phase, we follow the methodology developed by Costabloy et al. (2025) and Dione et al. (2023) applying specific thresholds to characterize:

- The inversion destruction near the surface with a TKE at 3 m exceeding $0.1 \text{ m}^2 \text{ s}^{-2}$ (Dhangar et al., 2021).

IOP6

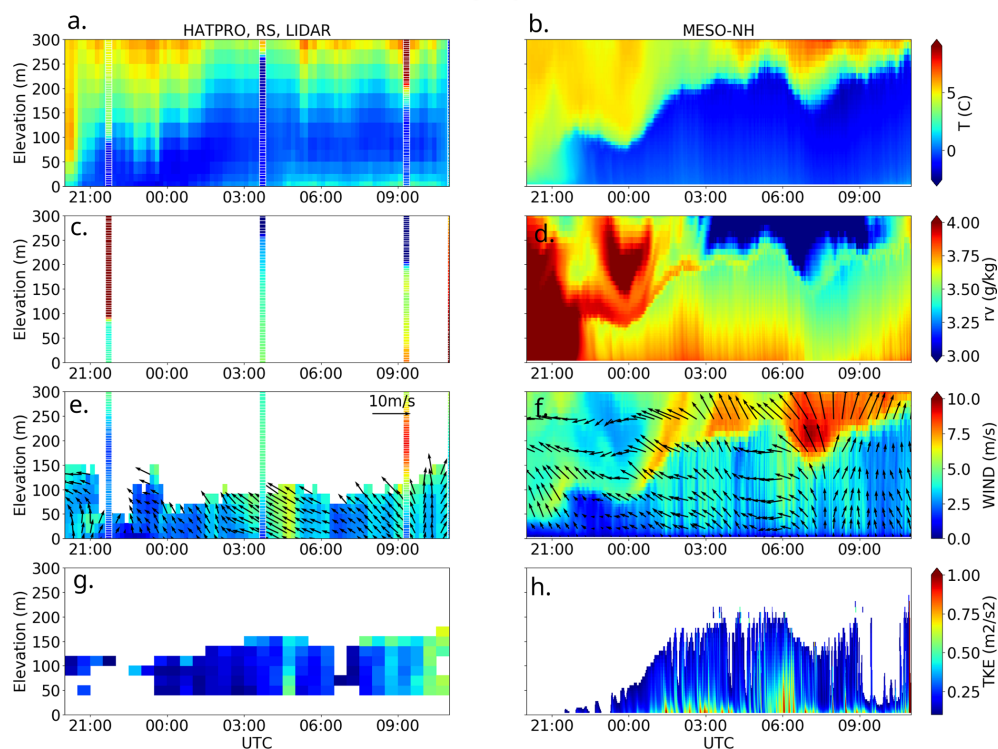


Figure 4. IOP 6: temporal evolution of vertical profiles of temperature (a, b), vapour mixing ratio (c, d), wind intensity and horizontal direction with north upward (e, f) and TKE (g, h) as measured by HATPRO (a), from RS (a, c, e), by lidar (e, g) and simulated by Meso-NH (b, d, f, h). The modeled TKE adds resolved and subgrid contributions.

IOP11

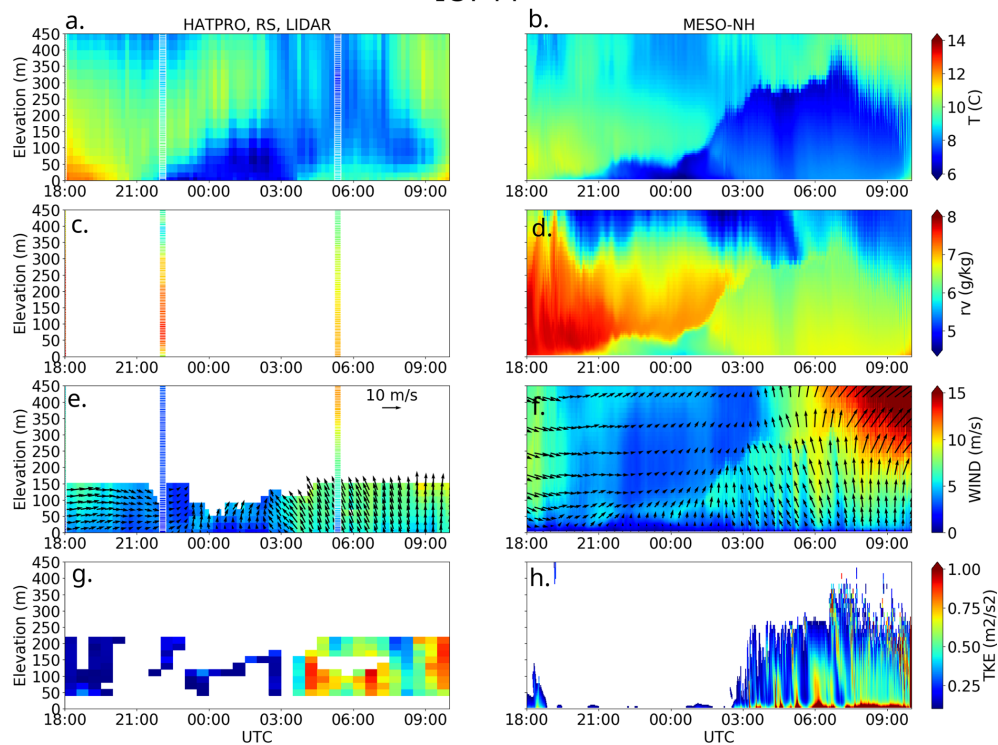


Figure 5. Same as Fig. 4 for IOP 11.

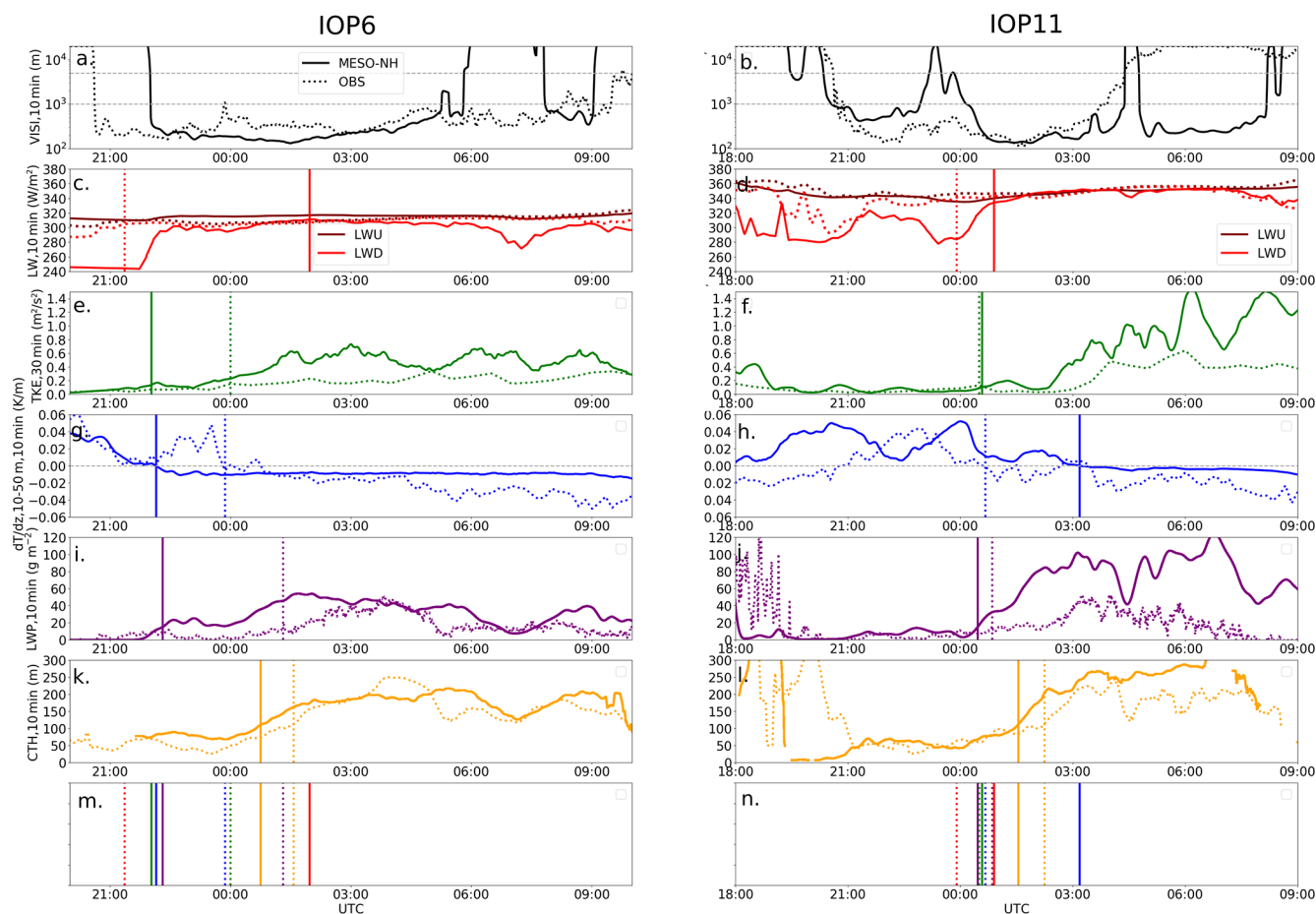


Figure 6. Temporal evolution of (a, b) visibility (c, d) downward (light red line) and upward (dark red line) longwave radiative fluxes, (e, f) TKE (green) with the modeled TKE adding resolved and subgrid contributions, (g, h) vertical gradient of temperature between 10 and 50 m retrieved from the HATPRO microwave radiometer, (i, j) LWP retrieved from the microwave radiometer, (k, l) CTH retrieved from the BASTA cloud radar, at Charbonnières. (m, n) Thin-to-thick transition times estimated for Meso-NH (plain lines) and the observations (dotted lines), using thresholds from the literature for all previous variables.

- The shift of the cooling area to the top of the fog layer with a negative vertical temperature gradient within the 10–50 m layer.
- The fog becoming optically thick with an absolute net longwave radiation difference ($|LW_{net}|$) lower than 5 W m^{-2} that can also be described by an LWP greater than 15 g m^{-2} according to Costablow et al. (2025) and a CTH greater than 110 m according to Wærsted et al. (2017).

Fields are previously averaged over a 10 min sliding window to reduce variability, except for TKE, which has a 30 min temporal resolution in order to avoid the influence of side-to-side horizontal sloshing of the fog on TKE measurements during the stable phase.

Before determining the phases, we first compare the characteristic fields between observations and simulations, and between the two IOPs.

Figure 6 shows the evolution of these quantities for IOP 6 and IOP 11, comparing observations with simulations. In IOP 6, visibility decreases abruptly in both observations and simulation due to advective contributions, whereas in IOP 11, the decrease is more gradual. After its formation, the evolution of fog during IOP 6 follows a classical radiation fog life cycle in both series: visibility drops to a minimum before gradually increasing, similar to the case previously studied by Mazoyer et al. (2019). In contrast, for IOP 11, the model temporarily dissipates the fog around 23:00 UTC, while the observations maintain it, with subsequent differences in evolution attributed to the approaching warm front.

Some discrepancies in the LW fields result from variations in fog formation between observations and simulations for both IOP 6 (before 22:00 UTC) and IOP 11 (before 00:00 UTC). However, as the fogs develop and become optically thick for LW radiation, LW downwelling (LWD) approaches LW upwelling (LWU) values, resulting in quasi-

similar intensities in both observations and model outputs. For IOP 6, the simulated LWnet is larger in absolute value than the observed values, which is likely due to cloud effects, with additional contributions from other sources of error, such as temperature errors, water vapour absorption or aerosol concentrations. Notably, the radiative effect of aerosols is not included in these simulations.

Before the fog and during its formation, the observed and simulated TKE are negligible for both IOPs. During fog development, TKE increases, but simulated values systematically overestimate observations, as highlighted above.

The mean temperature gradient (dT/dz) between 10 and 50 m reflects the air mass stability. For IOP 6, this gradient quickly becomes negative, due to fog formation via advection, even faster in the simulation than in the observations. In contrast, the period with a positive gradient lasts longer for IOP 11. During fog development, the thermal gradient is negative in all data sets, but the simulated values are greater than the observed ones, which explains the overestimation of simulated TKE by thermal production.

Regarding the fog cloud, the model systematically overestimates LWP (Fig. 6i, j), as highlighted above. In contrast, CTHs are fairly well reproduced. This may be due to excessive water production or insufficient depletion. The periods of overestimation of the LWP coincide fairly well with those of overestimation of the TKE. This may indicate a positive feedback, i.e. excess TKE produces excessive droplet activation, and excess droplets limit the negative temperature gradient, favouring the thermal production of TKE.

Focusing on the thin-to-thick transition phase, the different thresholds are shown in Fig. 6m, n: as expected, the observed ones are in good agreement with Figs. 9 and 10 of Costabloy et al. (2025); Dione et al. (2023), with slight discrepancies. These differences arise from two main factors: firstly, Costabloy et al. (2025) calculated the temperature gradient between 50 and 25 m, whereas we used the mean gradient between 50 and 10 m; secondly, we employed different sliding average methods with 30 and 10 min windows. As seen in Fig. 6, observational data fluctuate (e.g., LWP or temperature gradient), meaning small computational differences can lead to significant differences in threshold determination. Additionally, the slow evolution of some variables (e.g., LWnet, CTH, and TKE) emphasizes the importance of accurate threshold selection.

For both IOPs, in the observations, the earliest threshold is given by LWnet and the latest by CTH, with more than a 4 h gap for IOP 6, and a 2 h gap for IOP 11. In simulations, due to the rapid formation by advection in IOP 6, the TKE, thermal gradient and LWP thresholds coincide almost immediately after formation, while the CTH and LWnet thresholds are reached after 3 and 4 h, respectively. For IOP 11, the beginning of the thin-to-thick transition period is also given by the TKE and LWP thresholds, with the last threshold coming from the thermal gradient. Table 2 summarizes the different phases for IOP 6 and IOP 11, highlighting both observed and

modeled characteristics. Phase durations are relatively similar between model and observation, confirming the correct thermodynamic and dynamic representation, except for the adiabatic phase of IOP 11, which is affected by the approaching warm front.

As the times defined by the thresholds applied to the different parameters do not coincide, the first and last times are used to define the thin-to-thick transition phase. Now that the phases in the fog's life cycle have been determined, the validation of microphysics according to these phases can be addressed.

5 Microphysical evaluation

This section focuses on the microphysical evaluation of the simulated fog events, based on a comparison with the vertical profiles of LWC and N_c , as derived from CDP observations under the tethered balloon. First, we briefly evaluate the representativeness of the CDP range diameter measurements by comparing the reflectivity derived from the CDP with the reflectivity measured by the BASTA radar. Next, we analyse the microphysical vertical profiles and DSDs throughout the fog events and group them according to their respective evolutionary phases. Finally, we conduct a budget study to shed light on the microphysical processes involved throughout the fog life cycle, and to improve our understanding of the discrepancies between observations and simulation.

We note that a precise description of the observed vertical profile of LWC throughout the fog life cycle has been carried out in Costabloy et al. (2025), and a subsequent paper will hopefully analyze the observed N_c profiles and DSDs.

5.1 Representativeness of the CDP range diameter measurements

In this section, we first revisit Fig. 2 and compare the reflectivity derived from the CDP with that measured by the BASTA radar. This enables us to evaluate the range of droplet diameters present for each IOP. Since reflectivity depends on the sixth-order moment of the DSD, it is particularly sensitive to the largest droplets. The lower reflectivity derived from the CDP than from the BASTA suggests that there may have been droplets larger than the 50 μm upper limit of the CDP's measurement range. However, BASTA sometimes appears to underestimate reflectivity for reasons that must be explored. Also, note that, without proper instrumentation to measure droplets larger than 50 μm , it is impossible to determine their relative importance to N_c and LWC, since reflectivity is highly sensitive to even a small number of large droplets.

Figure 2 shows that, during IOP 6, the reflectivity derived from the CDP is lower than that derived from the BASTA, particularly between 02:00 and 05:00 UTC. However, during IOP 11, there is good agreement between the reflectivity derived from CDP observations under the tethered balloon

Table 2. Time range (UTC) of observed and modeled fog phases (in brackets the duration).

	Formation	Thin-to-thick transition	Adiabatic	Dissipation
IOP 6 OBS	20:37–21:22 (45 min)	–Day+1 01:34 (4 h 12 min)	–08:54 (7 h 20 min)	–09:54
MESO-NH	22:01–22:02 (1 min)	–Day+1 01:58 (3 h 56 min)	–08:32 (6 h 34 min)	–09:32
IOP 11 OBS	20:38–23:54 (3 h 16 min)	–Day+1 02:14 (2 h 20 min)	–03:25 (1 h 11 min)	–04:25
MESO-NH	20:34–Day+1 00:28 (3 h 54 min)	–03:11 (2 h 43 min)	–07:40 (4 h 29 min)	–08:40

and BASTA measurements during the formation and thin-to-thick transition phases (until 02:14 UTC).

This suggests that, despite numerous reverse reflectivity profiles reported by the CDP being a potential signature of the collision-coalescence process, droplets larger than 50 μm in diameter during the formation and thin-to-thick transition phases of IOP 11 may not have been so numerous. However, more of them may have been produced during IOP 6.

As LIMA has no specific drizzle parameterization, as in Khairoutdinov and Kogan (2000), no drizzle is modeled in IOP 6 (confirmed by the low simulated reflectivity value). However, a very low rain content is modeled in IOP 11. Although there are discrepancies between the model and the observations regarding drizzle and rain, the simulated droplet size shown later falls within the range of the CDP measurements. This enables a meaningful comparison to be made between the model and the observations within this diameter range.

5.2 General comparison of the vertical profiles

Figures 7 and 8 compare the vertical profiles of LWC and N_c measured by the CDP aboard the tethered balloon at Charbonnières with those simulated by the model for IOP 6 and IOP 11, respectively. For IOP 6, the tethered balloon was not launched during the formation phase, preventing an evaluation of the model at this stage. During the thin-to-thick transition, the vertical profiles of both LWC and N_c are initially quite vertically uniform. However, as the transition progresses, higher observed LWC values shift upward within the fog layer after 00:00 UTC. The model overestimates LWC values during this phase, particularly in the second part of the transition, where simulated LWC reaches 0.45 g m^{-3} compared to 0.3 g m^{-3} observed near the fog top, contributing to the LWP overestimation. Similarly, observed N_c also shift upward, with peak values reaching 200 cm^{-3} near the fog top. The model, in contrast, shows a too uniform N_c profile around 100 cm^{-3} and fails to capture the higher droplet concentrations observed at the fog top. During the adiabatic phase, observed LWC peaks near the fog top, with maximum values around 0.3 g m^{-3} , while the model continues to overestimate it, reaching 0.45 g m^{-3} , but a bit less than during the previous phase. After 04:00 UTC, simulated LWC decreases and becomes more aligned with observations, although still slightly overestimated. Observed N_c also peaks

at the fog top, reaching up to 500 cm^{-3} , likely due to droplet activation driven by radiative cooling. However, the model still produces too vertically uniform N_c profiles, with less pronounced activation at the top and mainly stronger values closer to the ground. Droplet concentrations increase significantly throughout the fog layer during the adiabatic phase, more than 200 cm^{-3} . The decrease in droplet concentration modeled towards 05:00 UTC is likely due to advection, as confirmed by the subsequent budget analysis. Some of the profiles observed deserve to be described in more detail. For example, the descent at 03:00 UTC shows low LWC/ N_c values associated with high reflectivity (Fig. 2a), suggesting enhanced coalescence and sedimentation. At 06:00 UTC during ascent, a strong LWC/ N_c peak with reasonable reflectivity was observed in the upper part, possibly due to intense top-mixing. During the dissipation phase, observed LWC decreases and becomes more homogeneously distributed, while the model continues to overestimate LWC in the upper fog layers. N_c concentrations rise again towards the top in the observations, but the model overestimates the fog depth, while producing correct values in the first 150 m. In summary for IOP 6, the observed LWC and N_c profiles evolve similarly. Initially, they are uniform profiles, followed by an increase at the fog top. However, the model consistently overestimates LWC over the fog layer and produces overly uniform N_c profiles by overestimating N_c near the ground and underestimating the peaks at the fog top.

For IOP 11, the focus is primarily on the formation and thin-to-thick transition phases. During the formation phase, both observations and simulations show a reverse LWC gradient, with maximum values (0.2 g m^{-3}) near the surface, and low N_c concentrations below 100 cm^{-3} , underestimated by the model. During the thin-to-thick transition phase, observed LWC and N_c vertical profiles become nearly vertically uniform, although the reverse gradient persists at 02:00 UTC for LWC. A short-lived but significant N_c peak (300 cm^{-3}) is measured near the fog top at 01:00 UTC, but overall N_c concentrations remain relatively low. The model strongly overestimates LWC in this phase, with values exceeding 0.5 g m^{-3} near the top, compared to observed values closer to 0.2 g m^{-3} . The model also simulates an N_c gradient, with peaks (300 cm^{-3}) near the fog top – this was not seen in IOP 6 simulation. As the transition continues, LWC remains overestimated in the model, and simulated N_c becomes more vertically uniform.

IOP6

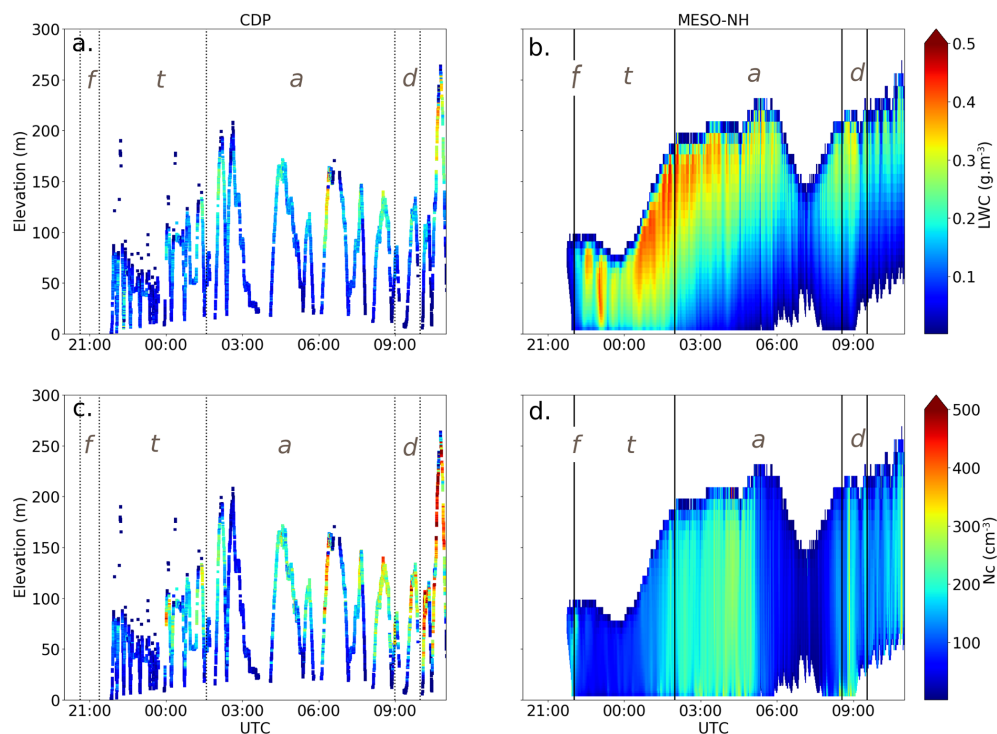


Figure 7. IOP 6: vertical profiles of LWC in g m^{-3} (a, b) and N_c in cm^{-3} (c, d) as measured by the CDP under the tethered balloon (a, c) and simulated by Meso-NH (b, d). The black vertical lines separate the fog phases, identified with grey letters (f: formation, t: thin-to-thick transition, a: adiabatic, d: dissipation).

IOP11

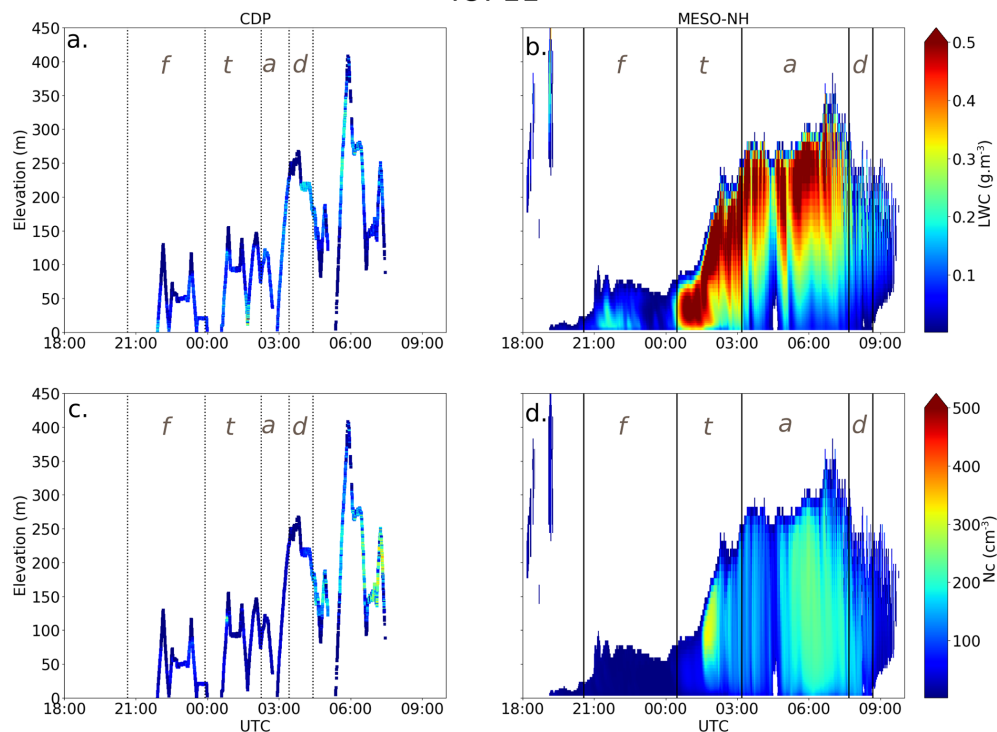


Figure 8. Same as Fig. 7 for IOP 11.

In summary, for IOP 11, the model overestimates the LWC across the entire fog depth, particularly in the upper part of the fog layer, significantly more than in IOP 6. For the two cases, this overestimation is consistent with the overestimation of LWP, and more pronounced in IOP 11 than IOP 6, while the CTH is correctly reproduced (Fig. 6). On the other hand, simulated N_c values fall within the correct order of magnitude, i.e. $\sim [10\text{--}250]\text{cm}^{-3}$. Notably, the model now avoids the systematic N_c overestimation reported in previous studies (Mazoyer et al., 2017; Ducongé et al., 2020). This improvement stems from the refined droplet activation parameterization introduced by Vié et al. (2024), particularly the consideration of continuous growth of existing cloud droplets by condensation.

For IOP 11, the LWC overestimation is consistent with the overestimated radar reflectivity values, whereas for IOP 6, the LWC overestimation in the fog layer contrasts with the previously noted underestimation of reflectivity. This suggests a potential issue with the DSD (including drizzle), as previously noted.

It is worth noting that the low droplet concentrations observed during IOP 11 may be influenced by oceanic air masses and the presence of large hygroscopic sea salt particles. These factors may limit activation under low supersaturation conditions (Barahona et al., 2010), while still allowing for vapour deposition and LWC growth.

To gain a better understanding of the model's limitations, we will analyse the vertical profiles of LWC and N_c , averaged by phase, to provide an overview of how the DSD vertical profile evolves. This will then be used to track the microphysical processes involved in the observations, as well as those identified in the previous thermodynamic analysis. We will then compare these with the processes involved in the simulation. This will help us to explain the processes behind the model misrepresentation.

5.3 Vertical profiles and DSDs averaged by phase

Figures 9 and 10 present the vertical profiles of temperature, LWC, and N_c averaged over the different fog phases, as well as the corresponding DSD, averaged over different vertical layers, for IOP 6 and IOP 11, respectively. This section aims to assess the representativeness of phase average with regard to the vertical profiles of temperature, LWC and N , in order to further evaluate DSD evolution.

During IOP 6, some discrepancies are observed between the temperature profiles measured by the HATPRO radiometer and those simulated by the model, across all fog phases. In the thin-to-thick transition phase, the model exhibits a warm bias near the surface, which had already been identified in previous comparisons with radiosoundings. In contrast, during the adiabatic phase, the radiometer indicates a cold bias relative to the simulation, even more marked during the dissipation phase. These inconsistencies may stem partly from limitations inherent to radiometric retrievals near the sur-

face, where the vertical resolution is coarse and the radiative transfer assumptions may be less accurate. Nonetheless, despite these uncertainties, the overall trends in the temperature profiles remain almost consistent between observations and simulations. Specifically, they reveal the formation and vertical growth of a well-mixed adiabatic layer during the transition from the thin to thick layer, and throughout the dissipation phase. Regarding LWC and N_c , IOP 6 exhibits significant variability during the thin-to-thick transition due to differences between the sub-phases, i.e. before and after 00:00 UTC. Averaging the 23 vertical profiles smooths out these changes, resulting in a nearly uniform LWC and a slight increase in N_c with height. While the model captures the structure, it overestimates the LWC at all levels (mean bias: $0.17 \pm 0.04\text{ g m}^{-3}$), which is consistent with the wet bias identified in Fig. 3. It also overestimates the N_c close to the surface. However, a comparison with the observed standard deviation shows a tendency to underestimate N_c in the upper layer, which is consistent with Fig. 7. During the adiabatic phase (16 observed vertical profiles) and the dissipation phase (3 observed vertical profiles), both the observed and the simulated profiles show a peak in LWC at the fog top, with similar vertical evolution. The observed N_c peaks at the top, while the modeled N_c remains uniform vertically.

During IOP 11, the simulation tends to be slightly too cold during the formation phase, which is also consistent with radiosounding data. In both observations and simulations, the temperature profiles capture the erosion of the near-surface stable layer during the formation phase, followed by the development of a well-mixed adiabatic layer during the thin-to-thick transition. During the formation phase, the average of the five observed profiles reveals a surface LWC peak and a vertically uniform N_c , both with low dispersion. While the model captures the profile shapes, it overestimates LWC (mean bias: $0.006 \pm 0.04\text{ g m}^{-3}$) due to the cold bias and significantly underestimates N_c . During the thin-to-thick transition, the average of the three observed profiles shows less dispersion than in IOP 6, probably due to the less numerous profiles. LWC tends towards a quasi-uniform profile, while retaining a surface peak; N_c remains vertically uniform. The model displays strong variation and overestimates both LWC (mean bias: $0.1 \pm 0.09\text{ g m}^{-3}$) and N_c (mean bias: $103 \pm 43\text{ cm}^{-3}$), particularly at the fog top. In summary, and in line with previous analyses, averaging N_c and LWC by phase confirms that, while Meso-NH captures the relative vertical profile of LWC quite well, it systematically overestimates the values during the thin-to-thick transition and the adiabatic phase for both IOPs. Thermodynamic biases contribute to this overestimation. It also produces N_c profiles that are too homogeneous vertically. This phase-based aggregation approach is considered reliable. The vertical profile of DSD is now described. DSD is also averaged by depth into three vertical layers, each 75 m in height.

During the thin-to-thick transition phase of IOP 6, the $10\text{ }\mu\text{m}$ droplet diameter mode is firmly observed across the

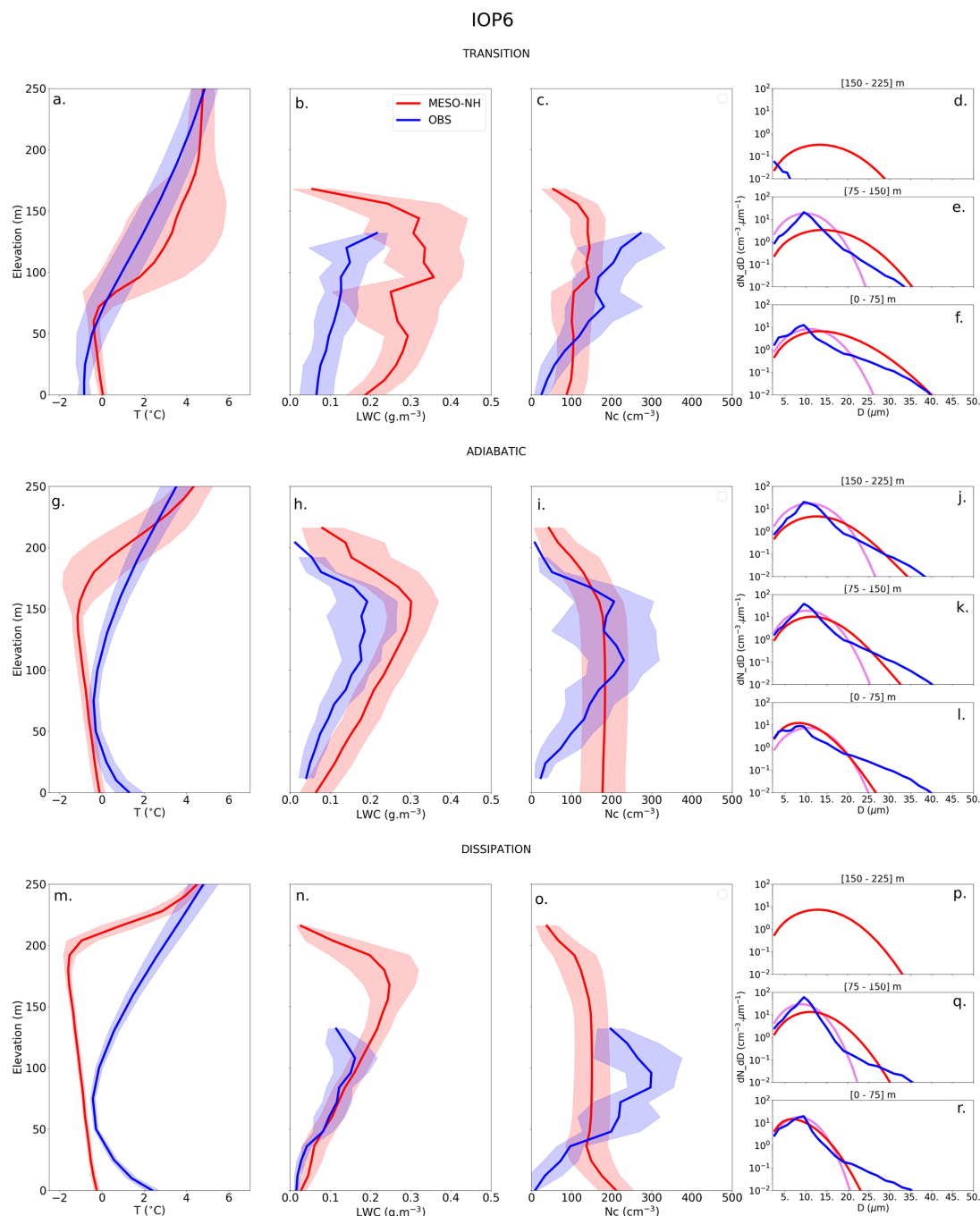


Figure 9. IOP 6: vertical profiles averaged by phases (23 for the thin-to-thick transition for the first line, 16 for the adiabatic for the second line, 3 for dissipation for the third line) of: Observed and simulated (**a**, **g**, **m**) absolute temperature, (**b**, **h**, **n**) LWC in g m^{-3} , and (**c**, **i**, **o**) N_c in cm^{-3} . Panels on the right represent droplets distribution integrated over different layers. Model results are in continuous red and observation in dashed blue. For the DSDs, the pink line represents the distribution that the model would represent with the observed LWC and N_c . The envelopes for the vertical profiles represent $\pm$ the standard deviation.

entire layer, which indicates that activation has occurred throughout. However, the droplet concentration within this mode is higher at the top of the fog, in the [75–150] m layer. This is consistent with N_c evolution during the second sub-phase of the transition (Fig. 7), as well as with stronger ac-

tivation at the top of the fog layer, which is probably the result of radiative cooling. Conversely, small- to medium-sized droplets (up to $30\text{ }\mu\text{m}$ in diameter) are more prevalent in the bottom layer, which is likely due to substantial condensational growth. This is consistent with LWC evolution during

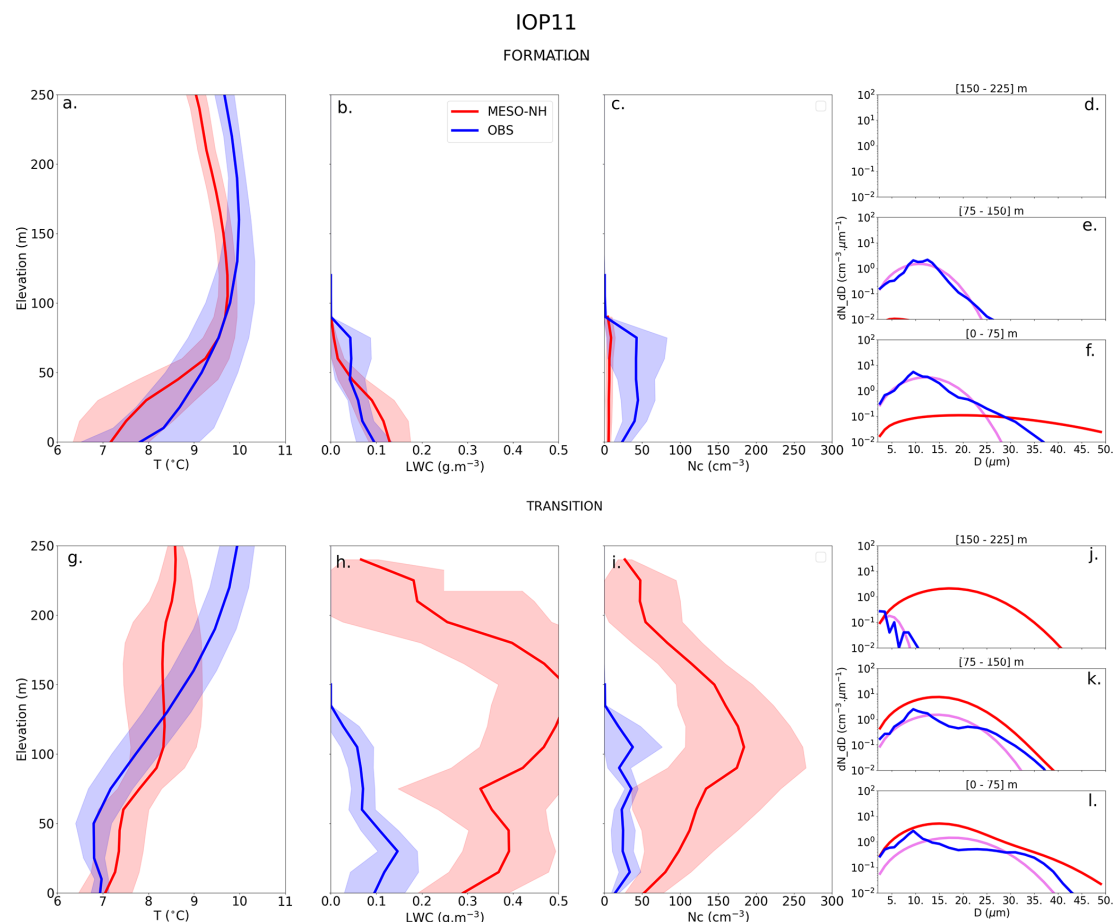


Figure 10. Same as Fig. 9 for IOP 11, but only for the 2 first periods, with 5 profiles averaged for the formation phase for the first line and 3 for the thin-to-thick transition for the second line.

the initial sub-phase of the transition (Fig. 7). Consequently, the observed averaged DSD increases in size towards larger diameters at the bottom. Droplets larger than $30\text{ }\mu\text{m}$ in diameter are recorded in the bottom layer, which suggests gravitational collision-coalescence, and is consistent with the CDP reflectivity analyses shown in Fig. 2.

During the adiabatic and dissipation phases, the $10\text{ }\mu\text{m}$ activation mode remains well established throughout the entire layer, and more pronounced at the top. This suggests continuous activation within the fog layer due to adiabatic motions, and to radiative cooling at the top of the fog layer. The concentration of small- to medium-sized droplets evolves slightly vertically, and large droplets are now present throughout the fog layer, with an increased concentration towards the bottom layers. A key finding of this analysis is the relative stability of small- to medium-sized droplets (up to $30\text{ }\mu\text{m}$ in diameter) in the lower fog layer compared to the upper layers throughout the adiabatic and dissipation phases, despite an increase in LWC towards the top. A $dM/d\log D$ (with M the mass) representation (not shown) reveals that differences in LWC are predominantly driven by the $10\text{ }\mu\text{m}$

mode diameter. This is consistent with the very high concentration of small droplets compared to small- to medium-sized ones (several orders of magnitude) in the $dN/d\log D$ representation. This indicates that activation and direct subsequent condensation growth are of primary importance in representing fog microphysics.

A comparison with the DSD from the model shows that, during the thin-to-thick transition phase, the modeled DSD increases in size towards larger diameters at the bottom, representing a similar range of droplet diameters to that observed. However, the model misses the $10\text{ }\mu\text{m}$ droplet mode and overestimates the concentration of small- to medium-sized droplets. During the adiabatic and dissipation phases, the model still struggles to represent the $10\text{ }\mu\text{m}$ diameter mode separately in the lower layer. It also misses the biggest droplets, particularly near the ground, explaining the underestimation of the radar reflectivity. Contrary to observations, the DSD depicts a decrease in the concentration of small- to medium-sized droplets in the bottom layer, despite the absence of large droplets. The increase in LWC towards the top is related to the enlargement of the DSD. The DSDs

presented in Figs. 9 and 10 include the distribution that the model would have with the observed LWC and N_c values represented by the pink line. This is to isolate the error arising from the Gamma law when considering a single-mode. For IOP 6, these distributions provide a more accurate representation of the $10\text{ }\mu\text{m}$ mode, particularly near the fog top. However, they fail to capture the enlargement of the DSD towards larger droplets. Using a single mode for droplets prevents a second mode from forming in large droplets.

The DSD observed during IOP 11 shows that only the $10\text{ }\mu\text{m}$ diameter activation mode is present from top to bottom, during the formation phase. This is present at a lower concentration than in IOP 6. The concentration of small- to medium-sized droplets increases through the bottom layer, probably due to condensation growth caused by surface cooling. Meanwhile, the $10\text{ }\mu\text{m}$ diameter activation mode persists: during the thin-to-thick transition phase, this mode is present throughout the fog layer again. However, a second, very distinct mode appears in small- to medium-sized droplets, affecting even the larger ones. This mode broadens towards the bottom layers. This is consistent with the reverse reflectivity profile, previously observed during the thin-to-thick transition phase of IOP 11, suggesting collisional growth. The $dM/d\log D$ representation (not shown) reveals that the differences in LWC are predominantly driven by small- to medium-sized droplets and large droplets. Unlike IOP 6, condensation growth and collision coalescence appear to have a stronger impact. The differences in IOP for condensation growth over small, small-to-medium, and large droplets could be explained by differences in N_c concentration. As N_c is lower in IOP 11, there should be less competition for vapor deposition. This may be due to a lower concentration of aerosols, or larger ones. Indeed, sea salt could have been present during IOP 11.

Comparisons with the DSD from the model show that, due to the concentration of droplets being too small during the IOP 11 formation phase, the DSD enlarges towards very large droplets, which should favor modeling rain formation. Consequently, small and medium droplets are missed, as shown by the pink DSDs produced with the observed LWC and N_c applied to the gamma law. During the thin-to-thick transition phase, the DSD enlarges towards the bottom layers, as observed. However, due to the excess in LWC and N_c , the modeled DSD appears too large compared to the observations. Due to structural limitations, LIMA is unable to accurately represent the bimodal nature of the DSD. This may explain why the $10\text{ }\mu\text{m}$ diameter mode is misrepresented during IOP 11 near the bottom of the fog layer, and why the largest droplets ($> 30\text{ }\mu\text{m}$ in diameter) are overestimated.

We will now use a budget analysis to gain a better understanding of the processes contributing to the discrepancies we have just described. Our focus is on identifying the causes of excessive LWC during the thin-to-thick and adiabatic phases, as well as the absence of an N_c vertical gradient, which we explore through a model budget analysis.

5.4 Analysis of the microphysical processes

Figures 11 and 12 present the simulated budgets of rc and N_c for IOPs 6 and 11, respectively. They show the components corresponding to the dynamical (advection and turbulence) and microphysical (evaporation/condensation, hereafter referred to as adjustment, and sedimentation/deposition) processes for the two variables. For N_c , the activation is also considered, as well as in the total tendency. Note that the contributions of self-collection and break-up to the N_c budget are not shown, as they are negligible. Unlike the N_c budget, turbulence is not presented separately in the rc budget (but it is included in the total contribution), as it applies to the conservative variable of total non-precipitating water, rather than to the mixing ratios of vapour and cloud water separately. In addition to improving the understanding of the microphysical characteristics, the budget analysis is also used to establish links with model biases identified in previous sections.

As previously shown, the simulated formation phase of IOP 6 is very short, which is consistent with advective formation; rc and N_c production are driven by transport. Conversely, during the IOP 11 radiative formation, rc production is driven by saturation adjustment, corresponding to a local process. This production is partly offset by sedimentation. N_c is only produced by activation, most likely due to radiative cooling. However, this is offset by transport and evaporation. Consequently, the total N_c tendency is close to zero, resulting in a lack of N_c (Fig. 10c). The main causes are the dry air above fog layer through entrainment/detrainment, inducing evaporation, and the excessive stable temperature profile, which prevents a positive turbulence term.

During the thin-to-thick transition phase, advection and turbulence are important sources of rc and N_c at the fog top, but mainly act as sink terms for both IOPs inside the fog layer. This may reflect the upward transport associated with the formation of weak convective movements during fog development. Saturation adjustment acts as a sink term for both rc and N_c at the fog top due to evaporation, and as an important source for rc by condensation just below the fog top due to radiative cooling.

Within the fog layer, transport can act as a source or sink term for N_c , but always as a sink term for rc . When the two are opposite, this may be due to the transport of a greater number of droplets with a lower liquid water content. Condensation also acts as a source term of rc within the fog layer, albeit less intensely than at the fog top. The condensation process is responsible for the excessive rc production. For IOP 6, this excess is due to high humidity, as shown in Figs. 3b and 4b. For IOP 11, the bias is due to the temperature being too low in the first hundred meters since formation. These biases contribute to the rapid vertical development of the fog, reinforcing the cooling at the fog top. This could generate excessive top-down buoyancy in the fog layer, favouring fog development. Excessive cooling at the fog top also favours the activation and excess of N_c , which is significantly

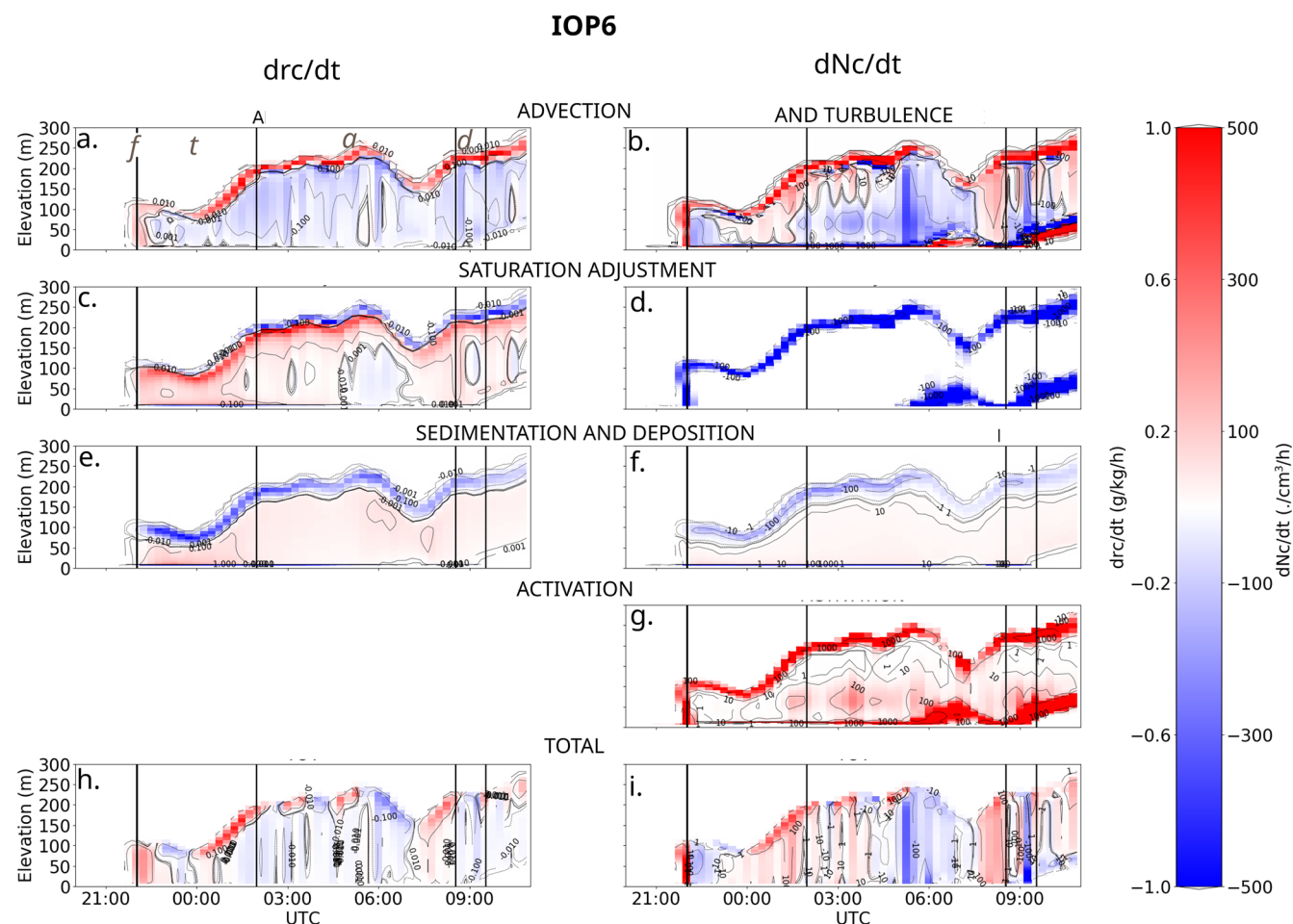


Figure 11. IOP 6: Vertical budget of rc and N_c during the fog life cycle: evolution (**a**: rc , **e**: N_c), due to advection (**b**: rc) and turbulence (**f**: N_c), saturation adjustment (**c**: rc , **g**: N_c), sedimentation and deposition (**d**: rc , **h**: N_c) and activation (**i**: N_c). Black contours indicate budget values of rc and N_c . The black vertical lines separate the fog phases, identified with grey letters (f: formation, t: thin-to-thick transition, a: adiabatic, d: dissipation).

larger for IOP 11 than for IOP 6. Since the transition phase, deposition acts as a sink term just above the ground. Logically, sedimentation acts as a sink term for rc and N_c at the fog top, and a source term within the fog layer. The sedimentation process should mainly contribute to the shift toward larger droplets in the DSD when moving downward within the fog layer for both IOPs (Figs. 9 and 10). This shift is in agreement with observations for IOP 6, while it is largely overestimated for IOP 11 due to excess of rc within the fog layer. Considering the total tendencies of this phase for both IOPs, rc is primarily produced at the top due to advection, and just below the top by condensation. Meanwhile, N_c is primarily produced at the fog top by transport and activation. These effects contribute to fog development.

During the adiabatic phase, which concerns only IOP 6 in terms of comparison to observations, the behavior of the processes is quite similar to that of the thin-to-thick transition phase, albeit more pronounced. We have seen that a

major shortcoming of the simulation comes from an overestimation of N_c near the ground, with an overestimation of droplets between 10 and 20 μm . This could be due to the activation process of N_c , which remains positive in the first 150 m throughout this phase and is not entirely consumed by the evaporation. This overestimation of activation is thought to be a consequence of the overestimation of the TKE near the ground, as seen in Fig. 6e. Conversely, concentrations of droplets between 25 and 40 μm , albeit lower, are missing. As previously mentioned, this is due to the single-mode assumption in DSD, as illustrated by the pink line in Fig. 9j, k, l. An inaccurate representation of the DSD shape could lead to incorrect sedimentation behavior.

During the dissipation phase of IOP 6, transport processes are primarily responsible for the sinks of rc and N_c . The same causes mentioned previously produce the same misrepresentation of DSD.

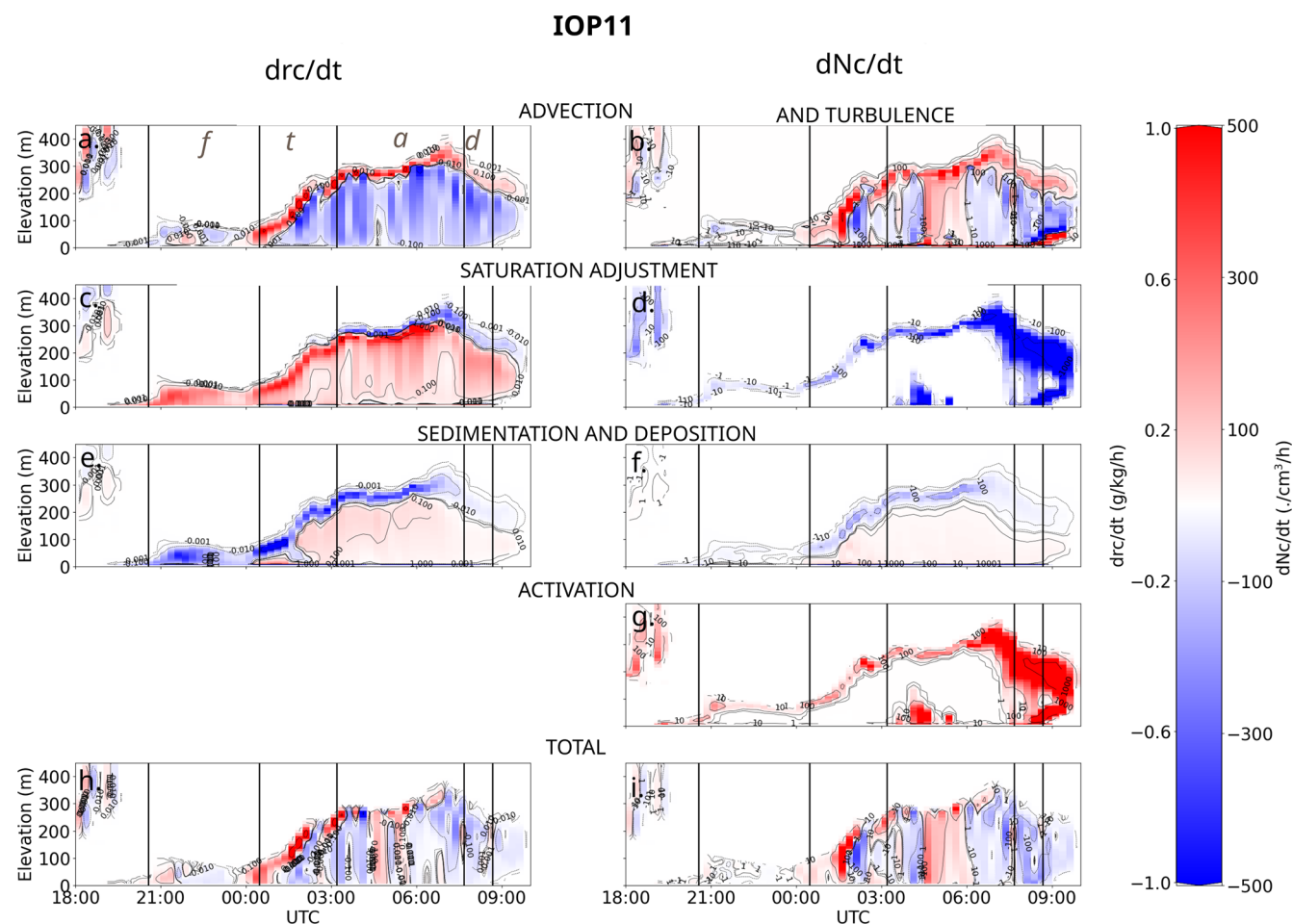


Figure 12. Same as Fig. 11 for IOP 11.

6 Discussion

Comparing simulated fog using the two-moment microphysical scheme LIMA in Meso-NH revealed that both the fog life cycle and the cloud top height were reasonably well simulated, but also highlighted discrepancies between the observations and the simulations. For IOP 6 and IOP 11, biases in the microphysical representation were identified and are now discussed.

First, there is overestimation of LWC during the thin-to-thick transition phase, which can lead to a rapid phase shift (Boutle et al., 2018), and during the adiabatic phase, which can delay fog dissipation (Bott, 1991; Zhang et al., 2014). This also leads to excessive radiative cooling, and subsequent strong activation at the fog top, as well as top-down buoyancy. The overestimated LWC is consistent with the overestimation of LWP for a well-estimated CTH. This is mainly due to thermodynamic biases: a wet bias during IOP 6 and a cold bias during IOP 11. It should be noted that humidity excesses of about 0.5 g kg^{-1} , as simulated during IOP 6, may substantially enhance liquid water content ($\sim 0.6 \text{ g m}^{-3}$

if fully condensed), although this effect is compensated by higher simulated temperatures. Conversely, a negative temperature bias of 1 K, as observed during IOP 11, leads to a change of about 0.5 g kg^{-1} in the saturation mixing ratio, although this effect is compensated by lower simulated water vapor mixing ratios. This strong sensitivity to small thermodynamic biases likely explains, at least in part, why fog remains particularly difficult to forecast.

Other sensitivity tests (not shown) using IOP 6 were conducted to mitigate LWC by reinforcing potential sink terms, particularly sedimentation. Firstly, reducing the aerosol loading slightly decreased N_c with little effect on the LWC or the N_c profile. However, the reduction may not have been strong enough. Secondly, increasing the deposition velocity from 2 to 6 cm s^{-1} (Katata, 2014) halved both the LWC and N_c values in the fog layer, but N_c remained homogeneous on the vertical. It should be noted that an accurate deposition velocity or a parametrization considering TKE, for example, could be a key avenue for improving the absolute values of LWC. During SOFOG3D, Price (2025) measured fog droplet deposition surface fluxes and reported typical

values of $50 \text{ g m}^{-2} \text{ h}^{-1}$. Our simulation produce close values with $20 \text{ g m}^{-2} \text{ h}^{-1}$ for fog droplets deposition flux and $35 \text{ g m}^{-2} \text{ h}^{-1}$ for gravitational settling flux. This is consistent with the previous comparison of observations and simulations conducted during LANFEX by Vié et al. (2024).

Furthermore, the impact of mixing at the fog top, particularly the related top-down mixing, should be better quantified, as highlighted by Yang and Gao (2020). Indeed, LWC is overestimated during IOP 6, whereas the upper layer is either quite wet (in the thin-to-thick transition phase) or dry (in the adiabatic phase) (Figs. 6 and 7). Using instrumentation dedicated to radiation, such as pyrgeometers under tethered balloons, may help to understand the radiative budget at the top of the fog layer, and better quantify top radiative cooling and its impact on top-down mixing. Fog dynamics could also be better qualified using vertical Doppler velocity measurements from BASTA. Initial preliminary results (not shown) reveal a very distinct difference in behavior between fog and stratus.

The second issue concerns the misrepresentation of the DSD, primarily during the adiabatic phase of IOP 6: the absence of droplets with diameters greater than $20 \mu\text{m}$ explains the underestimation of radar reflectivity, despite the overestimation of LWP. This issue did not arise for IOP 11, as the excess LWC allows the full range of droplet sizes to be represented. Failing to detect the largest droplets, as occurred in IOP 6, could result in inaccurate sedimentation modeling, leading to excessive LWC. It could also disrupt the activation processes, according to Bott (1991), by conserving a strong sink term for supersaturation, which may be particularly important at the fog top. This could also affect radiative cooling.

LIMA assumes a generalized gamma DSD, with fixed values for the shape parameters α and ν . Thus, a simple option to modify the DSD is to use different values for these two parameters. The selected DSD shape adheres to a mesokurtic distribution, whereas the observed distributions appear leptokurtic and positively skewed, showing a tendency toward a bimodal configuration (characterized by a scarcity of medium-sized droplets but an excess of both small and large droplets). Using $\alpha = 1$ and $\nu = 3$ for the DSD, instead of the other way around, provided a better fit for the DSD and enabled the representation of a leptokurtic and positively skewed distribution (not shown), yielding reflectivity values closer to the observations. However, this did not correct the N_c profile. A global fit to the entire SOFOG3D dataset could provide interesting insights; however, there is no doubt that these shape parameters would depend on aerosol loading and fog phases, which would make reaching a consensus difficult. Adding a third moment to the distribution, such as reflectivity, might allow for a better representation (Milbrandt and Yau, 2005). Ultimately, the use of bin schemes, as in Lábó and Geresdi (2016), should definitely help to constrain the shape of the DSD.

Our study has shown that the misrepresentation of DSD is partly due to the single-mode assumption. A bimodal DSD

is a common feature in fog, as observed in both IOP 11 and 6, and has been cited in previous studies (Choulaton et al., 1981; Gerber, 1991). Whether this second droplet mode serves as a critical LWC sink remains an open question. There have been numerous observations of drizzle in fog (Okita, 1962; Wendisch et al., 1998; Frank et al., 1998; Gul-tepe et al., 2007), including an observational stratocumulus campaign with specific drizzle instrumentation and measurements as described in VanZanten et al. (2005). The DSD distribution on the CDP diameter range was very close to that of fog. During IOP 6, vertical Doppler velocities of up to 0.3 m s^{-1} downwards (not shown) were associated with high reflectivity values (-15 dBZ). However, the fall velocity of a $50 \mu\text{m}$ diameter droplet (the upper threshold of the CDP) does not exceed 0.1 m s^{-1} . Additionally, the reflectivity derived from the CDP exhibited lower values than the BASTA ones during this IOP. A comparable underestimation of reflectivity was also observed for IOP 14 in Bell et al. (2022), indicating the absence of larger droplets. According to Matrosov et al. (2004), drizzle typically occurs when reflectivity values are higher than -20 dBZ . Such values appear to have been reached for both IOPs. Therefore, everything seems to indicate the presence of large droplets (greater than $50 \mu\text{m}$), which the instruments were unable to observe. In fact during SOFOG3D, a MPS (Meteorological Particle Spectrometer) from Droplet Measurement Technologies had measured particle size distribution ranging from 25 to $125 \mu\text{m}$. However, no validation has yet been performed during the campaign. Very early results suggest that droplets larger than $50 \mu\text{m}$ were present throughout, occasionally growing to sizes of up to $200 \mu\text{m}$ when reflectivity reached -20 dBZ . Further analysis and measurement of droplets larger than $50 \mu\text{m}$ in diameter may help to determine the effect of drizzle on reflectivity, LWC and N_c in fog.

There are two main hypotheses that seek to explain the bimodal DSD, but no consensus has been reached. The first hypothesis involves mass transfer from the first droplet mode to the second via gravitational and turbulent collisional coalescence (Xue et al., 2008; Zhao et al., 2013) or Ostwald ripening (Wendisch et al., 1998). The second hypothesis suggests that giant CCN, such as large sea-salt particles, are activated in the second DSD mode (Richter et al., 2021), which could also form from un-activated large hydrated particles (Frank et al., 1998). Meanwhile, the activation of aerosol particles from the accumulation mode would create the first DSD mode. This scenario is consistent with IOP 11, which shows a bimodal DSD and a maritime influence on aerosol loading. However, considering the DSD broadening from top to bottom during IOP 11, it seems unlikely that the effects of aerosols and Ostwald ripening are confined to the lowest level. Consequently, more weight could be given to the gravitational collision-coalescence theory for this specific case.

Currently, the LIMA scheme does not have the capability to model a bimodal DSD and two-mode activation. Although efforts have been made to implement the Khairoutdinov and

Kogan (2000) parameterization, which is better suited to drizzle formation than the original rain parameterization in LIMA, little success has been achieved using it. Modifications would be required to correctly position the drizzle mode in fog observations. Nevertheless, three-class parameterizations for warm microphysics, representing droplets, drizzle and rain could fill this gap, as demonstrated in Sant et al. (2013). In a similar way, the microphysics scheme of Saleeby and Cotton (2004), developed within the Regional Atmospheric Modeling System (RAMS), distinguishes between two categories of droplets: cloud1 (2–40 μm diameter) and cloud2 (40–80 μm diameter). Cloud1 results from CCN activation, while cloud2 forms through cloud1 autoconversion and giant CCN (type sea-salt) activation. Using a second droplet mode could potentially reduce supersaturation by providing more surface area for the first mode to grow. The authors also noted improved precipitation representation. Finally, a subgrid parameterization for drizzle, similar to that in Turner et al. (2012) or Marcel et al. (2026) for precipitation, could also be beneficial. Ultimately, a bimodal DSD representation could offer a better representation of the 10 μm diameter mode and radiative cooling at the fog top.

The third issue concerns the failure to capture the N_c gradient during the adiabatic phase, emphasizing the inaccurate DSD representation at the lowest levels. Regarding the N_c gradient at low droplet concentrations near the bottom layer, IOP 6 exhibits this gradient, whereas IOP 11, which has a lower aerosol loading, shows a nearly uniform profile in the observations. However, numerous case studies observed in SOFOG3D have indicated the presence of this gradient during adiabatic phases (not shown). The differences between IOP 6 and IOP 11 should be explored further to assess potential links between this gradient and aerosol loading.

The homogeneous vertical profile of N_c during the transition and adiabatic phases may be due to intense low-level activation caused by overestimation of TKE or vertical velocities within the first 150 m, which limits the formation of the vertical gradient. The timing and location of low-level activation coincide with high simulated TKE values, which may enhance supersaturation. An experiment in which the contribution of TKE to activation was set to zero produced a vertical N_c profile with slightly decreasing concentrations towards the surface. However, the gradient still deviated significantly from the observed one. Furthermore, as discussed by Thouron et al. (2012), the maximum supersaturation diagnostic used in this study is highly effective in the case of convective clouds. However, this assumes that the turbulent structure responsible for supersaturation has a sufficiently long lifespan to allow maximum supersaturation to be reached, which cannot be guaranteed in fog. Furthermore, when considering fine vertical resolution, activation will occur in the mesh crossed by the turbulent structure. However, this is not necessarily where maximum supersaturation is reached. Diagnosing maximum supersaturation in fog can lead to an overestimation of supersaturation,

particularly where vertical velocities are strongest. New formulations for supersaturation calculations and activation in fog are emerging, such as Peterka et al. (2024), which does not directly use vertical velocity, Field et al. (2023), which weights the TKE contributing terms, and Martín Pérez et al. (2024), which considers a sink term for supersaturation of large CCN. It should be noted that considering the continuous growth of existing cloud droplets through condensation as a sink term for supersaturation, as we did, may be overly significant in cases where excess LWC is modeled, which is a frequent bias in fog modeling. This consideration arises from the use of maximum supersaturation diagnosis, whereas other methods, such as prognostic supersaturation (Thouron et al., 2012), calculate supersaturation explicitly. The latter approach has been shown to be useful as a reference in LES fog simulations (Vié et al., 2024). However, this requires very high spatial and temporal resolution.

7 Conclusion

The present study forms part of the SOFOG3D field campaign, which aims to improve the assessment of NWP models' ability to represent subgrid-scale microphysical processes in fog. It provides a detailed and consistent evaluation of the two-moment microphysical scheme LIMA on the life cycles of two fog events during IOPs 6 and 11. Two 48 h simulations were performed, on a 100 m horizontal grid model covering an area of 60 km $\times$ 40 km, centred on the main observation site. IOP 6 was an advective radiative fog, whereas IOP 11 was a radiative fog whose evolution was affected by an approaching warm front. This study leverages the unique SOFOG3D dataset, which combines vertical measurements from remote sensing instruments and a tethered balloon equipped with a cloud droplet probe, to derive LWC and droplet concentration. To the authors' knowledge, this is the first study to evaluate vertical microphysical profiles simulated with observations throughout the entire fog life cycle.

The first part of the study focused on evaluating the fog life cycle using remote sensing (BASTA radar, HATPRO radiometer, and WindCube wind lidar), as well as radiosoundings. Fog evolution as inferred from simulated reflectivity generally agrees with observations for both IOP 6 and IOP 11, particularly after formation for IOP 6 and until 04:00 UTC for IOP 11. Meso-NH quite accurately reproduces the thermodynamic (temperature and vapour mixing ratio) and dynamic (wind intensity and direction) characteristics of the fog, but develops a wet bias for IOP 6 and a cold bias for IOP 11, and overestimates the TKE profiles in both cases.

The second part of the study focused on determining the fog phases, including the thin-to-thick transition period, according to Dione et al. (2023) and Costablos et al. (2025). The model shows a fairly good agreement with the observations, enabling a microphysical evaluation.

The third part of the paper examined microphysical processes, by relying on a process budget from the model. Meso-NH successfully represents the reverse LWC profile (with the highest values near the ground) during the formation phase of IOP 11, likely due to the sedimentation sink term and the presence of large simulated droplets ($> 50 \mu\text{m}$ diameter). During the thin-to-thick transition phase of IOPs 6 and 11, the model captures the mixed LWC profile, but significantly overestimates it, mainly due to the previously highlighted thermodynamic biases. The model does not reproduce the vertical gradient of N_c for IOP 6, but does so for IOP 11, albeit with an overestimated droplet concentration. During the adiabatic and dissipation phases of IOP 6, the model captures the LWC profile shape but overestimates its values, which coincides with a drier layer above. However, it fails to reproduce the droplet concentration gradient. Budget analyses indicate that excessive low-level activation by turbulent motions may lead to the overproduction of droplets and the capture of water that is not used for droplet enlargement through coalescence. During the thin-to-thick transition, and the adiabatic and dissipation phases of IOP 6, the simulated DSD underestimates the droplet mode at $10 \mu\text{m}$ diameter, and a significant proportion of the largest droplets ($> 25 \mu\text{m}$ diameter). The largest droplets in the DSD at the bottom are not represented in the model. Consequently, the vertical evolution of the DSD differs between the model and the observation. This misrepresentation of DSD is partly due to the single-mode assumption, which limits the sedimentation process and contributes to the overestimation of the LWC.

Overall, the model provides a reasonable representation of both fog cases. However, some biases remain, such as the overestimation of LWC, and errors in droplet number concentration and size. While only two cases have been studied and our results are consistent with the literature – notably with regard to the modelled excess LWC – more case studies are definitely needed to confirm them. If these biases are confirmed in other cases, several approaches to improving the presented model can be considered, that have been presented in the discussion session.

The extensive SOFOG3D dataset would be valuable in achieving these goals, particularly in constraining the DSD shape more effectively using observed and radar-derived DSD moments (reflectivity and Doppler velocity). It could also provide further insight into mixing processes at the fog top. However, additional field campaigns are needed to evaluate radiative cooling at fog top, dew and droplet deposition velocity, particularly over different surface types, following the approach of Tav et al. (2018). These campaigns should also investigate the presence and impact of drizzle and its possible link to aerosol loading. Additional instrumentation, such as drizzle probes or radar Doppler spectral functionality (Kollias et al., 2011), would be highly beneficial. Finally, given the significant profile-to-profile variability observed within fog phases, robust evaluation of the mean state would require a much larger number of vertical profiles. To

fully capture phase-dependent variability, a cloud radar with enhanced functionality would provide more continuous vertical information and measure vertical velocities and TKE inside the fog layer.

Based on the present results, several priorities emerge for future fog studies:

– *Short-term model developments.*

- Implement bimodal or drizzle-capable microphysics schemes in order to better represent the observed DSD variability and evolution, and the onset of light precipitation. At the same time, revisit the assumptions about the gamma DSD parameters for fog (α , ν).
- Evaluate alternative deposition parameterizations, that may involve dependencies to other parameterizations (such as turbulence), and are known to strongly influence fog characteristics.
- Revisit CCN activation in fog, especially the link with vertical movements, turbulence and radiation.

– *Priority observations for future campaigns.*

- Measure droplets larger than $50 \mu\text{m}$ and drizzle occurrence to evaluate bimodal or drizzle-capable microphysics schemes.
- Measure fog-top radiative fluxes to assess bimodal microphysics schemes.
- Measure aerosol concentrations and properties at different locations, together with supersaturation, to further investigate fog sensitivity to aerosol loading.
- Add Doppler spectral radar functionalities to retrieve DSD, vertical velocities and TKE within the fog layer and improve understanding of fog dynamics.
- Perform more frequent vertical profiles of turbulence, DSD, and thermodynamic variables to robustly evaluate the mean state and the variability of generalized gamma DSD parameters (α , ν) and the link with other parameters.
- Measure deposition fluxes to evaluate alternative deposition parameterizations.

Code and data availability. All the data used in this study are hosted by the French National Center for Atmospheric Data and Services (AERIS) at the following links: <https://doi.org/10.25326/689> (Burnet, 2024), <https://doi.org/10.25326/89> (Canut, 2020), <https://doi.org/10.25326/134> (Delanoë, 2020) and <https://doi.org/10.25326/207> (Martinet, 2021). Data access can be provided by following the conditions fixed by the SOFOG3D project. The surface–atmosphere coupled model Meso-NH

is open source and available at <https://mesonh.cnrs.fr/>, last access: 16 July 2024; Lac et al., 2018). The generated model output data supporting the results of this study are available from the corresponding author upon reasonable request from the corresponding author.

Author contributions. MM, CL, FB, BV, TC, SA and MF prepared the manuscript and analyzed the data. TC processed in situ data. MT and CL configured the model. MM performed the simulations. MF participated to the tethered balloon operations. FB was the PI of the SOFOG3D project, designed the field campaign and led the tethered balloon operations. CL was the PI of the high-resolution modeling task and co-led the process analysis task.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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