



The LMDZ Offline Radar Instrument Simulator (LORIS): A versatile tool for the evaluation of cloud and precipitation in LMDZ

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Abstract.

The representation of clouds in Atmospheric General Circulation Models (AGCMs) remains a major challenge, with important implications for surface precipitation and the Earth radiative balance. Evaluating cloud representation in models through comparisons with remote sensing observations is essential but remains difficult. This paper presents LORIS (LMDZ Offline Radar Instrument Simulator), a new tool designed to evaluate LMDZ which is the atmospheric component of the IPSL Coupled Model. LORIS includes an advanced subgrid cloud generator, built upon COSP (CFMIP Observation Simulator Package), to represent clouds and precipitation at the subgrid scale. The radiative transfer module of LORIS (PAMTRA), enables to simulate the Doppler spectra and radar moments based on more realistic hydrometeor scattering assumptions, especially for snow particles. LORIS can be applied to a wide range of observations, including ground-based, airborne, and satellite radar measurements. Comparisons between LORIS simulations and various radar observations in polar regions demonstrate its potential for future evaluations or refinement of cloud and microphysics parameterizations included in LMDZ. Designed to be flexible and versatile, LORIS provides multiple options in its subgrid generator (e.g. cloud overlap assumptions or condensate variability), and can be adapted to different microphysics schemes. It opens new perspectives for evaluating and refining cloud and precipitation processes in LMDZ, while offering a modular framework that can readily integrate future developments of the model physics.

1 Introduction

Clouds are a fundamental component of the climate system, generating precipitation and playing a major role in the Earth energy budget. However, the formation and evolution of clouds is governed by complex microphysical and dynamical processes. These small-scale processes are particularly challenging to represent in Atmospheric General Circulation Models (AGCMs), whose coarse horizontal resolution (typically a few hundred kilometers) is too large to explicitly resolve cloud physics. Therefore, small-scale dynamical processes, such as turbulence, which plays an important role in cloud formation through turbulent mixing, need to be parameterized as subgrid processes.



In addition to the difficulty of representing macrophysical properties such as cloud fraction and occurrence, cloud micro-
physics also represents a major challenge for the parameterizations of GCMs and numerical models in general (Morrison
et al., 2020). However, an accurate representation of clouds and their microphysical properties is crucial, as microphysics can
strongly influence precipitation (e.g., via riming) (Grazioli et al., 2015; Moiseev et al., 2017; Grzegorzczuk et al., 2025b), cloud
lifecycle (e.g. Fernández-González et al., 2016; Grzegorzczuk et al., 2025a), and radiative transfer (e.g. Matus and L'Ecuyer,
2017; Korolev et al., 2017).

Evaluating the representation of clouds in GCMs is therefore essential. One approach is to use remote sensing instruments,
such as radars and lidars, which provide information on cloud structure and properties throughout the atmosphere. However,
comparisons between model outputs and remote sensing observations are not straightforward, as they require the develop-
ment of instrument simulators that project model outputs into the instrument observation space. Another approach consists in
retrieving model parameters directly from remote sensing observations (e.g. Tridon et al., 2017; Raupach and Berne, 2017;
Billault-Roux et al., 2023; Billault-Roux and Berne, 2025), although this method can be less straightforward. Following the
simulator approach, several GCMs have been evaluated through the Cloud Feedback Model Intercomparison Project (CFMIP)
using the CFMIP Observation Simulator Package (COSP) (Swales et al., 2018), which enables the simulation of several in-
strument outputs. Model–satellite comparisons, in particular, have been widely used for this purpose (e.g. Cesana and Chepfer,
2013; Konsta et al., 2015). In addition to evaluating GCMs, active remote sensing can also provide constraints on the occur-
rence of microphysical processes and guide model development (e.g. Planat et al., 2021).

Given their coarse resolution, AGCMs generally characterize clouds using subgrid-scale statistical quantities, such as cloud
fraction, resulting in representativeness differences between remote sensing observations and the simulated grid-cell mean.
Therefore, COSP among other simulators (e.g Silber et al., 2022) are using a statistical approach that consists in decomposing
each model grid cell into subcolumns, to sample the subgrid variability of clouds. The necessity to use this approach instead of
a single column approach was especially highlighted by Di Michele et al. (2012). However, decomposition into subcolumns is
a delicate step since different assumptions regarding cloud overlap or the subgrid distribution of condensate and precipitation
within the grid cell can lead to different radar signals, thereby introducing biases in comparison with radar observations (Hill-
man et al., 2018; Song et al., 2018; Wang, 2022; Hashino et al., 2025).

Among the AGCMs evaluated through this approach, LMDZ (Hourdin et al., 2020), the atmospheric component of the Insti-
tut Pierre Simon Laplace (IPSL) coupled-model participating in CMIP (O'Neill et al., 2016, Coupled Model Intercomparison
Project:), has been widely evaluated using remote sensing observations, both through COSP simulators (Madeleine et al.,
2020; Roussel et al., 2026) or directly using ground-based (Wiener et al., 2024; Roussel et al., 2023) and airborne (Raillard
et al., 2024) precipitation radars precipitation retrieval.

LMDZ has recently benefited from several advances such as the incorporation of the icosahedral dynamical core DYNAM-
ICO (Dubos et al., 2015). Building on DYNAMICO, ICOLMDZ was further developed to run as a Limited Area Model



(LAM), allowing it to be forced with reanalysis products such as ERA5 (Hersbach et al., 2020), similarly to other numerical weather prediction (NWP) models. The LAM configuration of ICOLMDZ offers the opportunity to evaluate specific aspects of the model through data collected during dedicated field campaigns, such as cloud phase from airborne in situ measurements (Raillard et al., 2024) or the representation of the katabatic wind in Antarctica using ground-based meteorological instruments (Wiener et al., 2025).

The growing use of ground-based and airborne radars to evaluate and develop LMDZ physics in LAM configuration, is less compatible with the use of the COSP radar simulator (QuickBeam, Haynes et al., 2007). Indeed, COSP was originally developed for global simulations and satellite observations (e.g. Cloudsat instruments) and provide more large-scale features. Although Riley Dellaripa et al. (2021) adapted COSP for ground-based radars, it remains limited to providing the reflectivity (no radar Doppler spectra, higher radar moments or polarimetric variables) from relatively simple scattering assumptions (Mie or Rayleigh) for the ice phase of clouds. These limitations are particularly critical for evaluating the cloud physics representation in LMDZ, especially regarding precipitation, against more advanced multi-frequency radar instruments operated from the ground or onboard aircraft deployed during field campaigns.

In numerical weather prediction (NWP) models, more advanced radar operators can use more complex scattering assumptions and provide Doppler spectra and all radar moments which are essential for evaluating cloud microphysics. This is the case for CR-SIM for the WRF model (Oue et al., 2020) or the radiative transfer model PAMTRA (Mech et al., 2020). Furthermore, some simulators (Pfeifer et al., 2008; Wolfensberger and Berne, 2018; Matsui et al., 2019) focus on polarimetric variables (e.g., Z_{DR} , K_{dp}) in addition to reflectivity. However, polarimetry, among other aspects, remains challenging for radar operators with the presence snow particles (Vignon et al., 2021). The use of a more detailed radar operators than the one used in COSP would enable more detailed evaluations of LMDZ cloud physics by allowing direct comparisons with ground-based or airborne radar observations collected during field campaigns.

This paper presents the LMDZ Offline Radar Instrument Simulator (LORIS), a new tool that consider an upgraded subgrid generator (based on COSP) with a detailed and versatile radiative transfer model (PAMTRA) for radar simulations. LORIS enables a more detailed and consistent evaluation of LMDZ cloud physics, and is built to be adaptable to future model developments.

The paper is structured as follows: Section 2 presents the cloud physics parameterizations of LMDZ, while Section 3 describes the general working of LORIS. The subgrid generator is described in Section 3.1. The radiative transfer model (PAMTRA) used to simulate radar measurements, along with the scattering assumptions and hydrometeor properties derived from LMDZ cloud physics, is presented in Section 3.2. The different sensitivity tests conducted to evaluate the subgrid generator, as well as the operator of LORIS are shown in Section 4.



2 Cloud physics in LMDZ

The representation of cloud and precipitation of LMDZ used for this paper is the one under development for CMIP7, building on the statistical cloud scheme of Madeleine et al. (2020) used for CMIP6. This scheme relies on a probability density function (PDF) representing the subgrid variability of total water or saturation deficit (for free-tropospheric clouds). Indeed, the PDF is used to compute cloud fraction as well as to determine the total condensate mass through a water saturation adjustment (water vapor in excess of saturation is condensed). On this basis, LMDZ distinguishes between large-scale, shallow convective, and deep convective clouds. Large-scale clouds are represented using a single lognormal PDF, whereas shallow convective clouds are described by a bimodal lognormal PDF. Deep convective clouds are represented using a lognormal PDF, while convection is parameterized using the Emanuel mass-flux scheme (Emanuel, 1991), coupled with a parameterization of cold pools (Grandpeix and Lafore, 2010; Grandpeix et al., 2010). Recent developments have focused on improving the representation of large-scale and shallow convective clouds, particularly for polar regions.

The first development concerns cloud phase partitioning. Indeed, in the LMDZ CMIP6 version, the amount of supercooled liquid water in clouds was diagnosed using a temperature function which has been shown to be inappropriate for polar clouds (Forbes and Ahlgrim, 2014; Raillard et al., 2024). This aspect has been revised by introducing a new phase-partitioning scheme which is presented in Raillard et al. (2026) and Vignon et al. (2026b). This scheme diagnoses the presence of supercooled liquid water based on vertical air velocity, ice microphysics properties, and turbulence, allowing the presence of ice supersaturation in supercooled liquid water.

For precipitation, rain and snow are formed through autoconversion of ice and liquid clouds, as well as from new processes introduced by Raillard (2025) via the POPRECIP scheme (Process-Oriented PRECIPitation). This scheme accounts for additional microphysical processes, such as coalescence, riming, aggregation, and sublimation. For instance, hydrometeors in this scheme are considered to be monodisperse, with their size used as a tuning parameter. POPRECIP also introduces a distinction between precipitation falling in cloudy or clear-sky conditions, which is important for sublimation and evaporation, and therefore for surface precipitation.

In addition to cloud and precipitation parameterizations, a new parameterization developed for polar applications was implemented to represent blowing snow (Vignon et al., 2026a), which plays an important role in the surface ice mass balance. A final effort has been made for cirrus clouds by Borella et al. (2025), who introduced a prognostic treatment of ice water content, ice supersaturation, and cloud fraction in LMDZ. Those high-altitude oriented developments still need to be merged with the developments regarding mixed-phase clouds described previously and are not considered in the present paper.

Therefore, the LMDZ version used here is oriented toward CMIP7, encompassing all these new developments (except the one for cirrus clouds), and has been calibrated for polar regions. An additional effort has been made to retune the physical



package of LMDZ on single column simulations of warm and cold boundary-layer clouds and stable boundary layers, following Hourdin et al. (2021); Vignon et al. (2024, 2026a). The application of LORIS on this version of LMDZ will help to refine the new parameterization of clouds and precipitation.

3 LMDZ Offline Radar Instrument Simulator (LORIS)

130 The new LMDZ Offline Radar Instrument Simulator (LORIS) consists of two main components illustrated in Fig. 1: (I) a sub-column generator, which provides a stochastic representation of the cloudy condensate in LMDZ, and (II) a radiative transfer model that produces radar outputs. As visible in Fig. 1, LORIS only considers 1D vertical profiles from LMDZ and hence corresponds to vertically pointing radars. The detailed code structure of the two compounds of LORIS is presented in Fig. 2.

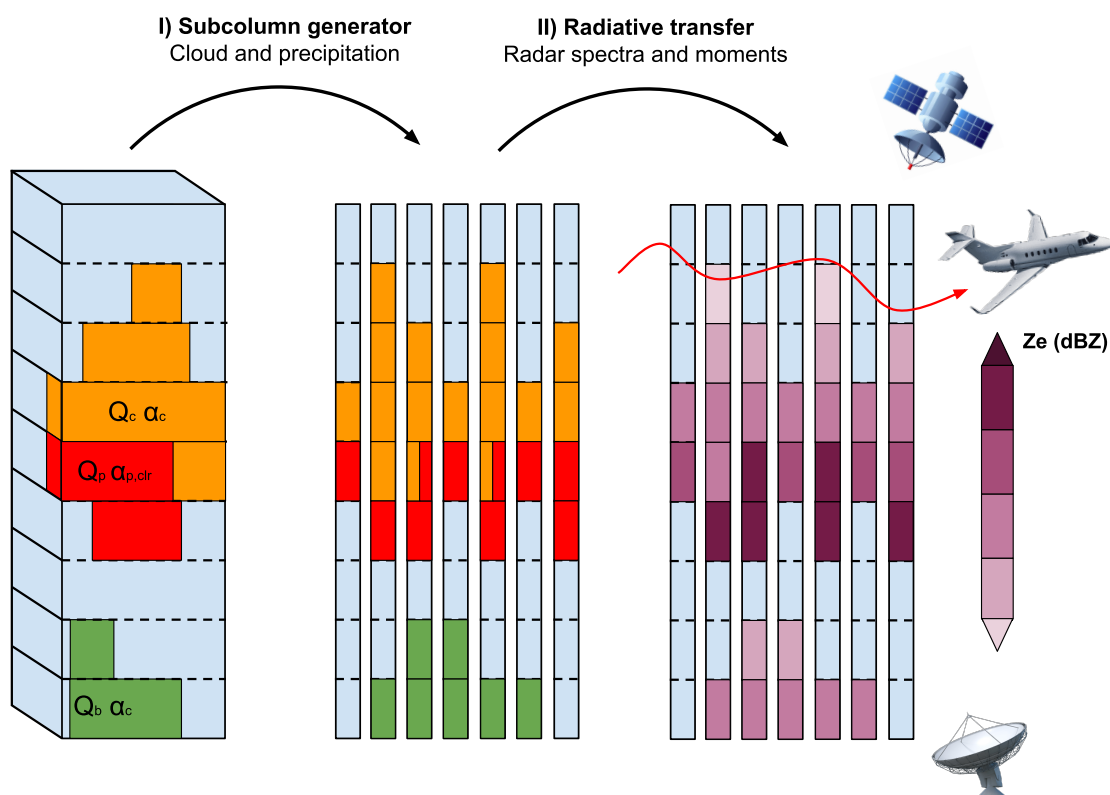


Figure 1. Schematized representation of the cloud and precipitation treatment in LORIS, with (I) the subcolumn generator, which provides subgrid profiles of cloud and precipitation condensate, and (II) the radiative transfer model, which produces radar variables from the sub-columns for ground-based, aircraft, or satellite instruments. Q_c , Q_p and Q_b are the mixing ratio of clouds, precipitation and blowing snow of LMDZ while α_c are the cloud and precipitation fractions.



LORIS is designed to run offline, which offers greater flexibility for short simulations, allowing users to adjust options, choosing different instruments, and obtain results quickly. In contrast, COSP (Swales et al., 2018) is coupled online within LMDZ, providing global diagnostics (e.g. day or monthly means) of the model that cannot be computed with LORIS, whose offline framework is more dedicated to evaluate shorter simulation in details.

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To run LORIS, some key parameters and options have to be set in its configuration file (Fig. 2a). These parameters both control the subcolumn generator options (Fig.1b) including the overlap assumption, precipitation treatment, number of sub-columns as well as the radiative transfer model settings (Fig. 1c) which are the scattering assumptions, runtime options, and radar characteristics. For each application, it is possible to define a specific configuration file, keeping the core code unchanged and ensuring reproducibility across different setups.

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The code structure of LORIS (implemented in Python and Fortran) is presented in Fig. 2, with the LMDZ output files used as input for a subcolumn generator. This subcolumn generator is based on the one implemented in COSP but includes important modifications developed by Hillman et al. (2018) (hereafter mentioned as H18). The cloud and precipitation condensate profiles generated by the subcolumn generator serve as input to the radiative transfer simulator (PAMTRA) in Fig.2c. Furthermore, to simulate radar variables in each subcolumn, the hydrometeor properties in PAMTRA are defined coherently with the physics of LMDZ. The details of those two steps are presented in the next sections.

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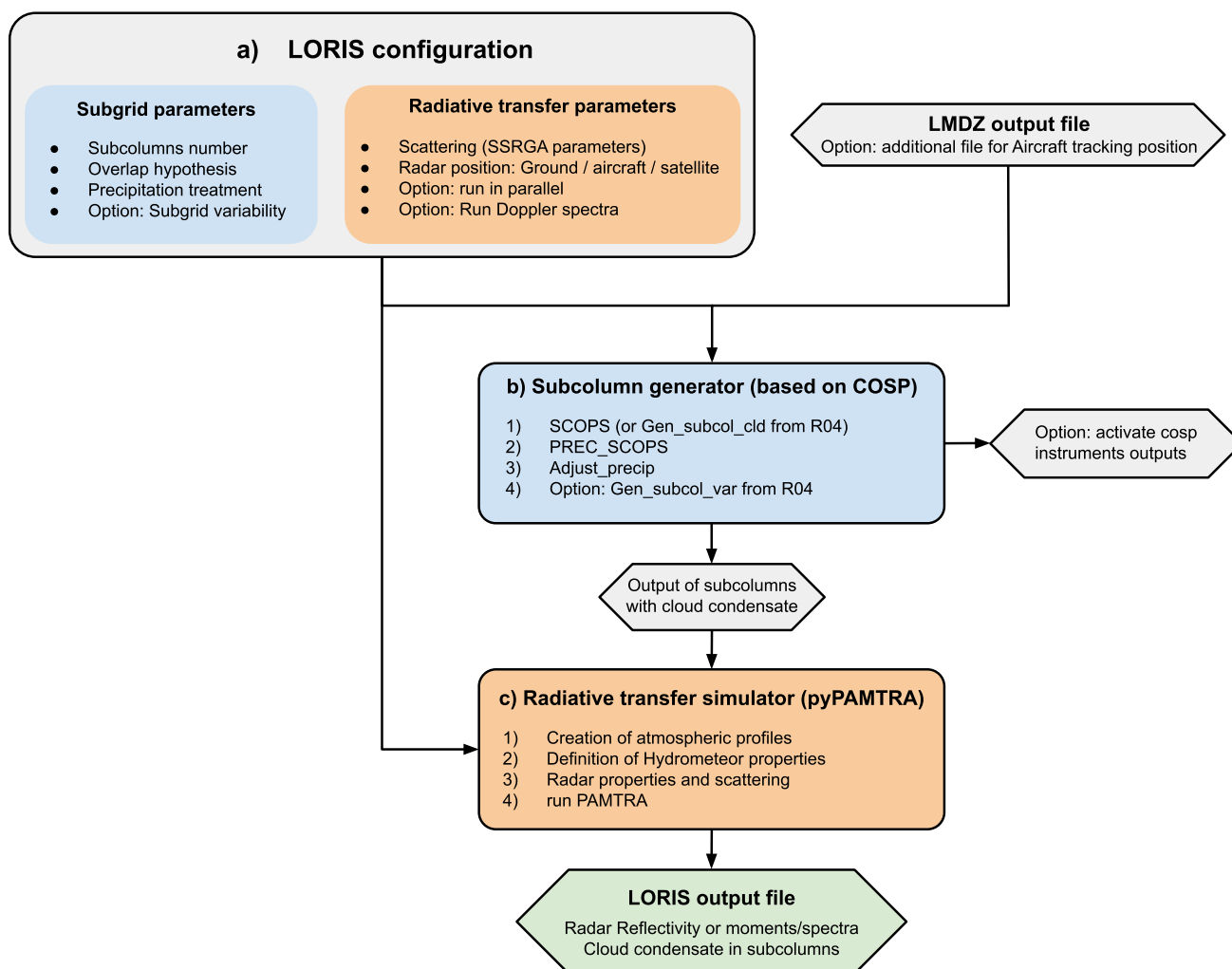


Figure 2. Schematic diagram of LORIS code structure (implemented in Python and Fortran). The boxes representing the subgrid generator are shown in light blue, whereas those corresponding to the radiative transfer simulator are shown in orange. The final output is presented in green, while the remaining options and steps are shown in gray.

3.1 Subcolumn generator

155 As illustrated in Fig.1(I), the main purpose of the subgrid generator is to subsample clouds and precipitation from the cloud fraction and mean grid-cell condensates provided by LMDZ. The following description will focus on shallow and large-scale clouds, while the specific treatment of deep convective clouds by LORIS is discussed in Section 3.3.



Considering a subgrid generator is essential to enable comparison of LMDZ (which represents cloud fraction and mean
160 cloud condensate) with remote sensing observations which sample clouds on finer spatial scales and can therefore distinguish
cloud and clear-sky.

The first step of the subgrid treatment of LORIS (Fig.2 b) is based on the Subcolumn Cloud Overlap Profile Sampler
(SCOPS: Jakob and Klein, 1999; Webb et al., 2001) implemented in COSP (Swales et al., 2018). It first assigns a cloudy or
165 clear-sky flag to each subcolumn (using the maximum-random overlap hypothesis) and then distributes the condensate mixing
ratios homogeneously (i.e. no subgrid variability) across subcolumns to match the mean gridcell values provided by LMDZ.

In a second step (Fig. 2b), precipitation is attributed to cloudy subcolumns using the algorithm `PREC_SCOSP`, which consid-
ers that all cloudy subcolumns are precipitating and that precipitation extends down to the surface, which implies a precipitation
fraction of 1 for cloudy areas. As for clouds, the precipitation mixing ratios are distributed homogeneously within precipitating
170 subcolumns.

However, several studies pointed out important biases in COSP subgrid representation of clouds and precipitation (Hillman
et al., 2018; Song et al., 2018; Wang, 2022) which are discussed below. Therefore, we adopt the improvement of COSP pro-
posed by H18 in LORIS, built upon the Räisänen et al. (2004) subgrid generator, which has been shown to provide a more
175 realistic subgrid representation than SCOPS (e.g. Oreopoulos et al., 2012).

The first modification that we consider is the treatment of precipitation. As mentioned previously, COSP assumes that all
subcolumns with cloud are precipitating down to the surface. However, it is quite common for precipitation to fall in a restricted
area below the cloud base or even evaporate. Therefore, ignoring precipitation fraction can lead to biases in the estimated inten-
180 sity and occurrence of surface precipitation. Furthermore, it can also generate an artificial dilution of precipitation condensate
across subcolumns compared to a situation with fewer precipitating subcolumns as shown in Di Michele et al. (2012). To ad-
dress this issue we adopted the `Adjust_precip` algorithm (see Fig.2b3) proposed in H18 based on the work of Di Michele
et al. (2012). The precipitation adjustment rearranges the precipitation flag in each subcolumn at every model level according
to the precipitation fraction. It also assigns the most precipitating subcolumns to the most cloudy ones. This treatment applies
185 only for the large scale precipitation of LMDZ for which the precipitation fractions have been introduced (Raillard, 2025).
Furthermore, accounting for the precipitation fraction for the subcolumns generation may become even more important as the
model resolution coarsens.

The second aspect of COSP refined by H18 and which currently considered a prospective improvement for LORIS, is the use
190 of a more realistic overlap hypothesis to determine cloud presence (or the cloud flag) among the subcolumns. Indeed, the sub-
grid sampler of COSP (SCOPS) consider the maximum-random overlap to determine the cloudy and non cloudy subcolumns.
It means that, vertically adjacent cloudy layers are maximally overlapped, while clouds separated by at least one clear-sky layer
are assumed to overlap randomly. This hypothesis is commonly used, as it is considered more realistic than purely random or



maximum overlap. However, observations from both ground-based and satellite remote sensing have shown that cloud overlap exhibits a more continuous behavior between these two extremes (Hogan and Illingworth, 2000; Barker, 2008). A more realistic subgrid representation is provided by the exponential-random (also called generalized) overlap. In this approach, an overlap parameter defines a smooth transition between maximum and random overlap. This parameter is derived from an exponential function depending on a decorrelation length scale and on the vertical separation distance between cloud layers. In addition, the condensate rank correlation parameter defined similarly as the overlap parameter can also be used to represent the probability of adjacent vertical layers with similar water contents to be aligned. These parameters can be derived from observations (Oreopoulos and Norris, 2011; Oreopoulos et al., 2012, 2022a). This method, used by Räisänen et al. (2004) and implemented in COSP by H18, is now available as an option (Fig. 2b) in LORIS in anticipation of future LMDZ developments. Indeed, the precipitation fraction given by the new precipitation scheme of LMDZ currently relies on the maximum-random overlap assumption.

The last improvement proposed by H18, also identified as a prospective development for LORIS, is the ability to account for the horizontal subgrid variability of cloud and precipitation mixing ratios across subcolumns (Fig. 2b), rather than assuming horizontally homogeneous condensate. This is achieved by prescribing a probability distribution function (e.g. a log-normal distribution) for the in-cloud mixing ratio, thereby generating variable in-cloud mixing ratios at a given model level and across the subcolumns. Accounting for the subgrid variability has been shown to significantly impact instrument simulator results (Song et al., 2018; Oreopoulos et al., 2022b; Wang, 2022). Similarly to the overlap assumption, representing such subgrid variability in LMDZ is currently under development. While recent work on cirrus clouds (Borella et al., 2025) accounts for this variability in LMDZ, its integration with the mixed-phase representation of clouds is still under development, making the subgrid variability a prospective development for LORIS to maintain consistency with future LMDZ physics.

The precipitation adjustment, along with the exponential-random overlap hypothesis and subgrid condensate variability (the two latter being prospective options) will be tested in Section 4.

3.2 Radiative transfer simulator

From the cloudy and precipitating subcolumns generated in LORIS, the radar variables are derived by the Passive and Active Microwave radiative TRAnsfer tool (PAMTRA, Mech et al., 2020) which consider plan parallel and horizontally homogeneous 1D atmospheric profiles as illustrated in Fig. 1 (II). The use of PAMTRA in LORIS is schematized in Fig. 2c and will be summarized in the next paragraphs. More details about PAMTRA can be found in Mech et al. (2020).

PAMTRA benefits from the pyPAMTRA framework introduced by Mech et al. (2020) based on its original Fortran code (Maahn et al., 2015), making it accessible and easy to use. The Python core facilitates the definition of the vertical atmospheric profiles (Fig. 2c1), where temperature, relative humidity, horizontal and vertical wind speed, eddy dissipation rate (ϵ), profiles are given by LMDZ. In addition, PAMTRA also takes in input the hydrometeor mixing ratio inside each subcolumns given by



the subgrid generator.

230 PAMTRA is able to monitor the definition of hydrometeors properties in a flexible manner (Fig. 2c). Indeed, for liquid or ice particles, monodisperse, gamma, exponential or even custom binned distribution can be set, which can correspond to a wide variety of microphysics representation. This allow to be consistent with the representation of microphysics in LMDZ (or any other models), or even to support any future developments. For instance, as mentioned in Section 2 hydrometeors in LMDZ are considered to be monodisperse particles. Therefore, for LORIS, liquid droplet sizes are supposed to correspond to
235 the effective radius used in LMDZ radiation scheme. Ice particle size are set to depend on temperature and ice water content following Sun and Rikus (1999) observations of cirrus clouds (i.e. close to $100\ \mu\text{m}$) which density is set to pure ice (i.e. $0.917\ \text{g cm}^{-3}$). The size of rain, snow and blowing snow precipitating particles are currently prescribed as tuning parameters. The snow density is set to depends on snow particle size based on Brandes et al. (2007) (i.e. around $0.1\ \text{g cm}^{-3}$). The terminal velocity of snow of LMDZ is set to $1\ \text{m s}^{-1}$, raindrops to $4\ \text{m s}^{-1}$ and blowing snow to $0.5\ \text{m s}^{-1}$. For clouds, as assumed in
240 LMDZ, droplets and ice particles are considered to be non-precipitating. From the different properties of each hydrometeor, their number concentration is derived and used to simulate observations from a radar instrument.

One important feature of PAMTRA is its ability to handle various scattering assumptions (Fig. 2c). Mie scattering is used for the liquid species (rain and droplets) as well as cloud ice since they are considered to be small (same order as the radar
245 wavelength), spherical and dense particles. The Self-Similar Rayleigh–Gans Approximation (SSRGA) method (Hogan et al., 2017) is used for snow particles.

SSRGA is particularly appropriate for snow aggregates as it accounts for their internal structure through a power law describing the backscatter cross-section. This power law depends on a set of parameters (see Table 2) that characterize the morphological properties of snow aggregates. κ represents whether the particle mass is concentrated toward its center, α , is
250 the aspect ratio, and β , γ , ζ represents the effect of small-scale structure and its fluctuations on the scattering amplitude. This method is considered, to provide realistic scattering properties of snow aggregates in comparison to the Discrete Dipole Approximation (DDA, Draine and Flatau, 1994) which is considered to be the most accurate scattering method for ice particles. However, SSRGA is often preferred, as it is computationally less expensive and is certainly sufficiently accurate to evaluate the simple microphysics of clouds used in LMDZ. More details about SSRGA can be found in Hogan and Westbrook (2014)
255 and Hogan et al. (2017).

In addition to scattering, PAMTRA (Fig. 2c) uses the radar characteristics provided in the LORIS configuration file (Fig. 2a), including the frequency, Nyquist velocity, Doppler spectrum resolution (N_{fft}), time resolution, radar noise level at 1 km (which defines the radar sensitivity) as well as the beamwidth. These radar properties can either be specified directly in the LORIS
260 configuration file (Fig. 2) or selected from a set of predefined radar types, for which all instrument characteristics are already implemented in LORIS. After defining all the important inputs, PAMTRA can be run in parallel or single core computation.



Based on the backscattering cross section defined from the scattering hypotheses and all hydrometeor properties, the radar reflectivity for each particle size can be obtained. The radar reflectivity factor Z_e can be directly computed by integrating this distribution. However, to obtain higher-order moments, the full Doppler spectra must be simulated following Kollias et al. (2011, 2014), which can be enabled in LORIS configuration (see Fig. 2). This is achieved by first using the fall velocity of each hydrometeor in the size distribution to convert the radar reflectivity per particle size to the reflectivity per terminal velocity (i.e. Doppler spectra). Secondly, the Doppler spectra is interpolated on a velocity grid that is defined from the number of fast Fourier transform points n_{fft} and the Nyquist velocity (v_{nyq}) defined in the radar characteristics (choice of the radar is an input of the configuration file of LORIS). Aliasing is also considered for terminal velocities exceeding v_{nyq} (i.e. velocities are folded back into the Doppler velocity range).

A translation of the Doppler spectra is considered due the vertical air velocity given by LMDZ. In addition, spectral broadening due to turbulence is also considered by two contributions. These contributions are the horizontal wind field contribution to the radial velocity due to the finite radar beamwidth as well as the turbulence (see Maahn et al., 2015). Both the horizontal wind speed and the eddy dissipation rate ϵ are provided by LMDZ.

Attenuation is added to the spectrum based on the extinction from gases and hydrometeors. The radar noise is added to the spectrum using the radar noise level at 1 km provided as input and is scaled at each level with range squared, thereby defining the radar sensitivity. The radar moments are derived from noise-corrected Doppler spectra, where the noise level is estimated and removed using the method described by Hildebrand and Sekhon (1974).

In the current version of LORIS, the radar signal is assumed to be nonpolarized, as the simplified microphysical assumptions in LMDZ (no particle shape representation) are not suitable for polarimetric applications. However, PAMTRA supports HV (horizontal receive, vertical transmit), VH, VV, and HH polarizations, offering the possibility to evaluate a future version of LMDZ through polarimetric variables such as ZDR and LDR among others. While LORIS is primarily focused on radar (i.e. active sensors), it can also use of the passive part of PAMTRA, providing brightness temperatures or polarized radiances of microwave radiometers across a wide range of frequencies (1 to 800 GHz).

3.3 Specific considerations for deep convective clouds

The previous sections described the treatment of large-scale clouds in LORIS. However, some specific aspects of the treatment of deep convective clouds differ need to be explained.

Regarding the subgrid generator, unlike COSP which prevents the co-occurrence of deep convective and large-scale clouds in the same subcolumn, in LORIS it is possible to have their co-occurrence in some cases. This ensures consistency with the cloud fraction constraint: $\alpha = \min(\alpha_{ls} + \alpha_{cv}, 1)$ where α and α_{cv} are the large-scale and convective cloud fractions respectively, following the approach used for radiative transfer in LMDZ. This means that when the sum of the large-scale cloud fraction α_{ls} and deep convective cloud fraction α_{cv} exceeds 1, deep convective and large-scale clouds coexist. In addition, no



precipitation fraction is provided for deep convective clouds, which prevents the use of the PREC_SCOSP algorithm.

Furthermore, no assumption for the hydrometeor size is made for clouds and precipitation since only the precipitation flux is diagnosed. As a first approximation for PAMTRA, hydrometeors from the deep convection scheme are assumed to have the same properties as those used for large-scale clouds. However, it is clear that the microphysical properties of convective clouds are different compared to more stratiform clouds (e.g. larger hydrometeors). Therefore, caution is required when interpreting the results obtained by LORIS in deep convection conditions, in absence of a clear microphysical description.

Since the current representation of deep convective cloud microphysics and precipitation can not fully benefit from the detailed subgrid and microphysical representations available in LORIS, the sensitivity analysis will focus on large-scale clouds, for which LORIS functionalities can be better evaluated. However, the development made in LORIS remains fully available for future improvements of deep convective cloud physics in LMDZ.

4 Sensitivity analysis of LORIS

To evaluate the sensitivity of the different components of LORIS on the simulated radar reflectivity, idealized profiles of cloud and precipitation fractions as well as mixing ratios are used (Fig. 3a,b). The sensitivity analysis is restricted to the ice phase (ice and snow) for large-scale clouds to simplify the interpretation of the different tests.

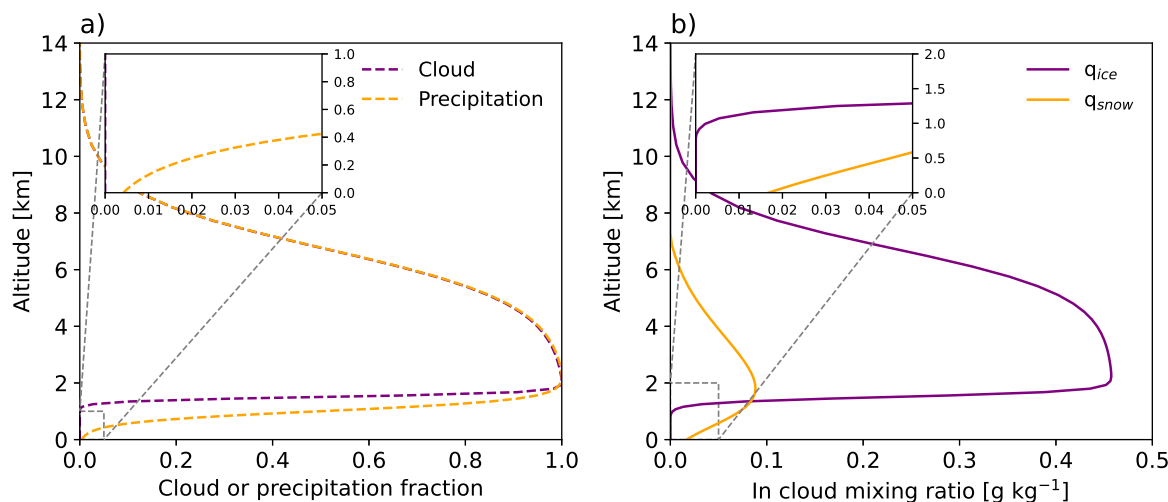


Figure 3. Vertical profiles of cloud fraction and precipitation fraction in a) as well as mixing ratio of ice and snow in b) used for the sensitivity tests of LORIS.



Both cloud and precipitation fractions (Fig. 3a) are set to increase from 0 to 1 between 12 and 2 km altitude. To represent the
315 cloud base, the cloud fraction drops to 0 near 1.5 km. The precipitation fraction follows the cloud fraction within the cloud and
gradually decreases from the cloud base to less than 1% at the ground, representing snowfall sublimation. This large variability
in cloud and precipitation fractions is designed to test the subgrid cloud generator and precipitation treatment in LORIS.

The ice mixing ratio (Fig. 3b) follows a similar evolution as the cloud fraction, reaching 0.45 g kg^{-1} near the cloud base
320 before quickly decreasing to 0 g kg^{-1} at 1.5 km. The snow mixing ratio increases from 7 km to a maximum of 0.1 g kg^{-1} at
1.5 km and then decreases to around 0.015 g kg^{-1} at the surface. Overall, these idealized profiles can be representative of a
stratiform cloud producing light snow precipitation that sublimates below the cloud base.

The remaining atmospheric profiles (temperature, humidity, and turbulence) are taken from a typical of polar contexts. These
325 variables play a minor role for the simulated reflectivity analyzed in the sensitivity tests, since reflectivity is primarily linked
to the choice of cloud and precipitation profiles (Fig. 3) and to the different settings of LORIS.

From the idealized atmospheric profiles of Fig. 3, different sensitivity tests, summarized in Table 1 are conducted. First, the
use of the precipitation treatment of H18 implemented in LORIS is tested. Second, the sensitivity to the number of subcolumns
330 is tested using 10, 30, and 100 subcolumns, which is within the range of values typically used in subgrid generators. The third
sensitivity test evaluates the effect of the scattering assumptions used in PAMTRA for snow particles by comparing Rayleigh,
Mie, and SSRGA scattering approaches. For SSRGA, different ice particle properties, and consequently sets of parameters
listed in Table 2 are tested.



Name	Precipitation treatment	Subcolumn number	Scattering	Condensate variability	Overlap
Default_N100_O1	Swales et al. (2018): default	100	O1	No	Max-random
H18_N100_O1	H18	100	O1	No	Max-random
H18_N30_O1	H18	30	O1	No	Max-random
H18_N10_O1	H18	10	O1	No	Max-random
H18_N100_O2	H18	100	O2	No	Max-random
H18_N100_H1	H18	100	H1	No	Max-random
H18_N100_H2	H18	100	H2	No	Max-random
H18_N100_Mie	H18	100	Mie-sphere	No	Max-random
H18_N100_Rayleigh	H18	100	Rayleigh	No	Max-random
H18_N100_O1_Var	H18	100	O1	Yes	Max-random
H18_N100_O1_Var_Exp	H18	100	O1	Yes	Exp-random

Table 1. Sensitivity experiments of LORIS examining the impact of the H18 precipitation treatment, different numbers of subcolumns, scattering assumptions, as well as two prospective tests including horizontal condensate variability and exponential-random cloud overlap.

The SSRGA parameters chosen in Table 2 follow those used by Billault-Roux and Berne (2025): O1 and O2 are derived from the microphysical and scattering model of Ori et al. (2021), while H1 and H2 correspond to the original values from Hogan et al. (2017) for two different aggregation models (Westbrook et al., 2004; Nowell et al., 2013, respectively).



name	aggregate type	κ	β	γ	ζ	α
O1	Mix of columns and dendrites (Ori et al., 2021)	0.22	2.52	2.36	0.049	0.6
O2	Rimed aggregate of dendrites (Ori et al., 2021)	0.15	4.98	3.53	0.036	0.6
H1	Aggregate of bullet and rosettes (Hogan et al., 2017)	0.09	0.86	2.0	0.28	0.83
H2	Aggregate of bullet and rosettes (Hogan et al., 2017)	0.16	0.15	2.33	0.22	0.82

Table 2. SSRGA scattering parameters for sensitivity tests. Same as one used in Billault-Roux and Berne (2025). O1 and O2 are derived from two different aggregation model

The results of the three sensitivity tests for the idealized profiles of Fig. 3 on W-band radar reflectivity are presented in Section 4.1.

Two additional sensitivity tests will be conducted and presented in Section 4.2 on the prospective aspect of the subgrid generator that were presented in Section 3.1. The first prospective test evaluates the effect of the subgrid condensate variability introduced in LORIS from H18. Therefore, the gamma distribution of Borella et al. (2025) used to describe the subgrid condensate distribution within cirrus clouds in LMDZ is adopted for these sensitivity tests. The parameters of this distribution are defined based on Kärcher et al. (2018) (2018) with $\nu = 4/3$ and $\lambda = \nu/q_i$ where q_i is the mean mixing ratio. The second prospective test evaluates the exponential-random overlap option implemented in LORIS following H18. Due to the prospective nature of this test, the parameters of the exponential-random overlap and condensate variability are taken directly from H18 (their Figure D1). These parameters correspond to decorrelation lengths: $L_\alpha = 1467$ m, $L_r^{cloud} = 970$ m and $L_r^{precip} = 2363$ m. The first is used to compute the overlap parameter, which defines the degree of cloud overlap in the subgrid generator, while the latter are used to compute the condensate rank correlation between two vertically adjacent layers (i.e. if similar condensate values overlap). Based on the results of Oreopoulos and Norris (2011) and Oreopoulos et al. (2012) the decorrelation lengths values used here are representative of high-latitude clouds, which is consistent with the idealized profiles defined in Fig. 3.

4.1 Effect of precipitation treatment, number of subcolumns, and scattering hypothesis in LORIS

To analyze the effect of the `Adjust_precip` algorithm, the snow mixing ratio across subcolumns is compared between *Default_N100_O1* (Fig. 4a) and *H18_N100_O1* (Fig. 4b). When using the default precipitation treatment (Fig. 4a), the snow mixing ratio is spread across all subcolumns of the model gridbox, which may overestimate the precipitation occurrence as mentioned by H18. In contrast, for the `Adjust_precip` algorithm (Fig. 4), precipitation is attributed to fewer subcolumns following the precipitation fraction. Since the mean snow mixing ratio (Fig. 3b) is preserved but concentrated into fewer subcolumns, it results in stronger mixing ratios within those subcolumns.

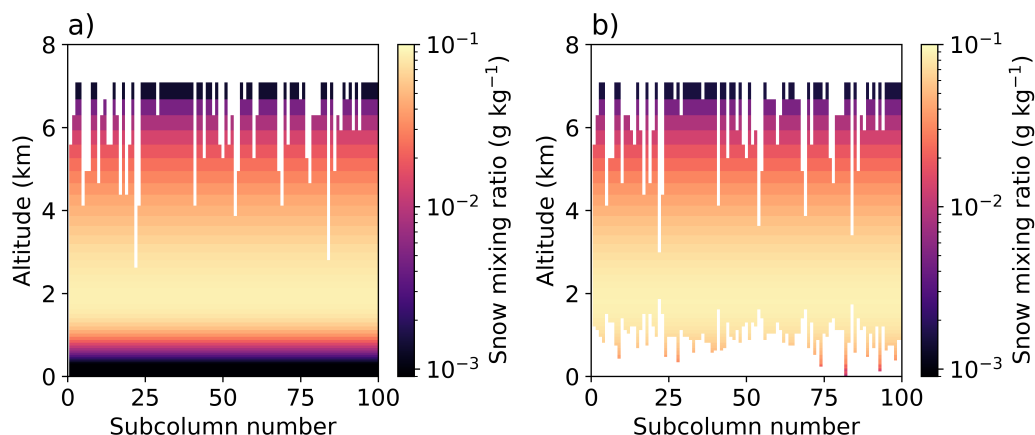


Figure 4. Subgrid snow mixing ratio across subcolumns derived from the profiles of Fig. 3, using the default precipitation treatment: *Default_N100_O1* in a) or using the *Adjust_precip* algorithm: *H18_N100_O1* in b).

The precipitation treatment has important consequences on radar reflectivity, as illustrated by the mean reflectivity profiles averaged across cloudy and precipitating subcolumns in Fig. 5a. With the dilution of the mixing ratio induced by the default precipitation treatment, Ze_W becomes progressively lower in *Default_N100_O1* compared to *H18_N100_O1*. The differences are visible from the cloud base, where the precipitation fraction starts to drop below 1 and reach up 20 dB near the surface. This is consistent with the results of Di Michele et al. (2012), who reported lower reflectivities when ignoring the precipitation fraction.

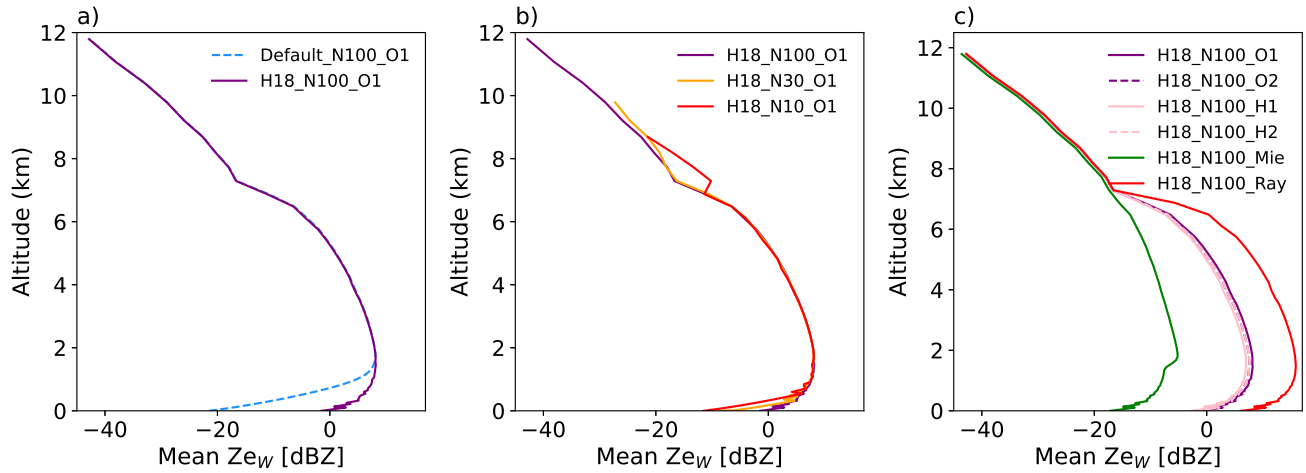


Figure 5. Mean simulated W-band radar reflectivity profiles averaged across cloudy and precipitating subcolumns for the sensitivity tests presenting in Table 1. Panel a) presents the effect of the `Adjust_precip` algorithm, b) the influence of the subcolumns number and c) the impact of the scattering assumption.

Fig. 5b presents the mean reflectivity profiles for a different subcolumn number. When the number of subcolumns is reduced (in *H18_N10_O1* or *H18_N30_O1*), the upper part of the cloud is not represented compared to *H18_N100_O1*. This is because, when the cloud fraction decreases, the number of subcolumns is not sufficient to resolve the subgrid cloud structure. For example, 100 subcolumns can represent cloud fractions down to 1%, whereas with 30 and 10 subcolumns, the minimum representable cloud fraction reaches 3.33% and 10% respectively. Consequently, this can strongly affect the representation of the upper part of clouds.

However, for all profiles in Fig. 5b, including those with a low number of subcolumns (e.g. *H18_N10_O1*), the reflectivity is simulated down to the surface, even if the precipitation fraction drops to around 1% near the ground (see Fig. 3a). Indeed, `Adjust_precip` assigns at least one precipitating subcolumn when the precipitation fraction is non-zero. With only 10 subcolumns (*H18_N10_O1*), this minimum of one precipitating subcolumn represents a minimum precipitation fraction of 10%, regardless the prescribed values that can be much lower (e.g. 1% at ground in our case). Therefore, the snow condensate is spread over a larger fraction which leads to its dilution. While this effect is similar to that observed for the default precipitation treatment in Fig. 5a, it remains smaller, with a maximum difference of around 10 dB between *H18_N10_O1* and *H18_N100_O1*.

To quantify the benefit of increasing the number of subcolumns, Fig. 6 presents the mean absolute error in Ze_W and its quartiles for configurations ranging from 5 to 100 subcolumns (same settings as *H18_N100_O1*) relative to a reference LORIS simulation using 10000 subcolumns. As expected, increasing the number of subcolumns reduces the mean error, although non linearly. Indeed, most of the improvement is achieved at around 20 to 30 subcolumns since the error diminution becomes more



limited beyond that. Furthermore, the third quartile (Q_3) approaches the mean beyond 20 subcolumns, indicating that extreme errors become less frequent.

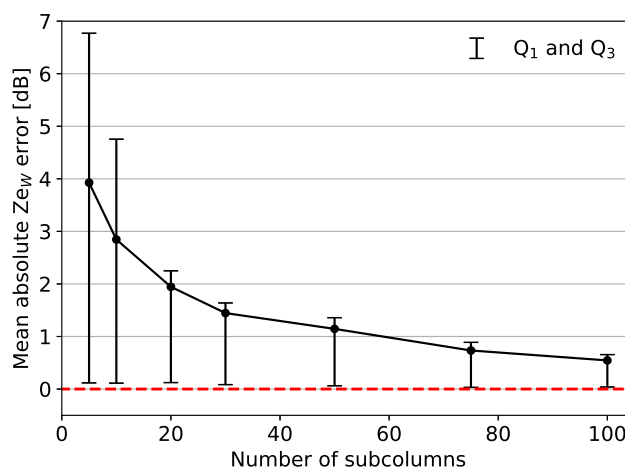


Figure 6. Absolute Mean error of Ze_W with 1st and 3rd Quartiles (Q_1 and Q_3) depending on the number of subcolumns. The same configuration as *H18_N100_O1* (Table 1) is used, but varying the number of subcolumns from 5 to 100. Errors are computed relative to a reference simulation with 10000 subcolumns.

385 While the results of Fig. 6 could depend on the profiles defined in Fig.3, Di Michele et al. (2012) also concluded that the difference induced by the number of subcolumn on the radar reflectivity starts to be less significant with 20 subcolumns or more. Therefore, it would be preferable not to use fewer than 20–30 subcolumns in LORIS. Furthermore, radars are generally considered to be calibrated within ± 1 dB, meaning that 50–60 subcolumns should be necessary to be within this range. The number of subcolumns should also be adapted to the case study, especially when cloud or precipitation fractions are low. This
390 situation could be more frequent for large model grid cells, which may therefore require a larger number of subcolumns.

The sensitivity of the mean radar reflectivity profile to the scattering assumption is shown in Fig. 5c. At upper levels (above 7 km), where only ice is present, all profiles are similar. Indeed, since monodisperse ice particles in LMDZ are around 50 to 100 μm in diameter, much smaller than the radar wavelength, this results in close agreement between Mie and Rayleigh
395 scattering. However, in the presence of snow (below 7 km), strong discrepancies arise between the scattering assumptions: Rayleigh produce the highest reflectivities, whereas Mie yields to the lowest, with differences reaching up to 24 dB. Using the SSRGA hypothesis in Fig. 5c, which is more realistic for snow particles (Ori et al., 2021), leads to reflectivities between these two extremes which is consistent with the results of Hogan and Westbrook (2014) (their Figure 5). This highlights the importance of the choice of scattering assumption for snow particles, particularly since GCM radar operators (e.g. QuickBeam
400 Haynes et al., 2007), primarily developed for satellite applications, rely on Mie or Rayleigh. Applying such assumptions to snow precipitation cases observed by ground-based radars may therefore introduce significant biases in simulated reflectivity.



It seems therefore necessary to use more sophisticated approaches such as SSRGA, currently available in LORIS.

The differences arising from the use of different SSRGA parameters, reflecting different assumptions about snow properties (Table 2), lead to maximum differences of around 2 dB in Fig. 6c. However, larger differences can be found for rimed snow particles (e.g. see Tridon et al., 2019; Maherndl et al., 2023) or for different particle size distributions. It should be noted that at lower frequencies (e.g. K-band), the reflectivity differences between the scattering hypothesis presented in Fig. 6c are smaller.

4.2 Prospective tests on subgrid treatment

To analyze the first prospective test presented in Section 4 and Table 1, Fig. 7a,b illustrate the ice and snow mixing ratios, influenced by the subgrid condensate variability. As shown in Fig. 7, using the subgrid variability leads to maximum mixing ratios of approximately 1 g kg^{-1} for ice and 0.15 g kg^{-1} for snow as well as low values close to 0.01 g kg^{-1} . Without subgrid horizontal variability, the mixing ratios across the subcolumns corresponds to the mean vertical profiles in Fig. 3.

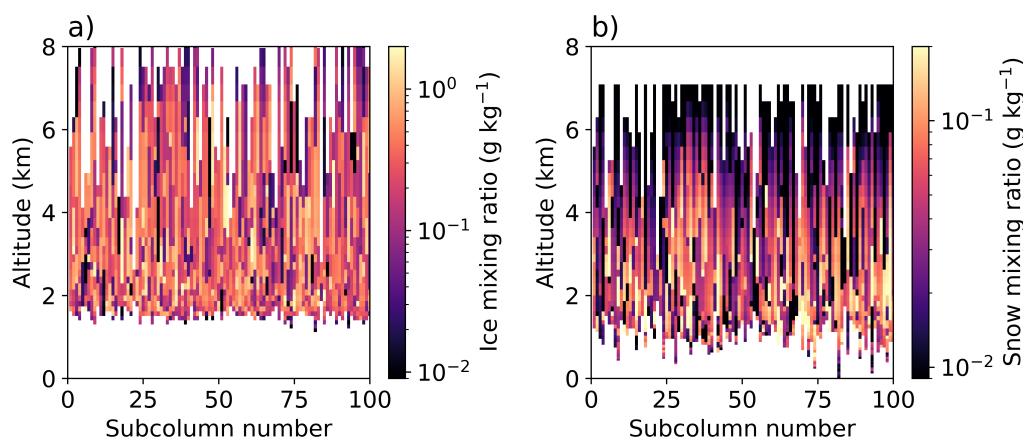


Figure 7. Subgrid mixing ratio of ice in a) and snow in b), for *H18_N100_O1_Var* that consider the condensate subgrid variability based on the gamma probability distribution function used in Borella et al. (2025).

The effect of subgrid variability on the simulated mean W-band radar reflectivity is presented in Fig. 8a. Considering the subgrid variability in *H18_N100_O1_Var* leads to decrease the reflectivity by about 2–3 dB compared to *H18_N100_O1*, for both snow near the ground and ice close to the cloud top. This overestimation of reflectivity in the absence of subgrid variability has also been identified by Song et al. (2018). Sensitivity tests on subgrid condensate distributions can be found in Wang (2022), but are beyond the scope of this study.

It important to note that the mixing ratio variability of ice and snow condensate is not correlated, as only the most precipitating subcolumns are linked to the cloudiest subcolumns (in terms of level numbers). Introducing a correlation between these variables could be a potential future improvement.

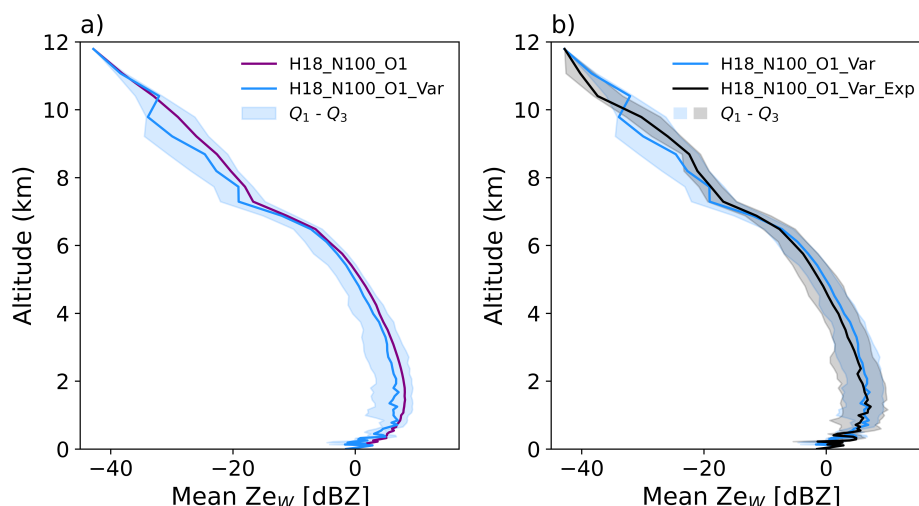


Figure 8. Simulated W-band radar reflectivity profiles averaged across cloudy and precipitating subcolumns for the prospective sensitivity tests presenting in Table 1. Panel a) presents the effect influence of the subgrid variability and b) the influence of the overlap hypothesis. The shaded areas indicate the 1st and 3rd quartiles.

The second prospective test evaluates the use of the exponential-random overlap hypothesis as an alternative to the maximum-random assumption. In Fig. 8b, the differences between *H18_N100_O1_Var_Exp* and *H18_N100_O1_Var* are quite small without any clear tendency toward an increasing or decreasing reflectivity. However, the differences are more pronounced near cloud top, where the cloud fraction decreases. While Fig. 8b suggests that the overlap hypothesis has a minor effect on simulated reflectivity, the present analysis is restricted to a single profile. Therefore, the overlap hypothesis could play a more important role when applied to global model outputs for satellite applications, as shown in H18.

5 Application examples

As presented in Section 3, LORIS can simulate radar observations from ground-based, airborne, and satellite radars. To illustrate this capability, we applied LORIS to LMDZ outputs to simulate radar observations from the three platforms, with the configuration adapted to the characteristics of each radar type. For all cases, 30 subcolumns are used, which is a good compromise between computational cost and subgrid representation accuracy, as suggested by Fig. 6. In this application, LORIS excludes the prospective options (subgrid variability and exponential random overlap), which are not yet compatible with the current version of LMDZ physics.



5.1 Precipitation event in Antarctica

435 To show the use of LORIS for ground-based and satellite radar, LMDZ was run to simulate a precipitation event that occurred in Antarctica between 15 and 18 February 2025 using a ICOLMDZ - LAM configuration. The simulation was performed over one week, from 12 to 19 February, with a horizontal resolution of 20 km, a time step of 30 s with the model physics called every 15 minutes. The domain, shown in Fig. 9, covers a radius of approximately 1300 km and is centered on the Antarctic coast to capture the entire event. This event was triggered by the arrival of an extratropical cyclone on the Adélie Land coast.

440 The advected warm and moist air, combined with terrain-induced uplift, led to a moisture intrusion that produced precipitation over the Antarctic continent, reaching the inner Antarctic plateau. During this event, a Mira-35C radar (Ka-band and manufactured by METEK) was operational at D17 (66°43'S 139°42'E), approximately 15 km from the French station Dumont d'Urville (located on the coast). The deployment of this instrument, along with other remote sensing instruments, took place in the context of the AWACA project (<https://awaca.ipsl.fr/>), which aims to improve the understanding of the atmospheric water

445 cycle in Antarctica. LORIS is applied to the LMDZ outputs at D17 to simulate the ground-based Mira-35C radar.

In addition, as shown in Fig. 9, the EarthCARE satellite passed over the region covered by the precipitation event, allowing measurements from its Cloud Profiling Radar (CPR) operating in W-band (Kollias et al., 2018). Since the EarthCARE track is located within the ICOLMDZ domain, LORIS will be also applied to simulate the CPR instrument.

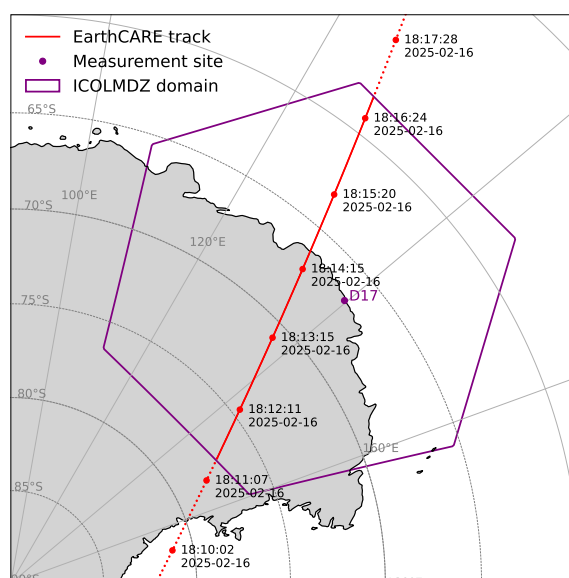


Figure 9. ICOLMDZ domain used to simulate the precipitation event of the 15-18 February 2025 over Antarctica, with the measurement site of the Mira-35C radar at D17 (66°43'S 139°42'E) as well as the EarthCARE satellite track.



450 5.1.1 Ground-based radar

Fig. 10 shows the observed radar moments of the Mira-35C radar at D17 (left column) and the LORIS-simulated moments (right column). As each model time step contains 30 subcolumns, individual subcolumns are randomly displayed at intermediate time steps, providing a full visualization of the subgrid variability of LORIS.

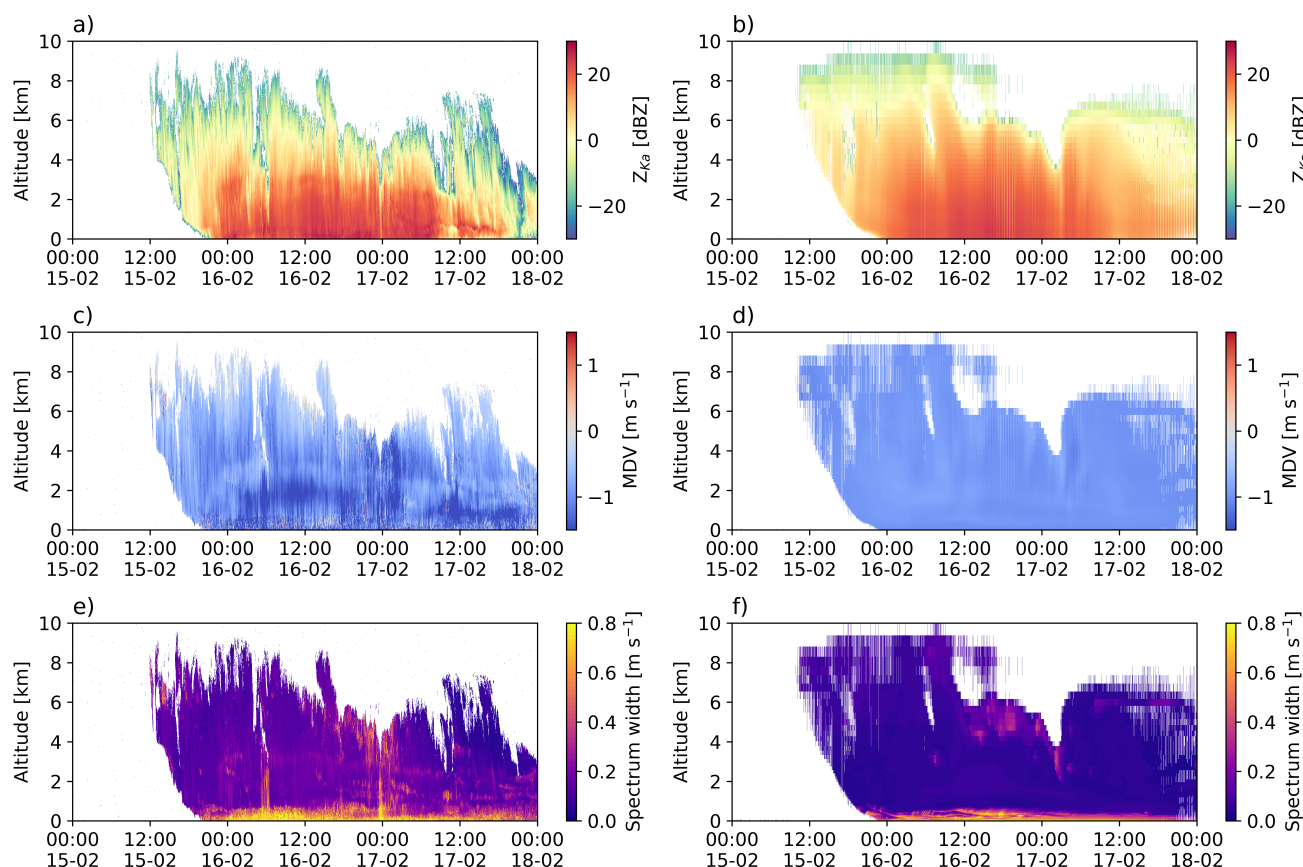


Figure 10. Radar moments from the Mira-35C Ka-band ground based radar deployed at D17 in Antarctica (left column) and LMDZ outputs processed with LORIS (right column) for the precipitation event of 15-18 February 2025: a) and b) Ka-band reflectivity, c) and d) Mean Doppler Velocity (MDV) as well as e) and f) Doppler spectrum width. For LMDZ simulations, all subcolumns are displayed and treated as intermediate time steps for visualization.

The observed global reflectivity time-height pattern in Fig. 10a is well reproduced by the simulation in Fig. 10b. While the timeline of the event seems to be well reproduced by LMDZ, the simulated reflectivity tends to be stronger near the cloud top. Since reflectivity depends on the number and size of particles, such biases can indicate that the hydrometeor properties (e.g. mass, size or concentration) are not well represented in the model. This aspect is particularly important since the hydrometeor properties both result and affect microphysical processes such as riming and aggregation that lead to the formation of



precipitation. The results of Fig. 10 illustrates the potential of LORIS to both evaluate and identify the possible improvement
460 in LMDZ physics. Furthermore, since LORIS can be applied to different radars, combining several radar frequencies could
further help evaluate LMDZ through the Dual Frequency Ratio (DFR) which depends on the hydrometeor size and not on their
concentration.

A second radar moment, the Mean Doppler Velocity (MDV) is shown in Fig. 10c and d for observations and LORIS respec-
465 tively. The MDV is close to -1 m s^{-1} in LMDZ while the observations show more variability, with values reaching -1.5 m s^{-1} .
Since the MDV depends on both the vertical air motion and the hydrometeor terminal velocity, it can provide valuable infor-
mation about the hydrometeor sedimentation in models. Furthermore, the MDV can also reflect the presence of microphysical
processes (e.g. riming) that can influence the terminal velocity of ice particles, even if this dependency is not yet represented
in LMDZ physics.

470 Finally, the Doppler spectrum width is shown in Fig. 10e and f for observations and simulation respectively. The spectrum
width reflects the variability of the Doppler spectrum, which depends on both air motion (horizontal wind and turbulence)
and the variability in hydrometeor size. Both the observations and simulation show significant spectrum width values near the
ground (more than 0.8 m s^{-1}) indicating the presence of turbulence in the Antarctic katabatic layer. The spectrum width
475 simulated by LORIS may therefore help evaluate the representation of this layer, for example in terms of its vertical extent. In
Fig. 10e, the observations also show a secondary peak in spectrum width between 2 and 3 km altitude, which is not reproduced
by LMDZ in Fig. 10f. Such discrepancies can help to assess both microphysical and dynamical processes representations in
LMDZ, as they may result from microphysical effects on hydrometeor size distributions (e.g. aggregation) or from turbulence.
However, such dependence on particle size is not yet represented in LMDZ.

480 Other radar moments such as skewness or kurtosis could also be informative, as they reflect the shape of the Doppler
spectrum. However, since the hydrometeors in LMDZ are currently monodispersed, these variables are of limited interest to
evaluate the present version of LMDZ cloud physics.

5.1.2 Satellite

485 For the same precipitation event, LORIS is applied to LMDZ outputs corresponding to the EarthCARE satellite trajectory (see
Fig. 8). Fig. 11a,b show the W-band radar reflectivity (Level 1b product) measured by the CPR instrument onboard EarthCARE
and the simulated reflectivity of LMDZ respectively.

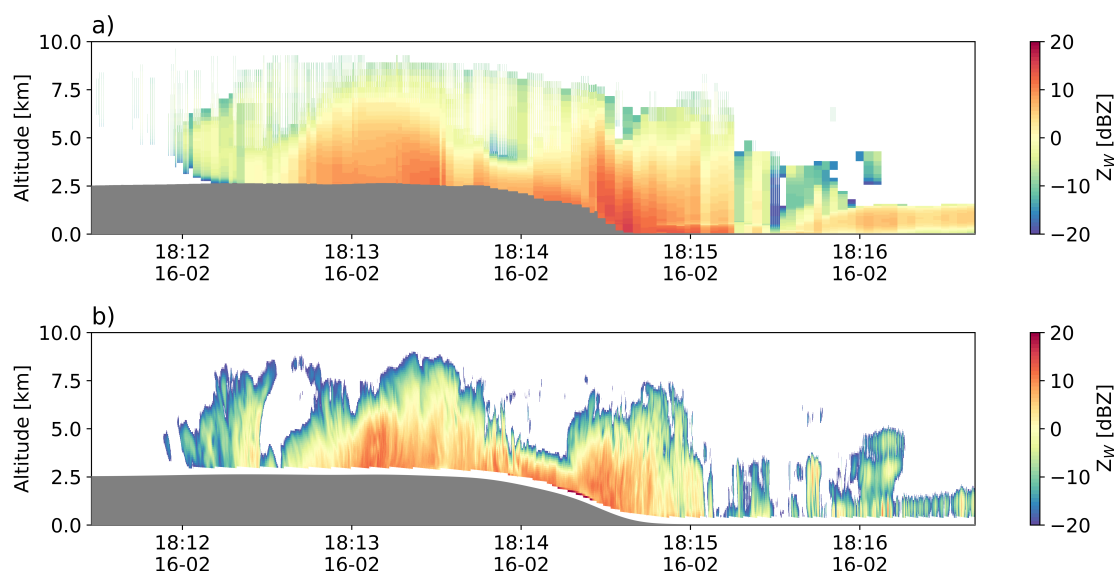


Figure 11. W-band reflectivity measured by the cloud profiling radar (CPR) of EarthCARE satellite in a), with the corresponding reflectivity simulated by LORIS from LMDZ outputs in b), for a precipitation event occurring between 15 and 18 February 2025 in Antarctica. For LMDZ simulations, all subcolumns are displayed and treated as intermediate time steps for visualization. Ground echoes are filtered out from Earthcare data.

The comparison of W-band reflectivity in Fig. 11 shows overall consistency between the simulated and observed values, except for some locally higher values in LMDZ. However, the simulated cloud system extends to higher altitudes (up to 8 km) along the EarthCARE track, while the observations exhibit a shallower cloud structure. As for the ground based Mira-35C radar, EarthCARE CPR also provides additional radar moments that could be compared with LORIS outputs to further evaluate the representation of cloud physics and dynamics in LMDZ, although this is beyond the scope of the present illustration. Overall using LORIS in a satellite configuration can inform us about larger-scale view of cloud coverage and horizontal cloud structure than ground-based observations based on more reliable simulated reflectivity than that given by COSP. The satellite approach is particularly valuable in remote areas where it is difficult to deploy ground-based instruments.

5.2 Airborne radar for an Arctic mixed-phase cloud case

The last illustration of LORIS, for airborne radar is based on a case study from the RALI-THINICE campaign (Rivière et al., 2024). During this campaign, measurements were conducted aboard the ATR-42 research aircraft equipped with microphysics probes and remote sensing instruments including the 95 GHz Doppler cloud radar RASTA (RAdar SysTem Airborne, Delanoë et al., 2013), which sampled clouds in both upward and downward directions. The presented case corresponds to an Arctic cyclonic event that occurred on 6 August 2022, close to the Svalbard archipelago (flight 40), with further details about this case that can be found in Raillard et al. (2024). For this case, we use the LMDZ simulation of Raillard et al. (2024), as an input



to LORIS for the comparison with the dual vertical pointing RASTA radar. Since two pointing directions are present for the RASTA, a special option was implemented in LORIS to handle dual vertical pointing radars, which consists in performing two
505 distinct runs that are combined into a single output.

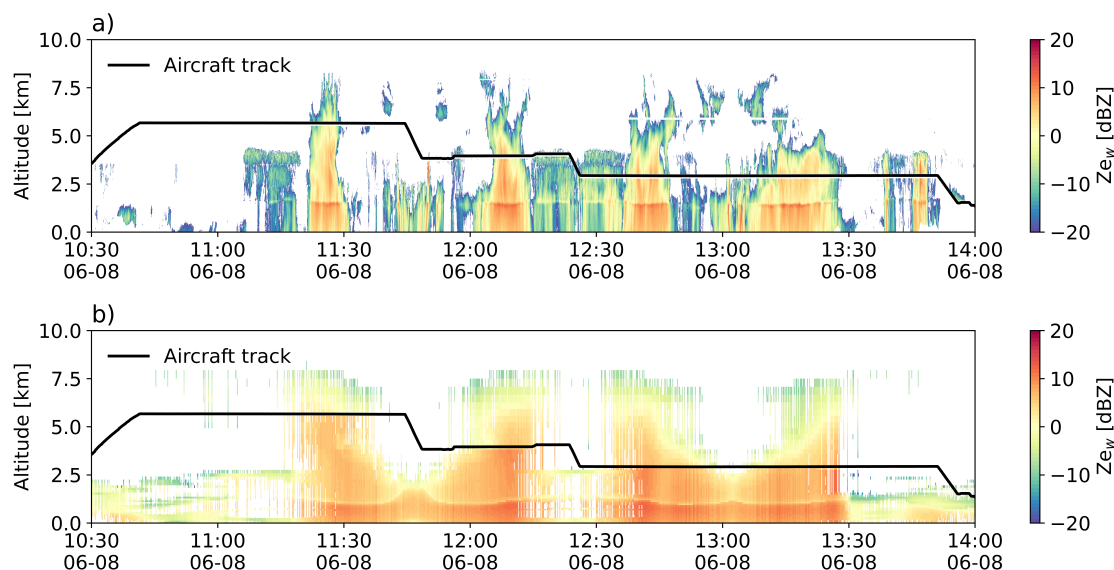


Figure 12. Upward and downward W-band radar reflectivity measured by the RASTA in a), with the corresponding reflectivity simulated by LORIS based on LMDZ outputs in b). This case corresponds to flight 40 of the RALI-THINICE campaign (Rivière et al., 2024), for which we use the LMDZ outputs of Raillard et al. (2024). For LMDZ simulations, all subcolumns are displayed and treated as intermediate time steps for visualization.

Fig. 12 presents the radar reflectivity obtained from the two RASTA radars in panel a) as well as from LORIS based on LMDZ output in b). The reflectivity of LMDZ is generally higher than in the observations, especially above the melting layer for the ice phase (i.e. above 2 km altitude). Regarding the general structure of the cloud system, the melting layer seems to be
510 located around 400 m lower than in the observations, which may reflect biases in the representation of the melting process. LMDZ also predicts the presence of shallow cloud systems with a depth of about 2 km before 11:15 UTC and after 13:30 UTC while these cloud systems are not observed by the W-band radar. In addition, the nimbostratus towers of LMDZ seem to be larger than in the observations. This example illustrates the utility of LORIS for evaluating the small-scale structure of clouds in LMDZ for specific cases, with the aircraft configuration which benefits from the successive measurements collected along
515 the aircraft path.



6 Conclusions

This paper presents LORIS (LMDZ Offline Radar Instrument Simulator), a new tool build to evaluate LMDZ, the atmospheric component of the IPSL (Institut Pierre Simon Laplace) GCM against detailed radar observations. LORIS was designed to evaluate LMDZ in a more detailed way than COSP, especially for regional case studies similarly as for NWP models. LORIS was therefore developed to bridge the gap between advanced NWP radar operators and global instrument simulators used in GCMs such as COSP which accounts for the subgrid representation of clouds and precipitation. Therefore, LORIS combines a subgrid generator with the versatile radiative transfer model PAMTRA (Mech et al., 2020), enabling the simulation of Doppler spectra and all radar moments with more realistic scattering assumptions. A key aspect of LORIS is its versatility, as it can be applied to ground-based, airborne, and satellite radars, making it adaptable to a wide range of evaluation methods as well as to upcoming developments in LMDZ cloud physics.

The subgrid generator of LORIS is build upon the one of COSP, incorporating in addition the improvements proposed by Hillman et al. (2018). The main improvement is the use of the precipitation fraction provided by LMDZ to generate more realistic precipitating and precipitation-free levels among the subcolumns, in contrast to COSP which assumes that all cloudy subcolumns are precipitating down to the surface. Two additional prospective improvements from Hillman et al. (2018), based on Räisänen et al. (2004), are also considered. First, the cloudy subcolumns of LORIS can be generated using the exponential-random cloud overlap hypothesis, which is more realistic than the maximum-random assumption. Second, it is possible to consider subgrid horizontal heterogeneity for cloud and precipitation mixing ratios based on probability distribution functions, rather than assuming a homogeneous in cloud mixing ratio across subcolumns. These two prospective features are available in LORIS in anticipation of the ongoing developments in the cloud physics of LMDZ.

The subcolumns generated by the subgrid generator serve as input to the Passive and Active Microwave radiative TRANSfer tool (PAMTRA, Mech et al., 2020) included in LORIS. The main advantage of PAMTRA lies in its versatility, since it combines a Python framework around a Fortran core that facilitates both the definition of hydrometeor properties and the specification of LMDZ input profiles (temperature, humidity, turbulence), making it easily adaptable to future developments of LMDZ. In addition, PAMTRA include different scattering assumptions, such as the Self-Similar Rayleigh-Gans Approximation (SSRGA, Hogan et al., 2017) used for snow particles in LMDZ. The SSRGA provides more realistic scattering properties compared to pure Mie or Rayleigh scattering, and is adaptable to different types of ice particles properties.

Sensitivity tests are conducted from idealized cloud and precipitation fractions as well as mixing ratio profiles to evaluate the influence of the different options available in the subgrid generator and radiative transfer model of LORIS. Accounting for the precipitation fraction in the subgrid generator leads to a more concentrated mixing ratio as the precipitation fraction decreases, resulting in higher reflectivities of up to 20 dB for the mean W-band reflectivity among precipitating subcolumns. We also find that the number of subcolumns plays a major role, especially when the precipitation fraction is low. In this case, too few



subcolumns can lead to a misrepresentation of cloud top height and to a dilution of the precipitation mixing ratio, resulting in an underestimation of the mean reflectivity by up to 10 dB near the ground. We also find that the benefits in terms of error reduction from increasing the number of subcolumns start to diminish beyond around 20 to 30 subcolumns, indicating a balance between computational cost and accuracy. Regarding the prospective aspects of the subgrid generator, introducing horizontal subgrid variability leads to a constant decrease in reflectivity of around 2-3 dB, while using the exponential-random overlap hypothesis instead of the maximum-random assumption shows no clear tendency, with only minor fluctuations. Finally, the sensitivity to the scattering assumptions of snow particles were tested. Using Mie or Rayleigh scattering leads to differences of up to 24 dB in reflectivity while SSRGA yields to values between these two extremes and can be considered to be more realistic. Using different SSRGA snow particle properties leads to smaller differences of around 2 dB.

Three application examples of LORIS were presented for ground-based, satellite, and airborne radar configurations and compared to observations. Radar reflectivity and other radar moments provide valuable insights into the hydrometeor representation and the microphysical as well as dynamical processes implemented in LMDZ cloud physics. Since these processes and properties are directly linked to precipitation, whose representation remains a major challenge in GCMs (Bony et al., 2015; Hellmuth et al., 2025), LORIS opens new perspectives for the evaluation and refinement of cloud and precipitation in LMDZ, while remaining adaptable to future developments of the model physics.

Data availability. The LMDZ Offline Radar Instrument Simulator (LORIS) is available at <https://doi.org/10.5281/zenodo.21475247> (Grzegorzczuk, 2026a). The ground-based MIRA-35C radar observations are available at <https://zenodo.org/records/21495051> (Grzegorzczuk, 2026b). The radar observations from the RALI-THINICE campaign used in this study are available in <https://doi.org/10.25326/472> (Delanoë, 2023). EarthCARE CPR data are available through the ESA EarthCARE data portal <https://doi.org/10.57780/eca-b3bc031> (European Space Agency, 2025) and the JAXA G-Portal <https://doi.org/10.57746/EO.01JDVCYDWJF63VPDBJP0VZ6V64> (Japan Aerospace Exploration Agency (JAXA), 2024).

Competing interests. The authors declare no competing interest.

Author contributions. PG: wrote original draft, LORIS development, ICOLMDZ simulations, data analysis. LR: RALI-THINICE simulations, manuscript editing. VW: ICOLMDZ Antarctic simulations, manuscript editing. EV: ICOLMDZ simulation, supervision, manuscript editing. CG: supervision, manuscript editing. AB: supervision, radar expertise for LORIS development, manuscript editing.

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