



Advancing the microphysical and subgrid treatment of precipitation in the LMDZ AGCM: parametrization development and first applications in polar contexts

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Abstract. The current assessment of the ice sheet surface mass balance and more generally of the atmospheric branch of the high latitude water cycle mostly relies on climate model simulations. The ability of climate models to reproduce polar precipitation depends not only on the simulation of the atmospheric dynamics and on the advection of moisture towards the poles but also on the representation of the subgrid-scale cloud and precipitation processes that govern the formation and growth of snowflakes and raindrops. The LMDZ atmospheric general circulation model, atmospheric component of the IPSL-CM climate model, is intensively involved in polar-oriented studies and recent developments were carried out to improve the representation of mixed-phase and ice clouds. However, recent studies also highlighted substantial shortcomings and biases that persist in the simulation of polar precipitation, both in the Arctic and in Antarctic. This study presents the development of a new precipitation scheme for LMDZ model that includes both an advanced microphysical treatment of snowfall and subgrid considerations to properly account for the interactions between hydrometeors and clouds. The scheme has been designed with a moderate-complexity approach, retaining strong approximations such as stationarity while avoiding additional prognostic variables that would require transport, thereby limiting its computational cost and making it suitable for global climate simulations and easy to tune. Particular attention has been paid to the numerical treatment of the different processes to ensure numerical convergence and stability at typical time steps used in global climate models. The scheme is then evaluated using regional simulations conducted over the Svalbard Archipelago and over Adélie Land, East Antarctica. The cloud water contents are assessed using airborne radar and lidar data collected during the RALI-THINICE campaign and the simulated vertical profiles of precipitation and microphysical tendencies are compared with observational data from a ground-based polarimetric radar deployed during the APRES3 Antarctic campaign. Perturbed parameter ensemble experiments are also conducted to assess the parameteric sensitivity of the model and to disentangle calibration issues from genuine structural biases. Results show that the model is now able to physically capture the vertical evolution of snowfall and to simulate more realistically the melting layer in the Arctic



case study. Future applications to simulate more reliably the present and future Antarctic surface mass balance with LMDZ as well as avenues for improvement particularly regarding some microphysical approximations can now be envisaged.

1 Introduction

In polar regions, over 90% of precipitation occurs as snow (Field and Heymsfield, 2015), playing a central role for both the surface and the atmosphere. Snowfall is the primary contributor to the surface mass balance of ice sheets, with direct consequences for global sea level. Moreover, when fresh snow accumulates on sea ice or continental ice, it modulates surface albedo and thermal insulation, thereby influencing the timing and magnitude of melt. In both the Arctic and Antarctic, snowfall also influences the phase partitioning and lifetime of mixed phase clouds (MPCs, McIlhattan et al., 2017). Precipitation production is particularly efficient in presence of supercooled liquid water that favour the Wegener-Bergeron-Findeisen, secondary-ice production, and riming processes (Silber et al., 2021). An accurate representation of the amount of precipitation, but also of its phase (Bintanja and Selten, 2014; Vignon et al., 2021), is paramount to simulating correctly both snow and ice-covered surface conditions and clouds at the poles.

Despite its central role, polar precipitation remains a significant source of uncertainty in atmospheric models and atmospheric reanalyses (Palermé et al., 2017b, a; Barrett et al., 2020). Reanalyses such as ERA5 and climate model ensembles from the Coupled Model Intercomparison Projects (CMIP) CMIP5 and CMIP6 systematically overestimate snowfall over Antarctica, particularly on the high interior plateau and coastal slopes (Roussel et al., 2020). This overestimation persists across resolutions and model generations, with little improvement from CMIP5 to CMIP6 (Roussel et al., 2020). While the seasonal cycle of precipitation is generally well captured at the continental scale, spatial patterns and magnitudes remain misrepresented. The model intercomparison carried out by Roussel et al. (2023) over the Dumont d'Urville station in coastal Adélie Land, East Antarctica, also shows a systematic overestimation of surface precipitation at this location and reveals difficulties in accurately capturing seasonal variability, orographic effects, and microphysical processes. These findings highlight that precipitation parameterizations require improvement and that increasing model resolution by itself is not sufficient to diminish biases.

Precipitation schemes in atmospheric models can follow either a prognostic or diagnostic formulation. In prognostic (non-stationary) schemes, hydrometeor species are explicitly evolving with time and are advected by the resolved dynamics, allowing for a more detailed representation of precipitation formation and transport (Gettelman et al., 2015). However, these schemes require the prognostic treatment and advection of additional water species, a careful coupling with cloud macrophysics, and often sub-timestepping to handle sedimentation processes (Chosson et al., 2014) which increase significantly their numerical costs and physical complexity. Moreover, this added complexity does not systematically translate into better physical realism or results. Van Dalum et al. (2024) show that snow advection can add more complexity and side effects than real improvement regarding the simulation of surface precipitation, and lead to unrealistically high precipitation extremes over regions with important topography, like southeast Greenland and the Antarctic Peninsula. Beyond these numerical costs, growing scheme complexity does not systematically translate into better physical realism. Proske et al. (2023) argue that excessive complexity



can hinder interpretability and mask unconstrained or compensating parameter choices rather than genuinely improve the model and show that a large part of the represented processes can be simplified or removed with limited impact on the results.

55 In this context, a stationary, diagnostic approach remains a pragmatic moderate-complexity alternative for large-scale climate models. It assumes that precipitation is in local equilibrium with the atmospheric column, i.e., that within a time step, microphysical sources and sinks balance and precipitation falls down to the surface with no horizontal advection. This assumption, widely used in traditional bulk schemes (Morrison and Gettelman, 2008), is reasonable at typical GCM resolutions ($\mathcal{O}(10\text{-}100\text{km})$) for many cloud regimes, including mixed-phase clouds and low-level ice clouds, where precipitation typically
60 reaches the surface over relatively short horizontal distances. It avoids introducing new prognostic variables and complex sedimentation resolution, thus reducing computational cost. However, the assumption may become less appropriate for slowly falling ice particles, particularly under strong wind conditions and at finer model resolutions.

The LMDZ atmospheric general circulation model (AGCM), atmospheric component of the IPSL-CM climate model (Hourdin et al., 2020), is more and more extensively used for polar climate simulations (Raillard et al., 2024; Wiener et al., 2025)
65 but suffers from notable precipitation representation biases (Madeleine et al., 2020; Roussel et al., 2020, 2023; Raillard et al., 2024) attributed to both ‘dynamical’ reasons – such as the calibration of horizontal dissipation (Lemonnier et al., 2021) – and shortcomings in the precipitation parameterization (Raillard et al., 2024). Moreover, despite recent developments to improve the representation of mixed-phase and ice clouds (Borella et al., 2025; Raillard et al., 2026; Vignon et al., 2026), biases persist in these clouds due to the precipitation representation itself. LMDZ is also increasingly used for water isotope studies in the
70 polar regions (Cauquoin et al., 2019; Dutrievoz et al., 2025, 2026), where an accurate representation of precipitation processes is essential for simulating the isotopic composition of snowfall. The current minimalist precipitation parameterization limits the LMDZ’s ability to represent precipitation processes adequately, motivating efforts to develop a more physically consistent precipitation scheme.

This work proposes the development of a new stationary and diagnostic precipitation scheme for the LMDZ model. It
75 includes both an advanced microphysical treatment of snowfall and subgrid-scale considerations to properly account for interactions between hydrometeors and clouds. Particular attention is also paid to the numerical treatment of the various processes to ensure convergence and stability at the typical time steps used in global climate models. Although this approach has limitations, it represents a first trade-off between the effort toward sophistication and the constraints of model complexity and computational cost. While the developments affect the precipitation treatment globally, we present a brief application and
80 evaluation on regional polar case studies in context where model biases have been clearly pinpointed. The paper is organized as follows: Section 2 introduces the LMDZ model and the shortcomings of its precipitation scheme. Section 3 presents the theoretical framework and implementation of the scheme in the model. A first test of the new parameterization is carried out using a single-column case in Section 4. Section 5 presents the evaluation and parametric sensitivity analysis, beginning with a brief description of the simulation configuration and the observational datasets used for evaluation in the Antarctic and Arctic
85 regions. Section 7 closes the paper with discussions and conclusions.



2 The LMDZ model and its shortcomings in the precipitation treatment

2.1 The LMDZ AGCM and the current precipitation scheme

This study uses the ICOLMDZ configuration of the LMDZ AGCM, which results from the coupling between the DYNAMICO icosahedral hydrostatic dynamical core (Dubos et al., 2015) and the physics of LMDZ (Hourdin et al., 2020). LMDZ is the atmospheric component of the IPSL climate model –named after the French climate Institute Pierre-Simon Laplace where it is developed– actively involved in the CMIP exercises (Boucher et al., 2020) and currently in preparation for CMIP7.

The reference cloud parameterization from the LMDZ6A model version used for CMIP6 is presented in detail Madeleine et al. (2020). In a nutshell, the LMDZ cloud scheme is based on a statistical scheme that assumes a subgrid probability density function (PDF) of total water vapour. Cloud fraction and condensates' specific content are diagnosed from the PDF under the assumption of instantaneous saturation adjustment. The shape of PDF depends on whether there is deep convection, surface initiated shallow convection or not (Vignon et al., 2026; Madeleine et al., 2020). Since CMIP6, cloud phase partitioning has been revised and now, vapour, liquid and ice are partitioned using a new turbulence-based phase partitioning parameterization (Raillard et al., 2026; Vignon et al., 2026).

The current precipitation parameterization is fairly minimalist and accounts for autoconversion of cloud ice crystals and liquid droplets, snow melting and rain freezing as well as the sublimation (resp. evaporation) of snow (resp. rain) coming from the overlying layer. The vertical profile of the precipitation flux is diagnosed assuming stationarity and equilibrium with the cloud field. The different processes are treated sequentially, with precipitation evaporation and snow melting applied before cloud condensation, and liquid and ice autoconversion and rain freezing applied afterwards. These processes are extensively described in Section 2.7 of Madeleine et al. (2020)

2.2 Shortcomings identified in the precipitation treatment

In the current scheme, several inconsistencies have been noted :

- Missing processes: none of the processes responsible for snowfall growth are explicitly parameterized, namely vapor deposition, aggregation, and riming.
- Precipitation evaporation: precipitation evaporates according to the grid-cell mean humidity. In reality, evaporation should occur primarily in the clear-sky fraction, while the cloudy fraction remains close to saturation. This can lead to an underestimation of the amount of precipitation reaching the surface (Touzé-Peiffer, 2021).
- Snow melting: the melting rate depends only on the temperature of the model layer; the warmer the layer, the more snow is melted. It does not depend on the layer thickness. Consequently, the same fraction of snow melts regardless of the vertical extent of the layer.



115 – Rain freezing: rain drops are assumed to freeze instantaneously whenever the air temperature falls below 0 °C. This prevents the simulation of freezing-rain events and supercooled liquid precipitation, which are frequent in the Antarctic (Silber et al., 2019).

– Snow autoconversion: as soon as cloud ice is present, snow is produced systematically. Furthermore, the current implementation of the autoconversion rate depends on the inverse of the layer depth following:

$$120 \quad \left. \frac{\partial q_i}{\partial t} \right|_{\text{auto}} = \frac{1}{\rho_a} \frac{\partial \rho w_i q_i}{\partial z} \approx \frac{\alpha_w q_i^{\beta_w + 1}}{\Delta z} \text{ with } w_i = \alpha_w q_i^{\beta_w} \quad (1)$$

where q_i is the cloud ice mixing ratio, w_i is the ice crystal fall velocity, ρ_a is the air density, Δz is the model layer thickness, and α_w and β_w are empirical coefficients controlling the fall-speed parameterization. This introduces an artificial sensitivity to the model's vertical resolution. Finer vertical grids result in stronger autoconversion rates, regardless of the physical conditions. This can result in the complete depletion of cloud ice near the surface in some cases.

125 3 Development of a Process Oriented Precipitation parameterization

Those shortcomings motivated the development of the POPRECIP (for Process Oriented Precipitation) scheme. Precipitation formation and evolution are parameterized with a one-moment approach considering ten processes that govern transfers between the specific contents of water vapor, liquid droplets, and ice crystals (respectively q_v , q_l , q_i), as well as the two precipitation fluxes of rain and snow (F_r , F_s), and the associated specific content q_r and q_s . These transfers are illustrated in Figure 1. Providing a given vertical overlap assumption for clouds – considered by default as a maximum-random overlap assumption– precipitation can evolve differently when falling into clear-sky (denoted by the *clr* superscript) and cloudy-sky regions (denoted by the *cld* superscript) (see section 3.1.1). The inclusion of a treatment of vertical overlap between cloud and precipitation is important, as it has been shown by Jakob and Klein (2000) to highly influence precipitation evaporation and its re-partitioning between clear-sky and cloudy regions.

135 This leads to the following four evolution equations:

$$\frac{\partial q_r^{cld}}{\partial t} + \mathbf{u} \cdot \nabla q_r^{cld} + \frac{1}{\rho_a} \frac{\partial F_r^{cld}}{\partial z} = \left. \frac{\partial q_r^{cld}}{\partial t} \right|_{\text{auto}} + \left. \frac{\partial q_r^{cld}}{\partial t} \right|_{\text{col}} + \left. \frac{\partial q_r^{cld}}{\partial t} \right|_{\text{melt}} + \left. \frac{\partial q_r^{cld}}{\partial t} \right|_{\text{fr}} \quad (2)$$

$$\frac{\partial q_r^{clr}}{\partial t} + \mathbf{u} \cdot \nabla q_r^{clr} + \frac{1}{\rho_a} \frac{\partial F_r^{clr}}{\partial z} = \left. \frac{\partial q_r^{clr}}{\partial t} \right|_{\text{melt}} + \left. \frac{\partial q_r^{clr}}{\partial t} \right|_{\text{fr}} + \left. \frac{\partial q_r^{clr}}{\partial t} \right|_{\text{eva}} \quad (3)$$

$$\frac{\partial q_s^{cld}}{\partial t} + \mathbf{u} \cdot \nabla q_s^{cld} + \frac{1}{\rho_a} \frac{\partial F_s^{cld}}{\partial z} = \left. \frac{\partial q_s^{cld}}{\partial t} \right|_{\text{auto}} + \left. \frac{\partial q_s^{cld}}{\partial t} \right|_{\text{agg}} + \left. \frac{\partial q_s^{cld}}{\partial t} \right|_{\text{melt}} + \left. \frac{\partial q_s^{cld}}{\partial t} \right|_{\text{fr}} + \left. \frac{\partial q_s^{cld}}{\partial t} \right|_{\text{rim}} \quad (4)$$

$$\frac{\partial q_s^{clr}}{\partial t} + \mathbf{u} \cdot \nabla q_s^{clr} + \frac{1}{\rho_a} \frac{\partial F_s^{clr}}{\partial z} = \left. \frac{\partial q_s^{clr}}{\partial t} \right|_{\text{melt}} + \left. \frac{\partial q_s^{clr}}{\partial t} \right|_{\text{fr}} + \left. \frac{\partial q_s^{clr}}{\partial t} \right|_{\text{sub}} \quad (5)$$

140 where the right-hand-side terms denote the tendencies associated with the different microphysical processes: autoconversion (auto), collection/accretion (col), aggregation (agg), riming (rim), melting (melt), freezing (fr), evaporation (eva), and sublimation (sub).



In our case, precipitation advection is not taken into account as the horizontal grid cells are generally too wide for precipitation to be advected in neighbouring grid cells before reaching the ground. Note that this assumption holds for typical grid cells of $\mathcal{O}(100\text{km})$, but it is no longer valid in strong wind conditions when running the model at $\mathcal{O}(10\text{ km})$ resolutions, especially for snowflakes. Following traditional bulk cloud scheme approach (Morrison and Gettelman, 2008), we consider that precipitation is close to equilibrium with the surrounding atmosphere and therefore close to stationarity, namely $\frac{\partial q_{r/s}}{\partial t} \sim 0$. Although questionable in conditions with slowly growing or decreasing hydrometeors such as in highly supersaturated conditions, those assumptions make it possible to:

- Avoid treating additional water species in the model physics, and advecting additional variables in the model dynamics, thereby saving computational resources. As we treat separately the in-cloud and clear-sky rain and snow water content, this would either require advecting four additional variables. Alternatively, one could advect only rain and snow, and reconstruct their partitioning between cloudy and clear-sky at each time step, but this would require an additional assumption (e.g., that the partitioning of precipitation content follows the cloud fraction), which is not necessarily valid.
- Diagnose the precipitation flux at the bottom of a given layer as the sum of the flux at the top of the layer and the source and sink terms within the layer. Therefore, there is no need to numerically treat the sedimentation term which is particularly challenging (Bouteloup et al., 2011) and often requires the use of sub-timestepping procedures (e.g., Chosson et al. (2014));

We do acknowledge the potential added value of the prognostic treatment of precipitation species especially when increasing the horizontal resolution of the model (Gettelman et al., 2015) and this may be an avenue for a further refinement of the scheme. However, the approach followed hereafter is a trade-off between sophistication effort and model complexity, and numerical cost

The processes are treated sequentially: those labelled as A_n in Figure 1 are applied before cloud condensation while those labelled as " B_n " are applied after cloud condensation.

The following section presents the analytical development of the parameterization followed by focus on its numerical implementation/treatment.

3.1 Clear-sky vs cloudy precipitation treatment and formulations of the different microphysical processes

Precipitation fluxes are by convention, defined positive downwards. We define the following writing conventions: grid box mean values (fluxes or specific quantities) are denoted with overlines ($\overline{F_Y}$, $Y \in (r, s)$, $\overline{q_X}$, $X \in (v, l, i, r, s)$), corresponding to water vapor, cloud liquid, cloud ice, rain, and snow, respectively whereas in-cloud variables are referred to as F_Y, q_X for conciseness. Additionally, as precipitation fluxes are defined in clear and cloudy sky, we introduce the superscript Z where $Z \in \{clr, cld\}$ refers to the variable in the clear sky or cloudy part of the grid cell. The cloud fraction α_c denotes the fraction of the grid cell covered by cloud, while α_{pr}^Z denotes the fraction of the corresponding clear- or cloudy-sky area occupied by precipitating hydrometeors. The link between grid-box mean and in-cloud specific quantities is:

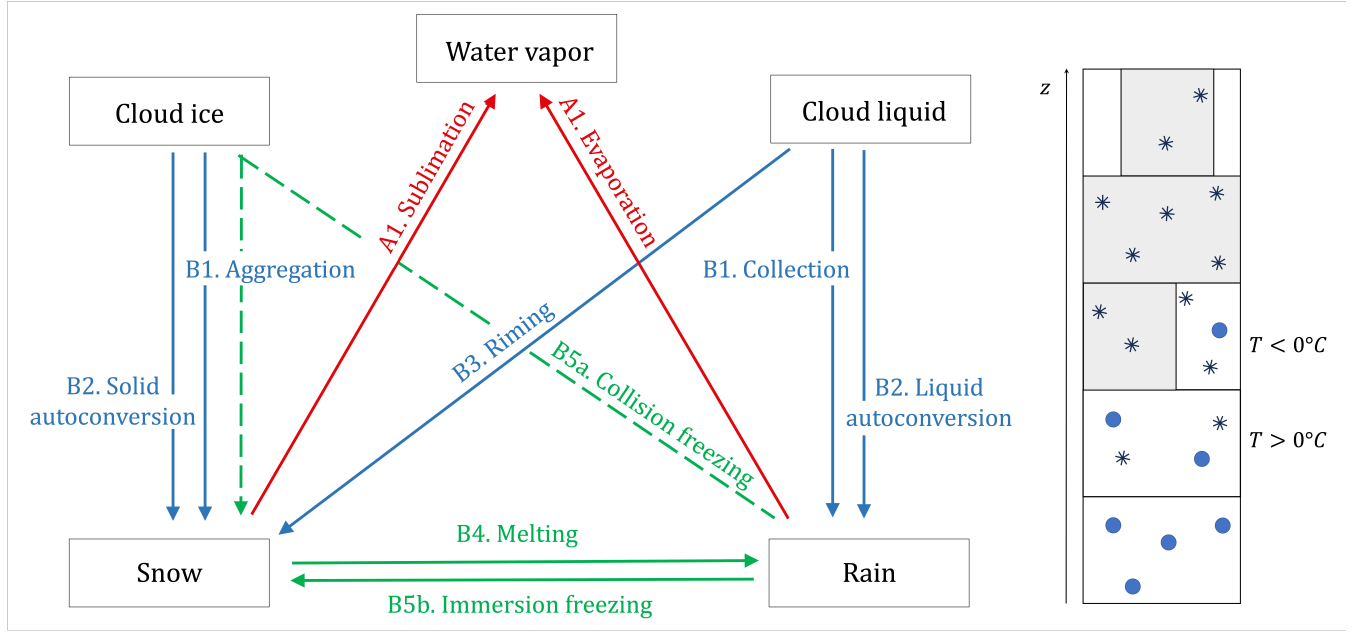


Figure 1. Left: schematics of the transfers of mass between the five water variables, through the 10 processes parameterized in the PO-PRECIP scheme. Red arrows indicate precipitation sinks, blue arrows indicate sources and green arrows represent transfer between the two precipitation phases. Numbers denote the order in which processes are treated within either the block A (before cloud formation) or B (after cloud formation). Right : schematics of a model column with cloud and precipitation overlap.

$$175 \quad \begin{cases} q_X = \overline{q_X} \\ F_Y^Z = \frac{\alpha_c}{\alpha_{pr}^Z} \end{cases} \quad (6)$$

Note also that for the state variables T, p the overline is omitted as we consider that temperature and pressure are homogeneously distributed within the grid-box.

The aforementioned microphysical processes are implemented sequentially in each layer, and within a vertical loop, from top to bottom. Each vertical level receives a precipitation flux from the level immediately above, with a precipitation flux
180 initialized to zero at the top level.

3.1.1 Cloudy and clear-sky precipitation fractions

Before computing the microphysical source and sink terms for rain and snow precipitation, the fraction of the precipitation flux falling in cloudy air and the fraction falling in clear sky are determined. This step is performed right after cloud formation, which provides the cloud fraction at level k , denoted α_c^k . Following an approach closed to that of Jakob and Klein (2000),
185 the total precipitation fraction and associated precipitation fluxes are partitioned into cloudy-sky and clear-sky contributions.



Such as in Chosson et al. (2014), it is assumed that i) precipitation comes from overlapping precipitation regions – including overlapping clouds and possibly clear-sky precipitation fraction – or is produced in the cloud fraction at the same model level, ii) rain and snow hydrometeors are homogeneously distributed in the precipitation fraction, and iii) possible horizontal tilt of precipitation drafts under clouds in case of wind shear is neglected. By default, a vertical maximum-random overlap between total precipitation fractions is assumed - i.e two precipitation fractions in adjacent vertical layers are assumed to maximally overlap, while the overlap of precipitation layer separated by a clear layer is randomly assumed. At each model level, the total precipitation flux and fraction are decomposed into a cloudy and a clear-sky contribution as follows:

$$\alpha_{pr} = \alpha_{pr}^{clr} + \alpha_{pr}^{cld} \quad (7)$$

where α_{pr}^{clr} and α_{pr}^{cld} denote the fractions of precipitation falling respectively in clear sky and in the cloud. Similarly, rain and snow fluxes are partitioned into cloudy and clear-sky components. The routine first diagnoses the total precipitation fraction in the current layer from the precipitation fraction in the layer above and the cloud fraction at the current level. Following the maximum-random overlap assumption, the updated total precipitation fraction at level k is computed following Jakob and Klein (2000):

$$\alpha_{pr}^k = 1 - (1 - \alpha_{pr}^{k+1}) \left(\frac{1 - \max(\alpha_c^k, \alpha_{pr}^{cld,k+1})}{1 - \alpha_{pr}^{cld,k+1}} \right) \quad (8)$$

where level k increases with altitude. The precipitation fractions are then updated to account for the change in cloud fraction between level $k + 1$ and level k . We denote by $\Delta\alpha_{pr}^{cld,k}$ and $\Delta\alpha_{pr}^{clr,k}$ the corresponding vertical changes (level-to-level) in the cloudy sky:

$$\Delta\alpha_{pr}^{cld,k} = \alpha_{pr}^{cld,k+1} - \alpha_{pr}^{cld,k} = \alpha_c^k - \alpha_{pr}^{cld,k+1} \quad (9)$$

$$\Delta\alpha_{pr}^{clr,k} = \alpha_{pr}^{clr,k+1} - \alpha_{pr}^{clr,k} = \alpha_{pr}^k - \alpha_c^k - \alpha_{pr}^{clr,k+1} \quad (10)$$

If the cloud fraction increases from level $k + 1$ to level k , then part of the clear sky precipitation flux is transferred to the cloud precipitation flux, and vice versa. The transferred precipitation flux is proportional to the corresponding fractional change in precipitation fraction.

Once all the processes have been applied, to avoid weak precipitation flux in clear sky with large corresponding fraction, the clear sky precipitation fraction can be updated. In practice if the clear sky precipitation flux is lower than an fixed threshold F_{min} the clear-sky precipitation fraction is reduced so that the local precipitation flux exactly matches the threshold.

Once the vertical overlap of precipitation has been specified, the different source and sink terms of rainfall and snowfall can be computed.

3.1.2 A1. Temperature adjustment, evaporation and sublimation

Note that for any phase change process, a tendency of the mesh temperature is computed accordingly. Before applying the individual microphysical processes, the mean air temperature in each layer is adjusted to account for the enthalpy flux associated

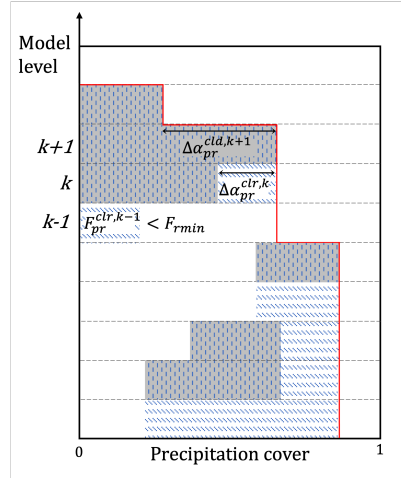


Figure 2. Schematic illustration of cloud and precipitation overlap in an atmospheric column. Gray shading denotes cloud fraction, assumed to follow a maximum-random overlap. Blue vertical hatching indicates the in-cloud precipitation fraction, while blue diagonal hatching indicates the clear-sky precipitation fraction. The red step line represents the cumulative maximum-random precipitation cover from the top of the atmospheric column to the current model level. At level k , the precipitation fraction is partitioned into clear-sky $\alpha_{pr}^{clr,k}$ and in-cloud $\alpha_{pr}^{cld,k}$, with the corresponding increments $\Delta\alpha_{pr}^{clr,k}$ and $\Delta\alpha_{pr}^{cld,k}$. If the precipitation flux F_{pr}^{clr} decreases down to the threshold F_{rmin} (illustrated at model level $k-1$), the clear-sky precipitation fraction and flux are reduced.

with the sedimentation of precipitation from the overlying layer. This adjustment ensures that the precipitation entering a given layer and the surrounding air have the same temperature, thereby ensuring energy conservation in the global coupled climate model. Evaporation of rainfall and sublimation of snowfall are parameterised in a similar way as in LMDZ6A, following the approaches of Sundqvist (1988), Klemp and Wilhelmson (1978) and Schlesinger et al. (1988):

$$220 \quad \left. \frac{\partial q_r^{clr}}{\partial t} \right|_{eva} = -\frac{1}{\rho} \frac{\partial F_r^{clr}}{\partial z} = -\frac{1}{\rho} \gamma_{eva} \left(1 - \frac{q_{tot}^{clr}}{q_{sl}} \right) (F_r^{clr})^{\delta_{eva}} \quad (11)$$

$$\left. \frac{\partial q_s^{clr}}{\partial t} \right|_{sub} = -\frac{1}{\rho} \frac{\partial F_s^{clr}}{\partial z} = -\frac{1}{\rho} \gamma_{sub} \left(1 - \frac{q_{tot}^{clr}}{q_{si}} \right) (F_s^{clr})^{\delta_{sub}} \quad (12)$$

where q_{tot} is the total specific humidity in the grid cell, q_{sl}, q_{si} the saturation specific humidity with respect to liquid and ice. $\gamma_{eva}, \gamma_{sub}$ are calibration parameters set by default to 2×10^{-5} . δ_{eva} and δ_{sub} are constant parameters set to 0.5 following Sundqvist (1988).

225 Note that in the model, this process has been implemented before cloud formation to include the evaporated water into the available total water for cloud condensation. However, as the cloud fraction at the corresponding layer is not known yet, we partition the clear-sky and the cloudy fractions of the precipitation flux considering the cloudy and clear-sky precipitation fractions from the layer immediately above. To compute the clear-sky humidity q_{tot}^{clr} , we distinguished two possible options. The first one assumes $q_{tot}^{clr} \approx q_{tot}$. Although this is a quite crude approximation, Touzé-Peiffer (2021) underlined that if $\alpha_c \ll 1$,



230 q_{tot}^{clr} is very close to q_{tot} and if $0 < \alpha_c < 1$, under a maximum-random assumption for cloud overlap, the clear-sky precipitation flux will be relatively small and the choice of q_{tot}^{clr} or q_{tot} for the calculation of evaporation will not strongly affect moisture tendencies in the grid box To compute the vertical decrease in precipitation flux associated with precipitation and sublimation within the clear-sky fraction of a given layer, Eq. 11 (resp. 12) is resolved numerically with an exact treatment of F_r^{clr} (resp. F_s^{clr}) that is, the resolution of classical linear ordinary differential equation, and an explicit treatment of q_{tot}^{clr} . Such a numerical
235 resolution guarantees the positivity of the precipitation flux. Evaporation and sublimation are also limited such that the humidity in the clear-sky fraction does exceed saturation.

3.1.3 B1. Collection by liquid drops and aggregation of snowflakes

The collection of liquid droplets in suspension by raindrops and the aggregation of snowflakes and ice crystals in suspension are considered here. Note that we consider only the collection and aggregation between cloud condensates and precipitation fluxes.
240 These processes are treated before autoconversion to avoid collecting and aggregating the newly formed fluxes, which have already interacted with the cloud. Aggregation and collection depend on collisions between cloud condensates in suspension and precipitating hydrometeors, and therefore occur only in the cloudy part of the grid cell. The expressions are derived from Muench and Lohmann (2020) assuming mono-dispersed populations of spherical rain drops and snowflakes:

$$\left. \frac{\partial q_r}{\partial t} \right|_{col} = - \left. \frac{\partial q_l}{\partial t} \right|_{col} = \gamma_{col} \frac{3}{4} \frac{1}{r_{eff,r} \rho_r} F_r^{cld} q_l E_{rl} \quad (13)$$

$$245 \quad \left. \frac{\partial q_s}{\partial t} \right|_{agg} = - \left. \frac{\partial q_i}{\partial t} \right|_{agg} = \gamma_{agg} \frac{3}{4} \frac{1}{r_{eff,s} \rho_s} F_s^{cld} q_i E_{si} \quad (14)$$

where $\gamma_{col}, \gamma_{agg}$ are tuning coefficients that control the efficiency of the processes. The parameter ρ_r is the rain (liquid water) density and snow density is computed following Brandes et al. (2007) :

$$\rho_s = 178 (2r_s 1000)^{-0.922} \quad (15)$$

with the snowflake radius $r_s = r_{eff,s}$ being under the sphericity assumption. The parameters E_{rl}, E_{si} are the collection and
250 aggregation efficiencies respectively. The collection efficiency is assumed to equal unity $E_{rl} = 1$. For aggregation efficiency, we use the temperature-dependent formula from Lin et al. (1983):

$$E_{si} = \exp(0.025(T - T_0)) \quad (16)$$

where $T_0 = 273.15$ K is the freezing point of water. As we consider a one-moment approach for the treatment of hydrometeors, the effective radii of rain drops $r_{eff,r}$ and snowflakes $r_{eff,s}$ must be fully parameterised. For rain drops, as we assume a mono-
255 dispersed population for collection processes, we set it to constant value equal to $500 \mu\text{m}$ by default. For snowflakes, as a first approach we decide to set $r_{eff,s}$ as constant value (equal to 1 mm by default). This quite crude approximation is an avenue for improvement for the scheme. However, it avoids introducing additional parameters – such as parameters relating $r_{eff,s}$ with the ice water content and/or snowfall flux and/or temperature – and therefore avoid complexifying even more parametric



exploration exercises (see Section 5.3.2). Eq. 13 (resp. 14) is solved exactly in q_l (resp. q_i): the solution of this equation has an
260 exponential form, which guarantees the positivity of the cloud condensates mass during the process.

3.1.4 B2. Autoconversion

The autoconversion of cloud droplets to precipitating rain drops remains unchanged compared to LMDZ6A. It is based on the
original ideas of Sundqvist (1988) which assumes that precipitation generation occurs at a given rate when the in-cloud liquid
water content exceeds a given threshold value (Liu et al., 2006). Given the deficiencies of the previous formulation for ice
265 autoconversion, a similar formulation is adopted for the conversion of cloud ice to snow, owing to its simple formulation. The
two autoconversion rates read:

$$\left. \frac{\partial q_r}{\partial t} \right|_{\text{auto}} = - \left. \frac{\partial q_l}{\partial t} \right|_{\text{auto}} = \frac{q_l}{\tau_{\text{rain}}} \left(1 - \exp \left[- \left(\frac{q_l}{q_{th,r}} \right)^{\kappa_r} \right] \right) \quad (17)$$

$$\left. \frac{\partial q_s}{\partial t} \right|_{\text{auto}} = - \left. \frac{\partial q_i}{\partial t} \right|_{\text{auto}} = \frac{q_i}{\tau_{\text{snow}}} \left(1 - \exp \left[- \left(\frac{q_i}{q_{th,s}} \right)^{\kappa_s} \right] \right) \quad (18)$$

where $\tau_{\text{rain}}, \tau_{\text{snow}}$ are rain and snow autoconversion timescales, $q_{th,r}, q_{th,s}$ are thresholds in liquid and ice amounts above
270 which autoconversions sharply increase, and κ_r, κ_s are power values set to 2 following Sundqvist (1988). We make the snow
autoconversion timescale depend on the liquid content of the cloud, to account for the fact that snow production is more
efficient in presence of supercooled liquid water where the Wegener-Bergeron-Findeisen process is active. It varies between
two values $\tau_{\text{snow,min}}$ and $\tau_{\text{snow,max}}$ as follows:

$$\tau_{\text{snow}} = \tau_{\text{snow,max}} + (\tau_{\text{snow,max}} - \tau_{\text{snow,min}}) (x_{liq})^{\kappa_\tau} \quad (19)$$

275 where $\tau_{\text{snow,min}}, \tau_{\text{snow,max}}$ and κ_τ are adjustable parameters, and x_{liq} denote the liquid fraction in the cloud. The numerical
resolution of the autoconversion equations is particularly critical as the resolution with an explicit scheme of Eq. 17 and 18 at
typical GCM time steps (15 min for the standard LMDZ physics) often leads to artificial total depletion of the cloud liquid and
ice contents. In LMDZ6A, a sub-timestepping procedure was employed for the autoconversion of liquid droplets to rain but
increased substantially the numerical cost in this part of the code. Here, we choose a numerically cheaper one-step alternative
280 consisting in treating the right-hand side term of Eq. 17 and 18 explicitly – that is, taking q_l and q_i in the exponential at the
beginning of the time-step – such that the equations can then be integrated exactly as first-order linear differential equations in
 q_l and q_i . This ‘semi-exact’ scheme shows a good agreement with the explicit resolution with sub-time steps as illustrated in
Fig 3.

3.1.5 B3. Riming

285 Riming of cloud liquid droplets occurs through collisions with falling snowflakes, converting liquid droplets in suspension into
snow. This process takes place only within the cloudy part of the grid cell. The formula, which has the same form as those for
aggregation and collection, is inspired from Muench and Lohmann (2020):

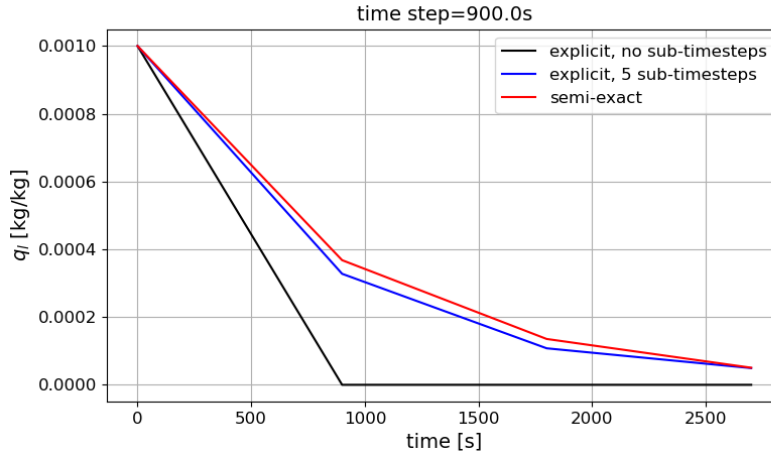


Figure 3. Comparison of the numerical integration in time of the liquid droplets autoconversion equation (Eq. 17). For this example, the initial q_l value has been set to 0.001 kg kg^{-1} , τ_{rain} to 3600 s and $q_{th,r}$ to $6.5 \times 10^{-4} \text{ kg}^{-1}$.

$$\left. \frac{\partial q_l}{\partial t} \right|_{rim} = -\gamma_{rim} \frac{3}{4} \frac{1}{r_{eff,s} \rho_s} F_s^{cld} q_l E_{sl} \quad (20)$$

where γ_{rim} is a tuning coefficient (Table 1) and E_{sl} the riming efficiency set to 1 (Ferrier, 1994).

290 3.1.6 B4. Melting

To derive a formulation for melting we adopt a heuristic approach considering the melting of one snowflake and then apply this to a population of snowflakes. We start from an equation for one snowflake mass m_s evolution due to melting from Wilson and Ballard (1999):

$$\left. \frac{\partial m_s}{\partial t} \right|_{melt} = -4\pi \frac{K_a C_s}{L_f} (T_{wi} - T_0) f_v \quad (21)$$

295 where K_a is the thermal conductivity of air, C_s the snowflake capacitance, L_f the latent heat of fusion, T_{wi} the ice wet-bulb temperature, and f_v the ventilation factor and $T_0 = 273.15 \text{ K}$ is the pure water freezing point. Melting is triggered when $T_{wi} > T_0$. The ice wet-bulb temperature accounts for evaporative cooling during snow melting in subsaturated air (Heymsfield et al., 2021). As meltwater evaporates, it cools the particle, allowing snow particles to survive at slightly warmer temperatures when the relative humidity is low.

300 Assuming spherical snowflakes of radius r_s , mass density ρ_s , and a fall speed v_s^Z , the number of snowflakes N_s^Z can be derived from the snow flux F_s^Z .



$$N_s^Z = \frac{F_s^Z}{v_s^Z \frac{4}{3} \pi r_s^3 \rho_s} \quad (22)$$

$$\left. \frac{\partial q_s^Z}{\partial t} \right|_{\text{melt}} = N_s^Z \left. \frac{\partial m_s^Z}{\partial t} \right|_{\text{melt}} \quad (23)$$

where $Z \in \{clr, cld\}$ refers to the variable in the clear sky or cloudy part of the grid cell. The ventilation factor is absorbed in
305 a tuning coefficient γ_{melt} (Table 1) and the following formula for melting is obtained :

$$\left. \frac{\partial q_s^Z}{\partial t} \right|_{\text{melt}} = -\gamma_{\text{melt}} N_s^Z 4\pi \frac{K_a C_s}{L_f} (T_{wi}^Z - T_0) \quad (24)$$

Note that as the cloudy part is saturated, $T_{wi}^{cld} = T$. For now, the snow fall speeds in clear and cloudy sky are assumed equal with $v_s^{clr} = v_s^{cld} = 1 \text{ m s}^{-1}$. The snow effective radius is also prescribed with $r_s = 1\text{mm}$ and snow density is computed with Eq. 15.

310 Finally, a maximum snow quantity $\Delta q_{s,max}$ that can melt is imposed to ensure the temperature is not brought back below freezing:

$$\Delta q_{s,max}|_{\text{melt}} = \frac{C_p (T_0 - T)}{L_f} \left(1 + \left(\frac{c_{pv}}{c_{pd}} - 1 \right) q_{tot} \right). \quad (25)$$

3.1.7 B5. Rain freezing

First, the freezing of rain into snow upon collision with ice crystals is considered. During this process, falling raindrops freeze
315 after colliding with ice crystals, which act as ice nucleating particles. This process occurs only in the cloudy part of the grid cell. The collision rate between ice crystals in suspension and falling raindrops is estimated by adapting the riming formulation of Muench and Lohmann (2020), yielding:

$$\left. \frac{\partial q_i}{\partial t} \right|_{\text{fr,col}} = -\gamma_{fr} \frac{3}{4} \frac{1}{\rho_r r_r} F_r^{cld} q_i E_{ri} \quad (26)$$

with the freezing efficiency of rain drops and ice crystals $E_{ri} = 1$. The portion of rain that freezes is then added to q_s . To
320 determine this amount, the number of raindrops that freeze N_r is assumed to equal the number of collected ice crystals N_i (i.e. $N_r = N_i$). Assuming spherical particles, the corresponding changes in rain and ice specific content, respectively dq_r and dq_i are related by their particle masses:

$$dq_r = \frac{4}{3} \pi \rho_r r_r^3 dN_r = \frac{4}{3} \pi \rho_i r_i^3 dN_i \frac{\rho_r r_r^3}{\rho_i r_i^3} \quad (27)$$

with the ice crystals radius parameterized as a function of temperature and ice water content following Sun and Rikus (1999).



325 Ultimately, the amount of rain that has frozen follows:

$$\left. \frac{\partial q_r}{\partial t} \right|_{\text{fr,col}} = \left. \frac{\partial q_i}{\partial t} \right|_{\text{fr,col}} \frac{\rho_r r_r^3}{\rho_s r_s^3} \quad (28)$$

and the snow mass specific content increases as:

$$\left. \frac{\partial q_s}{\partial t} \right|_{\text{fr,col}} = - \left. \frac{\partial q_i}{\partial t} \right|_{\text{fr,col}} - \left. \frac{\partial q_r}{\partial t} \right|_{\text{fr,col}} \quad (29)$$

330 The second pathway for rain to freeze is immersion freezing. The parameterization is inspired from the original ideas of Bigg (1953) and consists in an exponential decrease of rain specific humidity with a given characteristic freezing timescale. The latter depends exponentially on temperature so that it is equal to 1 for temperature of homogeneous freezing T_{hom} and is close to 0 for water melting point T_0 . This step is applied separately in clear and cloudy sky.

$$\left. \frac{\partial q_s}{\partial t} \right|_{\text{fr,imm}} = \frac{-q_r}{\tau_{fr,imm}} \quad (30)$$

$$\frac{1}{\tau_{fr,imm}} = \beta_{fr,imm} \exp \left(-\gamma_{fr,imm} \frac{T - T_{hom}}{T_0 - T_{hom}} \right) \quad (31)$$

335 where T is the temperature, $T_{hom} = 235.15\text{K}$, $\gamma_{fr,imm}$, $\beta_{fr,imm}$ are tuning coefficients (Table 1). As for melting, the total amount of rain that can freeze is limited by a threshold to prevent the temperature from rising above 0°C .

$$\Delta q_{r,max}|_{\text{fr}} = \frac{C_p(T - T_0)}{L_f} \left(1 + \left(\frac{c_{pv}}{c_{pd}} - 1 \right) q_{tot} \right) \quad (32)$$

4 Illustration of the parameterization effects in Single-Column simulations of the COMBLE case

340 The impact of the POPRECIP parameterization on the representation of polar snowfall is first illustrated using single-column simulations of the Cold-air Outbreaks in the Marine Boundary Layer Experiment–Model Intercomparison Project (COMBLE-MIP) case (Juliano et al., 2026).

345 The COMBLE-MIP case is an intercomparison exercise involving Large-Eddy Simulations (LESs) and Single-Column Models (SCMs). It adopts a quasi-Lagrangian framework to investigate the formation and evolution of the boundary layer, clouds, and precipitation during a major marine cold-air outbreak that originated over the sea ice west of the Svalbard archipelago and propagated over the open ocean toward the Norwegian coast. The strong increase in surface turbulent fluxes over the open ocean leads to a rapid deepening of the boundary layer downstream of the sea-ice edge, accompanied by vigorous convective activity and the development of mixed-phase convective boundary-layer clouds. During the approximately 20-hour transit over the open ocean, these clouds produce substantial snowfall rates (approximately $0.2\text{--}0.3 \text{ mm h}^{-1}$), while the boundary-layer organization undergoes the characteristic transition from rolls to open cells. Throughout the event, the supercooled liquid water



Name	Description	Range	Units
<i>Cold precipitation parameters</i>			
$q_{th,s}$	Threshold triggering snow autoconversion	$[1e-6 - 1e-4]$	$[kg\ kg^{-1}]$
$\tau_{snw,max}$	Typical maximum timescale of snow autoconversion	$[1800 - 7200]$	[s]
$\tau_{snw,min}$	Typical minimum timescale of snow autoconversion	$[1800 - 7200]$	[s]
γ_{sub}	Controls the strength of sublimation	$[1e-4 - 5e-4]$	$[m^{-1}]$
γ_{rim}	Controls the strength of riming	$[0.5 - 2]$	[-]
γ_{agg}	Controls the strength of aggregation	$[0.5 - 2]$	[-]
<i>Warm precipitation parameters</i>			
$q_{th,r}$	Thresholds triggering rain autoconversion	$[1e-4 - 1e-3]$	$[kg\ kg^{-1}]$
τ_{rain}	Typical timescale of rain autoconversion	$[4000 - 15000]$	[s]
γ_{col}	Controls the strength of collection	$[0.5 - 2]$	[-]
γ_{eva}	Controls the strength of rain evaporation	$[5e-5 - 5e-4]$	$[m^{-1}]$
<i>Mixed parameters</i>			
γ_{melt}	Controls the strength of melting	$[0.5 - 2]$	[-]
$\gamma_{fr,imm}$	Controls the strength of immersion freezing	$[0.5 - 2]$	[-]
$\beta_{fr,imm}$	Controls the time of immersion freezing	$[0.5 - 2]$	[-]

Table 1. Name, description, and range of acceptable values for the adjustable parameters of POPRECIP. Parameters in bold are used for the sensitivity experiment.

content remains moderate, with liquid water path values generally ranging between 0 and $0.1\ kg\ m^{-2}$. Details of the initial conditions and forcing prescribed for the single-column simulations are provided by Juliano et al. (2026).

LMDZ participated in the COMBLE-MIP intercomparison using a physics package based on LMDZ6A, but including the recent turbulent diffusion (Vignon et al., 2024) and cloud-phase partitioning (Raillard et al., 2026; Vignon et al., 2026) parameterizations. Figure 4 illustrates the impact of activating the POPRECIP parameterization (POPRECIP simulation), using the default values of its free parameters, relative to the standard LMDZ precipitation scheme (CTRL simulation).

Figures 4a–c show that activating POPRECIP substantially reduces the cloud condensed water content and maintains the liquid water path within the range of observed values. An analysis of the snowfall growth tendencies (not shown) indicates that this reduction primarily results from enhanced snowfall production through the autoconversion of ice crystals in the presence of supercooled liquid water near cloud top, with smaller contributions from riming and aggregation. Consequently, cloud condensate is converted into precipitation more efficiently in the POPRECIP simulation. The only exception occurs during the final two hours of the simulation, when the strong increase in supercooled liquid water in the CTRL simulation enhances the production of frozen precipitation through the freezing of rain.

In the CTRL simulation, snowfall exhibits a sharp and likely unrealistic increase near cloud base. This behavior arises from a subtle interaction with the cloud-phase partitioning parameterization, which produces relatively large ice water contents near



cloud base. As the model layers become progressively thinner toward the surface, this ice is almost entirely converted into snowfall (see Eq. 1), resulting in an unrealistically sharp increase in the snowfall flux.

Figures 4d and e further show that, in the subsaturated layer below cloud base, the snowfall flux remains nearly constant in the CTRL simulation, whereas it decreases toward the surface in the POPRECIP simulation, despite sublimation being computed from the grid-box mean relative humidity by default. In POPRECIP, snowfall below cloud base is confined to the clear-sky precipitation fraction, making the local clear-sky snowfall flux and thus the sublimation rate (see Eq. 12) more intense. This leads to enhanced sublimation and a more realistic decrease in snowfall toward the surface.

Other single column model cases have been performed and are summarized in Appendix A. It is worth emphasizing that, beyond modifying the representation of snowfall and mixed-phase polar clouds, the POPRECIP scheme and in particular the introduction of droplet collection and the new treatment of rain evaporation, inevitably affects the simulation of warm clouds. Appendix A shows that, after retuning the cloud and boundary-layer parameters, the representation of typical warm, weakly precipitating boundary-layer clouds in single-column simulations remains comparable to that obtained with the standard LMDZ precipitation scheme.

5 Methodology, observations and model configuration for the Arctic and Antarctic case studies

5.1 Evaluation strategy

Evaluating the representation of snow precipitation is challenging because snowfall measurements are subject to substantial observational uncertainties. Evaluating the added value of a precipitation parameterization is also inherently difficult because precipitation is the end product of a chain of interacting processes, including model dynamics and numerous physical parameterizations. Consequently, our evaluation strategy focuses less on quantitative performance metrics and more on process-oriented diagnostics that provide insight into the behaviour of the scheme and can be used to guide its calibration.

We evaluate the POPRECIP parameterization using two complementary case studies: a frontal precipitation event over coastal Antarctica and a month-long airborne campaign in frontal clouds over Svalbard region. These case studies provide access to different observational datasets and physical processes, allowing the scheme to be evaluated in distinct polar environments. For each case study, a dedicated regional model configuration is set up and compared with the available observations. The parameterization is first evaluated using a reference set of parameters, before a perturbed parameter ensemble (PPE) is performed to investigate its parametric sensitivity.

5.2 Observational datasets

5.2.1 The RALI-THINICE campaign

The RALI-THINICE campaign was the French component of the international THINICE field campaign conducted over Svalbard in August 2022 to investigate synoptic-scale Arctic cyclones and their interactions with cloud microphysics (Rivière

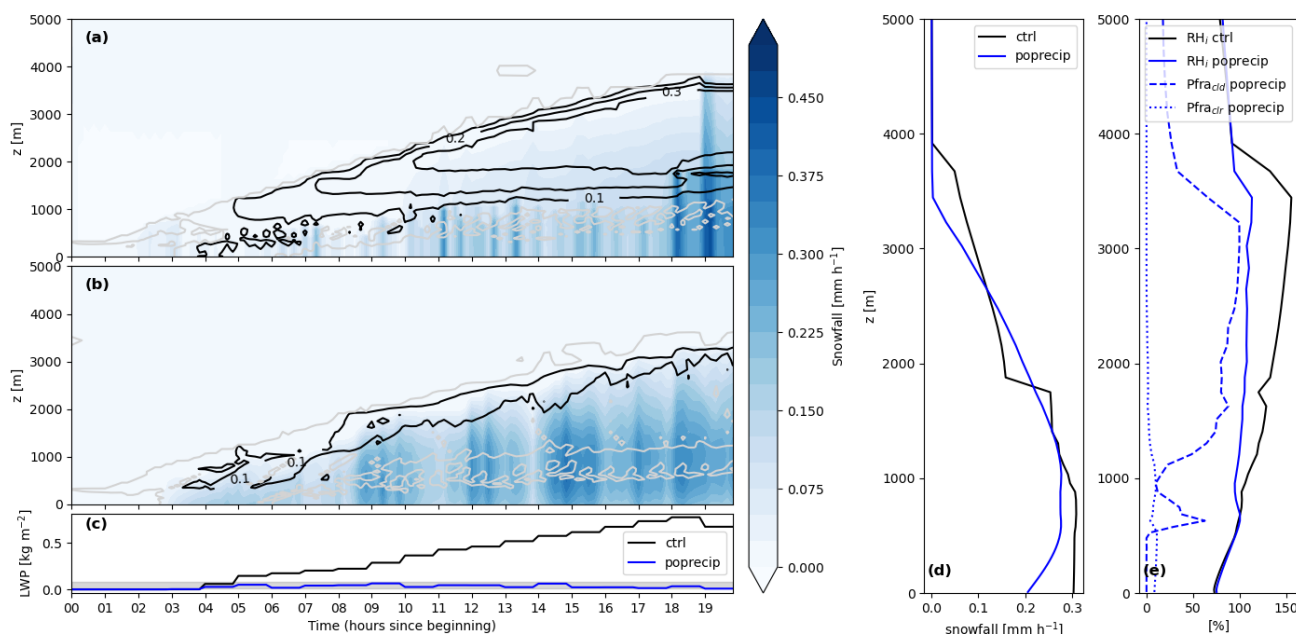


Figure 4. Analysis of the COMBLE-MIP single column simulations with LMDZ. Panels a and b: Time evolution of the snowfall vertical profile (shading) and of the condensed (liquid + ice) cloud water contents (black contours every g kg^{-1}) in the CTRL (a) and POPRECIP (b) simulations. Grey contours indicate the 50% cloud fraction line. Panel c shows the time evolution of the liquid water path (LWP). Grey shading shows the range of satellite-based and ground-based remote-sensing retrievals shown in Juliano et al. (2026). Panel d shows the vertical profiles of snowfall averaged over the last 2 hours of simulation. Panel e shows the vertical profiles of relative humidity with respect to ice (RH_i) averaged over the last two hours of simulations in the two configurations (solid lines). Dashed (resp. dotted) lines show the cloudy (resp. clear-sky) precipitation fractions in the POPRECIP simulation.

et al., 2024). During the campaign, the French ATR42 research aircraft was based in Longyearbyen (78.23°N , 14.90°E) and performed 16 flights sampling a variety of cloud systems associated with cyclones, primarily mid-level frontal clouds.

The simulated cloud and precipitation properties are evaluated using both in situ and remote-sensing observations. In situ cloud microphysical properties are derived from the combined measurements of the CDP-2, 2D-S and HVPS probes. The total liquid water content (LWC) is inferred from the combined particle size distribution (PSD) measured by the CDP-2 and 2D-S probes, assuming spherical droplets. The total ice water content (IWC) is retrieved using the combined PSD from the 2D-S and the HVPS and the modified Brown and Francis mass-diameter relationship (Hogan et al., 2012). The Radar System Airborne (RASTA), a 95 GHz Doppler cloud radar, provides vertical profiles of cloud properties above and below the aircraft. Of particular interest here are the retrieved cloud mask and ice water content, which are derived from the radar measurements (Delanoë and Hogan, 2008; Cazenave et al., 2019). Snowfall flux is also estimated by multiplying the IWC retrieval with the retrieved vertical velocity (Raillard et al., 2024).



5.2.2 The APRES3 campaign

The APRES3 (Antarctic Precipitation, Remote Sensing from Surface and Space) campaign was conducted during the austral summer of 2015–2016 at the French Antarctic station Dumont d’Urville (DDU), located in coastal Adélie Land, Antarctica (66.66°S, 140.00°E; 41 m above sea level).

410 During APRES3, several instruments were deployed at DDU. Their deployment and the data processing is thoroughly described in Grazioli et al. (2017a); Genthon et al. (2018). The three instruments used in this study are listed below:

- a Micro Rain Radar (MRR): Installed inside a radome and operating in vertical mode, the MRR scans the atmosphere from 341 to 3141 m a.s.l. with a vertical resolution of 100 m.
- a Mobile X-band Dual-Polarization Radar (MXPOL): A scanning radar capable of operating at multiple elevation angles
415 (from vertical to low-level radial scans), with a maximum radial range of 30 km and a radial resolution of 75 m. It performed plan position indicator (PPI) scans, range height indicator (RHI) scans as well as vertical profiles of the atmospheric column every 5 min.
- a Weighing Precipitation Gauge (Pluvio2): Measures the liquid water equivalent content of snowfall. The gauge is installed at approximately 3 m above ground level to minimize contamination from blowing snow and data were processed
420 to avoid the occurrence of ‘phantom’ precipitation (Grazioli et al., 2017a).

Vertical profiles of snowfall fluxes at DDU can be estimated using a combination of the MRR and the Pluvio2, under a set of assumptions that link radar reflectivity to snowfall rate. While these retrievals provide valuable products, they remain subject to uncertainties related primarily to the choice of reflectivity–snowfall (Z-S) relationships. In this study, we use two independent retrieval approaches based on MRR observations during the APRES3 period. The first method follows Grazioli
425 et al. (2017a), who derived a Z-S rate relationship with its associated uncertainties, specifically for DDU using 3 months of MRR and Pluvio2 data. Their approach accounts for site-specific snowfall microphysics and includes corrections for blowing snow contamination and near-surface sublimation. The second method is taken from Wiener et al. (2023), who derived a Z-S relationship using a seven-year time series and a refined retrieval algorithm with more restrictive filters on the data. To estimate the uncertainty on the mean snowfall rate over a given time period, a Monte Carlo approach is applied using the available
430 uncertainty estimates on instantaneous snowfall values. For each time and height point, the snowfall rate is assumed to obey a Gaussian distribution defined by the mean value and its standard deviation. By generating a large number (here, 1000) of random realizations sampled from these Gaussian distributions across the full time series, we build an ensemble of possible snowfall profiles. For each realization, the temporal mean snowfall profile is computed, resulting in a distribution of mean snowfall profiles that reflects the measurement uncertainty propagated through the averaging process. The final mean snowfall
435 estimate is obtained by averaging across all realizations, while the uncertainty on this mean is quantified as the standard deviation of these realizations. This approach explicitly accounts for the temporal variability and measurement uncertainty, providing a statistically robust estimate of the uncertainty on the mean snowfall radar data. Note that this method has been applied on the data by Grazioli et al. (2017a) only, as the uncertainties for Wiener et al. (2023) are not yet provided. To further



investigate the microphysical processes occurring in the precipitation evolution, we use a product derived from the MXPOL
440 radar scans. This product provides probabilities that a given process dominates the vertical evolution (growth or decay) of
snowfall. The classification relies on the PIVS (Process Identification based on Vertical gradient Signs) method developed by
Planat et al. (2021), which uses vertical gradients of radar reflectivity (Z_H) and differential reflectivity (Z_{DR}). Based on the
combined gradient signatures, the PIVS method categorizes hydrometeor growth into three main regimes: aggregation and
riming (AR), sublimation (SUB), and crystal growth by vapor deposition (CG). This information provides valuable constraints
445 for evaluating the occurrence of microphysical processes and to qualitatively assess those that dominate snowflakes' growth
and decay.

5.3 Model configuration and sensitivity experiments design

5.3.1 Regional model configuration

To carry out the evaluation, we ran Limited Area Model (LAM) simulations performed with the ICOLMDZ configuration of
450 LMDZ, consisting in the coupling between the DYNAMICO icosahedral dynamical core and the LMDZ physics. An Arctic
domain is set-up south of Svalbard (76°N, 29°E) and simulations span August 2022, corresponding to the RALI-THINICE
campaign. An Antarctic domain is centred on Dumont d'Urville station (66.66°S, 140.00°E) and simulations cover the period
from 20 to 31 December 2015, encompassing a precipitation event observed during the APRES3 campaign that already served
to assess the representation of snowfall in the Polar WRF model in (Vignon et al., 2019a). The two configurations share
455 the same model setup. The horizontal resolution is about 30 km, the model uses 95 hybrid-pressure vertical levels, with a
resolution ranging from approximately 150 m near 2 km altitude to 400 m near 7 km altitude. Lateral boundary conditions
are prescribed from hourly ERA5 reanalyses, while daily sea-surface temperature and sea-ice concentration are also taken
from ERA5. To constrain the large-scale dynamics and ensure concurrent 4-dimensional comparisons between simulations
and airborne observations, horizontal winds are nudged toward ERA5 within the domain, above 750 hPa using relaxation time
460 scale of 1 hour.

5.3.2 Physics package choice, calibration and sensitivity experiments design

Simulations are performed using a development version of the LMDZ physics package for CMIP7. This version is based on
the LMDZ6A version of the physics used for CMIP6 (Hourdin et al., 2020) and tuned for the 95-level vertical grid (Hourdin
et al., 2021) but includes several recent developments, particularly relevant for representing polar clouds and precipitation.
465 First, a new TKE-1 turbulent diffusion parameterization (Vignon et al., 2024) is used, as it exhibits better numerical properties
compared to the previous scheme of LMDZ. This scheme has been carefully calibrated for the representation of stable boundary
layers. Second, cloud phase partitioning has been substantially revised, introducing interactions between the cloud and turbulent
diffusion schemes and a coupling between microphysics and subgrid-scale dynamics (Raillard et al., 2026; Vignon et al., 2026).

The physical package used in this study is still under development and has not yet undergone a comprehensive tuning
470 exercise in the three-dimensional global version of LMDZ. Nevertheless, both the TKE and POPRECIP schemes have already



been extensively calibrated in single-column and three-dimensional configurations, with particular emphasis on warm-cloud cases (temperatures above 0°C). This previous calibration effort resulted in a consistent representation of warm clouds in single column simulations (see Appendix A). Relevant parameters for cloud-phase partitioning have been calibrated separately on the M-PACE single-column case (see Vignon et al. (2026))

475 As precipitation results from complex interactions between many physical processes, its parametric sensitivity necessarily depends on a large number of model parameters in different parameterizations. Assessing the parametric sensitivity of precipitation would require exploring a huge parameter space. We decide here to perform a non-exhaustive but simpler exercise, considering the sensitivity to a subset of POPRECIP parameters, keeping the parameters of the other parameterizations fixed from the previous calibration exercises.

480 We thus perform a Perturbed Parameter Ensemble (PPE) exercise with six parameters. Five adjustable parameters belongs to POPRECIP itself: $\tau_{snw,max}$, $q_{th,s}$, γ_{sub} , γ_{agg} and γ_{rim} (Table 1). These parameters respectively control key processes governing snow production and evolution, including autoconversion, sublimation, aggregation, freezing and riming. In addition, the parameter ζ which controls the width of the subgrid-scale total-water distribution is included in the PPE. Although not part of POPRECIP, ζ influences the amount of condensable water and exerts a strong control on precipitation formation.

485 We generate an ensemble of 60 – that is, 10 times the number of free parameters (Couvreur et al., 2021; Hourdin et al., 2021) – simulations with different parameter combinations. The parameter space is sampled using the HighTune explorer tool of Couvreur et al. (2021), which relies on Latin Hypercube Sampling to efficiently explore multidimensional parameter spaces.

The reference simulation, hereafter referred to as 'CTRL', uses the standard LMDZ precipitation scheme. Simulations including the POPRECIP scheme are referred to as POPRECIP. For illustration purposes, figures showing a single POPRECIP
490 simulation in fact illustrated the result from a randomly selected member of the ensemble.

6 Evaluation in regional simulations of Arctic and Antarctic case studies

6.1 Arctic case study

6.1.1 Illustration on Flight 40 of the RALI-THINICE campaign

The aim of this section is to qualitatively illustrate the differences between the CTRL and POPRECIP simulations and their
495 agreement with the observed IWC and LWC during a particular flight of the RALI-THINICE campaign: the Flight 40. This flight took place on 6 August 2022 and sampled a deep cloud and mid-level MPCs associated with the warm front of a moderate cyclone southeast of Svalbard (Rivière et al., 2024). The synoptic situation is well reproduced by LMDZ, with good agreement between the simulated and observed mid- and high-level cloud structures, allowing for a robust comparison of cloud properties (Raillard et al., 2024). Figures 5a–f present cross sections of the liquid fraction, LWC and IWC for a portion of the flight. The
500 LWC and IWC fields include both cloud condensates and precipitation, and in-cloud values are shown. Note that the considered segment of the flight trajectory (colored in Fig. 5i) sampled the same nimbostratus cloud at two different altitudes.



POPRECIP has a significant impact on the liquid fraction. The CTRL simulation produces clouds that are almost entirely liquid (Fig. 5b), whereas POPRECIP generates a more variable liquid fraction (Fig. 5a), with enhanced liquid content near cloud top (between 11.6 and 12 UTC) and a distinct liquid layer at higher altitudes. Visual inspection of every single PPE member reveals that this pattern is a robust feature in the ensemble. The simulated values of liquid percentage in clouds are overall in better agreement with the in-situ airborne observations. To understand this improvement, let us analyse the individual liquid and ice components. POPRECIP reduces the LWC throughout the cloud (Figure 5c–d). This is reflected in the mean liquid water path between 11.4 and 12.2 UTC, which decreases from 746 g m^{-2} in CTRL to 330 g m^{-2} on average across the POPRECIP ensemble. The illustrative POPRECIP member shown here has a mean LWP of 358 g m^{-2} , close to the ensemble mean. Where observations are available, the simulated values are generally closer to measurements than in CTRL, leading to an overall fair agreement. In contrast, the IWC distribution changes only modestly between CTRL and POPRECIP (Figs. 5e–f), with even higher IWC values in the core of the nimbostratus cloud in the POPRECIP simulation. Comparison with the radar-retrieved IWC (Figure 5h) shows that POPRECIP leads to slightly more concentrated values of ice water contents within the nimbostratus. However, both POPRECIP and CTRL overestimate the IWC. Overall, POPRECIP improves the agreement with observations in terms of LWC and liquid fraction in clouds, but the IWC remains overestimated.

Regarding the simulation of snowfall intensity, we now compare the simulated snow flux above the 0°C isotherm with the radar-retrieved snowfall flux immediately above the top of the detected melting layer (Fig. 5g). Both CTRL and POPRECIP reproduce a snowfall flux pattern within the deep nimbostratus that is broader than observed (regions of high precipitation values are broader than those of the observations). High snowfall values are thus not concentrated enough in space. This feature appears to be relatively insensitive to both the precipitation scheme and the choice of POPRECIP parameters, suggesting that its origin may instead be related to other aspects, such as the location of the cloud structures or the model's coarse resolution to represent such a cloud.

The new scheme offers the possibility to investigate the tendency associated with each of the microphysical processes, i.e. the amount of snow produced/depleted by unit of time (Fig. 6). The snow specific content (Fig. 6b) reveals the high snow content is concentrated in a quite narrow band in the deep nimbostratus. This snow is primarily generated by the autoconversion of ice crystals that grow by aggregation below (Figs. 6d,f). High snow contents in the lower but thick clouds between 11.5 and 12.1 are rather formed by riming and rain freezing (Figs. 6a,c), consistently with the presence of supercooled liquid water in those clouds.

6.1.2 Statistics on all flights and parametric sensitivity

To make the conclusions regarding the impact of POPRECIP on cloud water contents more robust, we now compare statistics across the 15 research flights of the campaign.

To this end, we construct scatter plots of simulated versus in-situ measured IWC and LWC for the POPRECIP and CTRL configurations. Two metrics are used to quantify model performance: the geometric mean bias, defined as the exponential of the mean logarithmic ratio between simulated and observed values and therefore interpreted as a multiplicative bias (values above or below unity indicating overestimation or underestimation, respectively), and the Root Mean Square Logarithmic Error

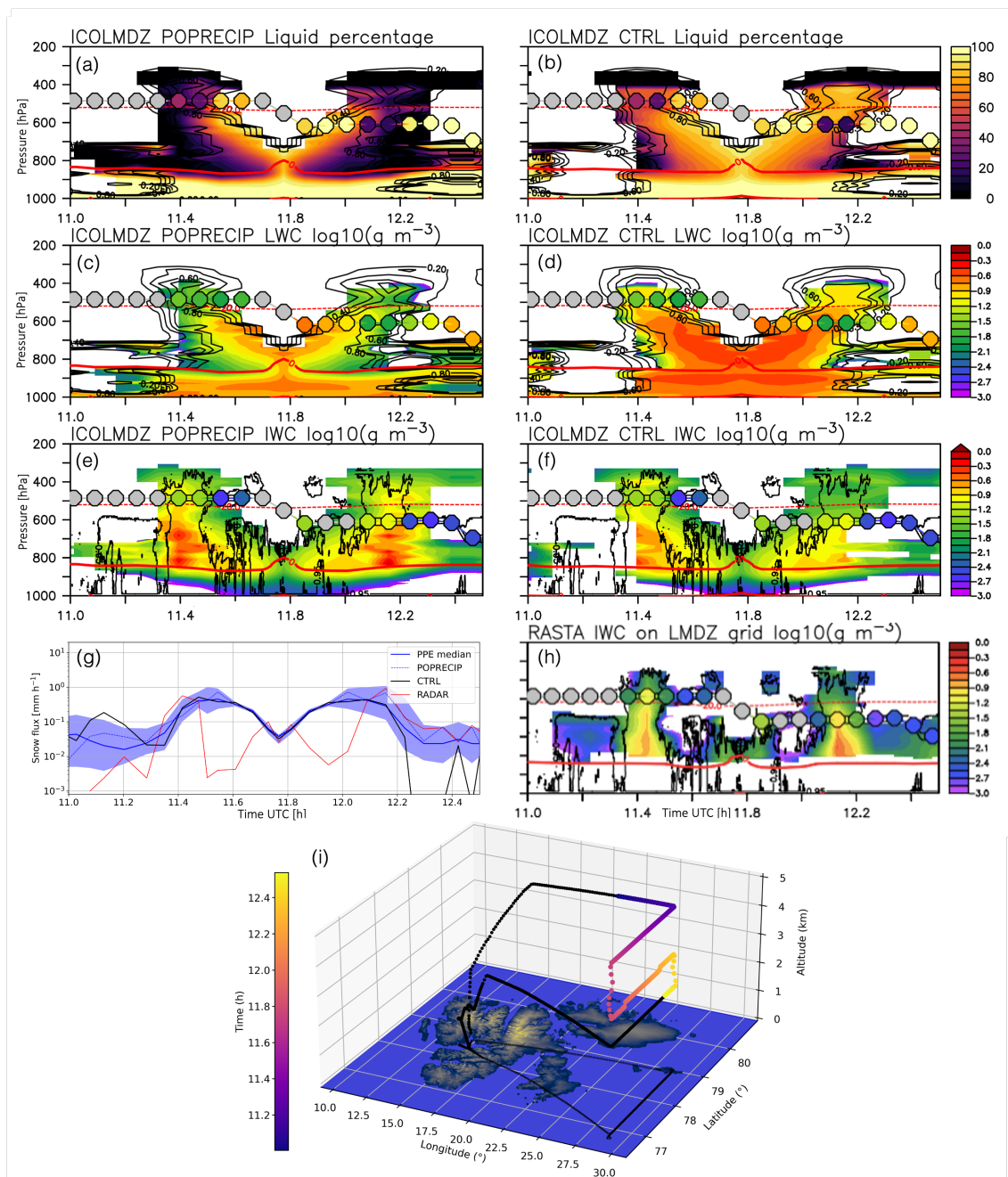


Figure 5. Time-pressure plots along flight path for simulated liquid percentage (a-b), simulated LWC c-d), simulated IWC (e-f), snowflux above melting layer for radar IWC retrieval (red curve), POPRECIP simulation (blue curve) and the 60 PPE simulations (blue envelope) (g), IWC radar retrieval, 3D trajectory of the flight 40 (i). For panels (a) to (f) : left panel shows the simulations with the POPRECIP scheme, right panel shows the control simulation and colored circles are the corresponding quantities estimated from the microphysical probes measurements. Black contours are simulated cloud fraction (a-d) and radar cloud mask (e-h). Red lines are isotherm 0°C (solid) and -20°C (dashed).

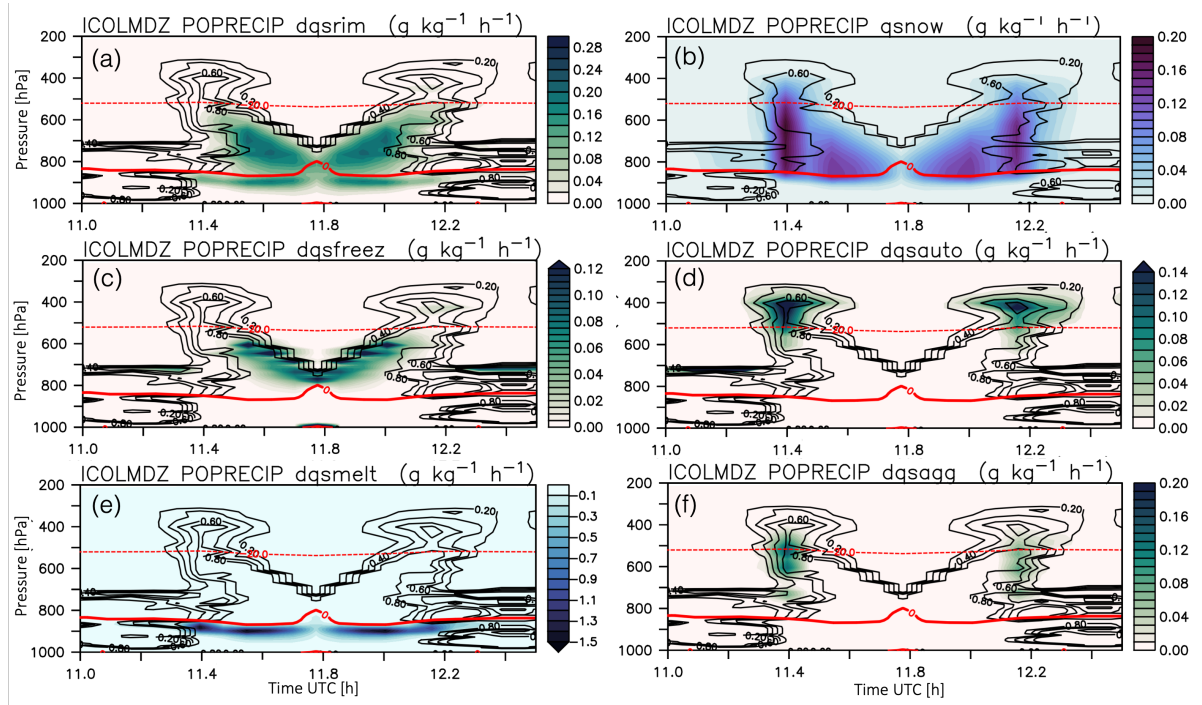


Figure 6. Time-pressure plots along flight path for different precipitation microphysical tendencies. a-d&f represents source terms for the snowflux: a) riming, b) snow content (mais pourquoi les unités sont les mêmes que pour a), c) freezing, d) autoconversion, f) aggregation while (e) is a loss term: melting. Black contours are simulated cloud fraction (a-d) and radar cloud mask (e-h). Red lines are isotherm 0°C (solid) and -20°C (dashed).

(RMSLE). Note that some liquid cloud points are not reproduced by the model (triangles on Fig. 7, correspond to a cloudy model grid cell but with negligible liquid content). Such points are excluded from the computation of these metrics.

As already noted for Flight 40, the simulated LWC differs substantially between CTRL and POPRECIP. The geometric bias of LWC decreases from 4 in CTRL to 1.9 in POPRECIP (Figs. 7a,c), indicating that the mean overestimation is reduced from approximately a factor of four to a factor of two. The LWC therefore exhibits a much weaker systematic bias, although the lowest observed values still tend to be overestimated.

In contrast, POPRECIP and CTRL exhibit very similar behaviour for IWC, with a nearly systematic overestimation of ice water content (Figs. 7b,d), particularly for low observed IWC values. The geometric bias and RMSLE remain very similar in the two simulations.

It is worth noting that the systematic overestimation of IWC and the decrease of the overestimation of the LWC also hold across the PPE ensemble. To assess this, we examined the sensitivity of the IWC and LWC RMSLE, as well as the corresponding geometric biases, to variations in the free parameters of the scheme. The only clear signal emerges for the LWC



geometric bias, which appears strongly constrained by the riming parameter γ_{rim} . No comparable sensitivity is found for the IWC (Figs. 7e,f).

550 6.2 Antarctic precipitation event

6.2.1 Case overview

We focus now on a precipitation event happening between 28 and 30 December 2015 at DDU. The synoptic conditions are described in details in Vignon et al. (2019a). The event was associated with the passage of a warm front of an extra-tropical cyclone located west of DDU and that advected heat and moisture towards the ice sheet along its eastern flank. The katabatic winds flowing down the slopes of the ice sheet were intensified due to the increase of the large-scale pressure gradient during the approach of the cyclone. As the katabatic boundary layer is relatively dry (Grazioli et al., 2017a), substantial partial or total snowfall sublimation at low levels occurred during the event.

DDU is located on the Antarctic coastline, near the junction of two grid cells in our model configuration. The closest grid cell contains both land-ice and ocean fractions and has a mean altitude of 275 m above sea level (asl), while the second closest grid cell is entirely oceanic and has a mean altitude of 20 m. As shown in Fig. 8, snowfall exhibits strong horizontal gradients along the Antarctic slope. We therefore chose the southern grid point, as it provides surface and boundary-layer conditions that are more representative of DDU, even though the incoming air masses have already begun to experience the orographic lifting associated with the coastal slope and the resulting precipitation depletion.

Figure 8 presents a latitudinal cross-section at the longitude of DDU at 00:00 UTC on 30 December, near the peak of the event. The wind vectors particularly evidence the katabatic flow descending from the Antarctic Plateau towards the coast. This katabatic flow transports dry continental air into the lowest kilometers of the atmosphere beneath the precipitating cloud layer and leading to precipitation sublimation. The figure also illustrates the strong horizontal variability in snowfall especially near the coastline where the maritime air mass encounters the katabatic outflow and where air masses are lifted up.

We first compare the time–height evolution of the precipitation flux with snowfall retrievals from the MRR at DDU (Grazioli et al. (2017a); Wiener et al. (2023)) (Fig. 9). The MRR observations show that snowfall often does not reach the ground due to low-level sublimation (Grazioli et al., 2017b; Jullien et al., 2020). In contrast, LMDZ tends to simulate precipitation fluxes that are generally more intense and more frequent, with precipitation reaching the surface more often than observed.

The surface snowfall rate (Fig. 9c) shows that both CTRL and POPRECIP capture the timing of the different precipitation peaks. The intensity of the precipitation peaks is reasonably represented at the beginning and end of the event, with a slight improvement in POPRECIP compared to CTRL. However, the main precipitation peak around 00:00 UTC on 30 December is strongly underestimated. The spread across the POPRECIP ensemble indicates that varying the sublimation parameter has only a limited impact on this behaviour, suggesting that it is more likely associated with the representation of the dry katabatic layer, an aspect that is beyond the scope of this study. Nevertheless, POPRECIP substantially improves the accumulated snowfall over the whole period between 28 Dec and 31 Dec (Fig. 9d), reducing the positive bias found in CTRL. The observed accumulated

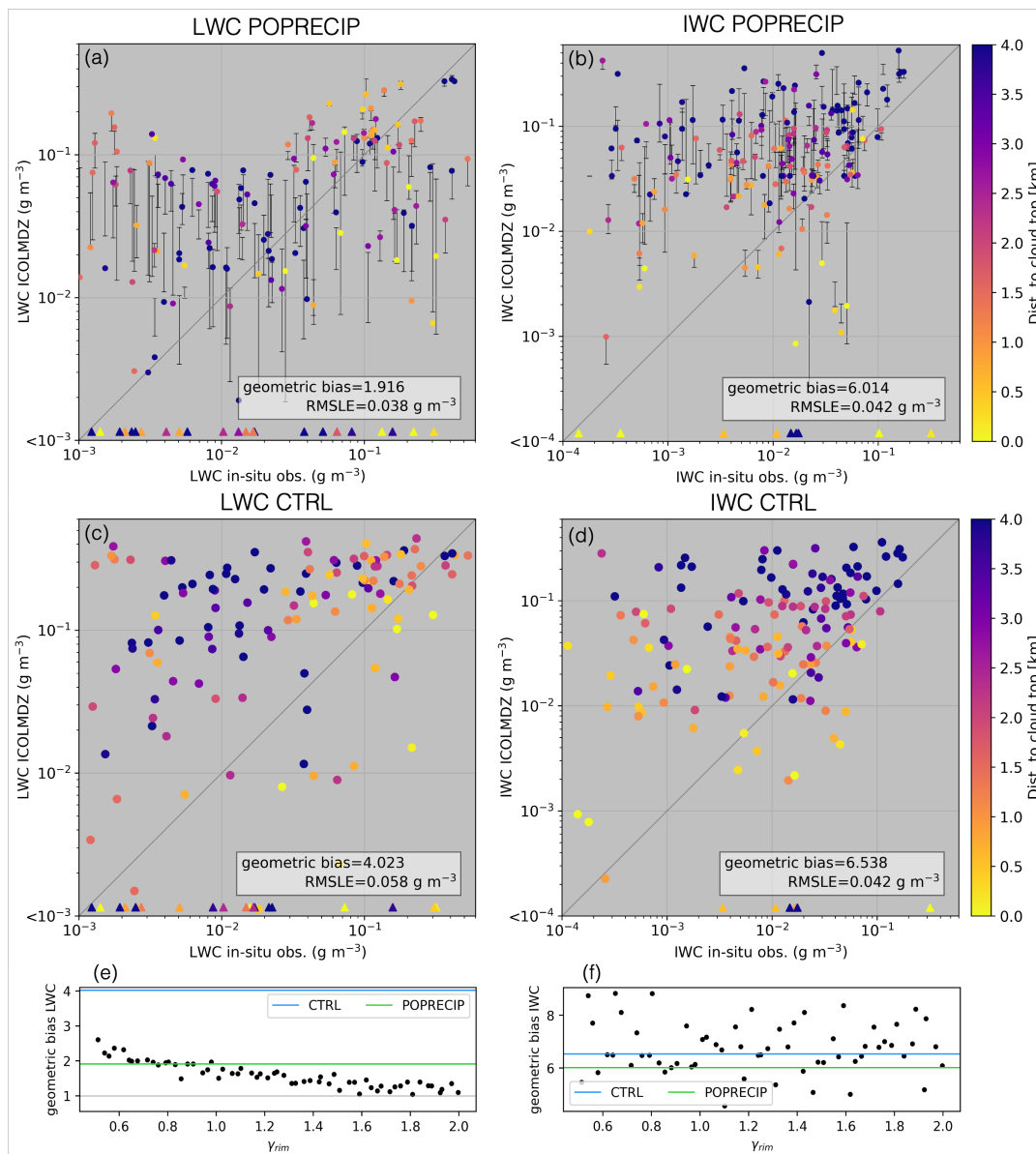


Figure 7. Scatter plots of in-situ observed vs. simulated liquid water content (left column) and ice water content (right column) using data from the 15 flights of RALI-THINICE. The POPRECIP simulation is shown on the top row and the control simulation on middle row. The color refers to the distance between the altitude of the model grid cell and cloud top in the model. Triangles correspond to points with a cloud fraction $\alpha_c > 0.1$ and IWC $< 10^{-4} \text{ g m}^{-3}$ (resp. LWC $< 10^{-3} \text{ g m}^{-3}$) in the model. They have been (artificially) plotted at the very bottom of the graph to make them visible. The error bars on panels a and b indicate the 25th and 75th percentiles of the PPE ensemble. The bottom row displays the evolution of the geometric bias of the LWC (e) et IWC (f) with the tuning parameter controlling riming. Details regarding the comparison methodology can be found in Raillard et al. (2024).

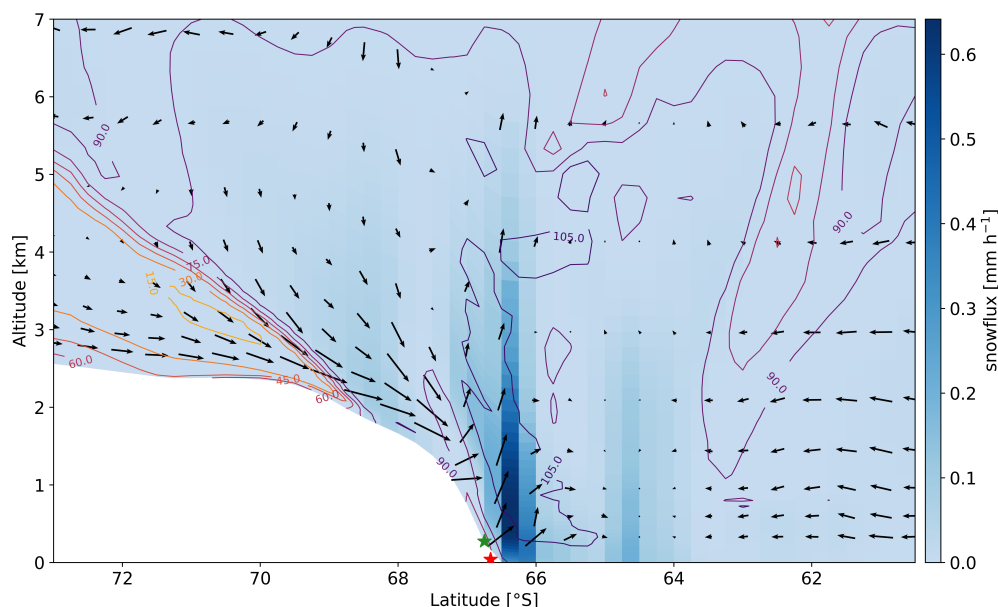


Figure 8. Latitudinal cross section of the precipitation flux at 00:00 UTC, 30 December and at the longitude of DDU (140.34° E). Relative humidity with respect to ice is shown in contours and the arrows represent the wind in the lat-z plane. Red and green stars correspond to the latitude-altitude coordinates of respectively DDU and its nearest grid point in the model.

580 snowfall is approximately 4 kg m^{-2} , whereas CTRL overestimates it by about a factor of two, while POPRECIP produces values that are much closer to the observations.

6.2.2 Microphysical processes and parametric sensitivity

Figure 10 presents the vertical structure of the precipitation event. Panel (a) shows an instantaneous radiosounding and model profile of relative humidity with respect to ice (RH_i) at 00:00 UTC on 30 December 2015. Figures 10b-d present vertical profiles
585 averaged over the duration of the event of snowfall fluxes, cloud water contents, and microphysical process tendencies from PPE and evidences key features during the precipitation event. First, the two MRR snowfall flux retrievals differ in magnitude but agree on the overall profile shape: the flux increases downward up to a maximum value, around 1 km agl and then decreases, as precipitation enters the drier katabatic layer and sublimates. Moreover, the value of the precipitation flux at the lowermost and highermost radar gates are similar between the two methods. Both CTRL and POPRECIP simulations produce snowfall
590 fluxes of the correct order of magnitude but overestimate the flux in altitude and near the surface. This overestimation is reduced in POPRECIP, which better matches the observed flux values at both ends of the profile, consistent with the improved precipitation representation shown in Fig 9. However, both simulations fail to reproduce the observed vertical structure of the snowfall flux. In particular, they lack sufficient growth near 1.0-1.5 km. The maxima of the profile is located lower in the column, around 500 m agl, and is less pronounced than in the observations. Furthermore, the decrease in flux towards

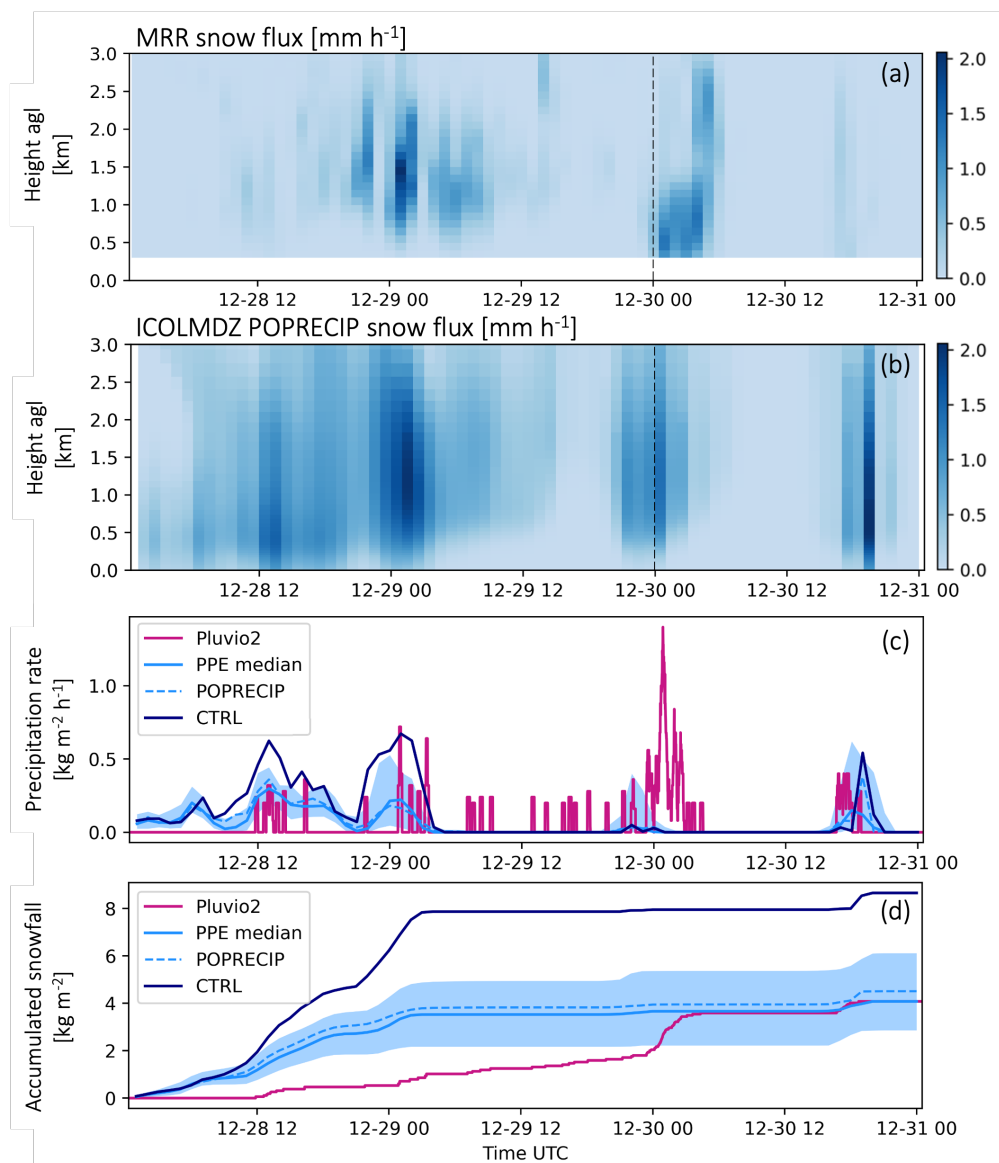


Figure 9. Precipitation flux time-height for the precipitation event considered for the MRR-Grazioli data (a) and LMDZ POPRECIP simulation (b). Panel (c) displays the surface snowfall measured by the Pluvio2 and simulated by the CTRL and POPRECIP simulations while panel (d) displays the accumulated snowfall during the event. In panels (c-d), the shaded envelope indicates the range of values across the PPE (see hereafter), the plain blue line the median value across the PPE and the dashed line the POPRECIP single member used for illustration on Figures 5 to 8. The vertical black dashed line corresponds to the time at which Figure 8 is done.



595 the surface is too weak. Caution is required when interpreting the upper part of the MRR profile, where precipitation growth is observed. First because the MRR sensitivity decreases with height, and second because the Z–S relationship, to retrieve the snowflux from the radar reflectivity, is calibrated near the surface, which may not hold aloft where ice crystal habits and velocities differ.

In the lower portion of the profile, the decrease in snowfall flux is primarily controlled not by the choice of parameters (see
600 the narrow envelop of blue profiles on Fig. 10) but by the vertical structure of the atmosphere, particularly the relative humidity profile and the representation of the katabatic layer in the model. The instantaneous RHi profiles (Fig. 10a) indicate that in both POPRECIP and CTRL simulations, the near-surface dry layer (up to about 1 km agl) is reasonably well represented, although it is slightly too shallow and too humid in the model. This likely contributes to the underestimated decrease in snowfall flux within this layer by limiting sublimation. Vignon et al. (2019b) showed that atmospheric models commonly underestimate the
605 boundary-layer depth above DDU, leading to a dry sublimation layer that is too shallow, particularly at coarse resolution. They attributed this bias to difficulties in representing the coastal transition of the katabatic flow, known as the katabatic jump. Similar low-level flow biases were also reported for LMDZ by Wiener et al. (2025) (see their Fig. 10). In the present case, the model reproduces the overall boundary-layer structure but exhibits a surface jet that is too strong and confined (not shown), together with a dry layer that appears slightly too shallow. Whether these biases are directly related to an inadequate representation of
610 the katabatic jump remains unclear.

The overly smooth shape of the snowfall flux profile, characterized by gradients that are too weak in both the growth and decay regions, is a robust bias across the PPE and therefore suggests that additional deficiencies exist and that work must be done to correct this. One possible explanation could be that collection processes, such as aggregation and riming, are underestimated because the current microphysical scheme assumes a constant snow particle radius, thereby limiting the
615 variability of collection efficiencies. Another possible avenue is to account for the effect of turbulence on collection processes, as proposed by Chellini and Kneifel (2024), which may be particularly relevant within the strongly sheared katabatic boundary layer. Further investigation of these processes represents a promising direction for future model developments that could help improve the vertical evolution of snowfall in LMDZ.

Regarding condensates (Fig. 10c), the POPRECIP simulation exhibits a clear reduction in LWC and a slight increase in IWC
620 relative to CTRL, in accordance with the observations made for the RALI-THINICE analysis. This behavior can be explained by the inclusion of the riming process and reduction of the ice crystal autoconversion (see Fig. 10d, blue shading). More complex feedbacks are also likely involved: greater ice content implies a higher surface area, which enhances vapor deposition onto ice crystals. This, in turn, reduces the liquid fraction in the phase partitioning parameterization (Raillard et al., 2024).

Although both simulations exhibit quantitative biases in snowfall flux, an important question is whether they nevertheless
625 succeed in capturing the microphysical processes responsible for precipitation evolution throughout the atmospheric column. Based on the MXPOL product from Planat et al. (2021), which provides the probability that a given microphysical process dominates the vertical evolution of snowfall, three distinct layers are identified, each characterized by a different dominant microphysical process. In the observations, the upper layer is primarily associated with crystal growth by vapor deposition, the middle layer with aggregation, and the lower layer with riming. Based on these findings, three corresponding vertical layers



are defined (shaded regions in Fig. 11d, with the colors indicating the dominant process identified from the observations). For each of the 60 PPE members the dominant microphysical tendency is then identified within each layer. Note that snow autoconversion in the model is considered to represent crystal growth by vapor deposition. The dominant processes are then related to the perturbed parameter values to assess whether clear dependencies emerge.

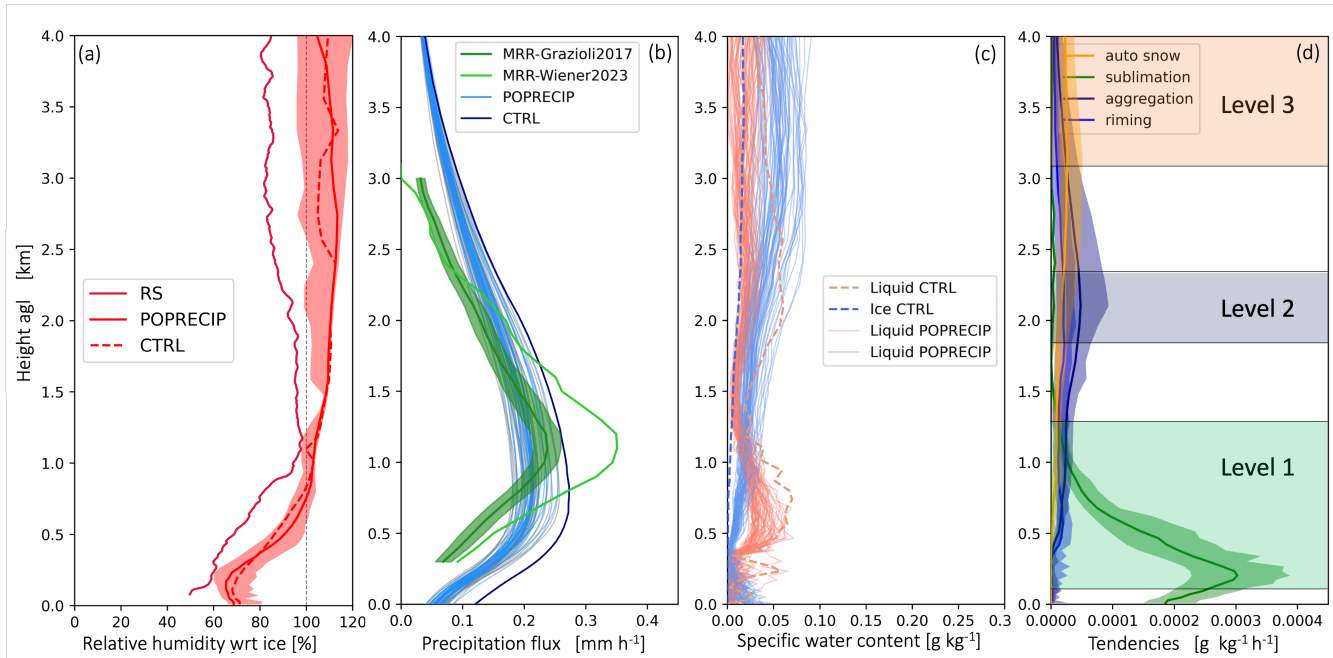


Figure 10. (a) Instantaneous vertical profile of relative humidity with respect to ice at the Dumont d’Urville radiosonde launch at 00:00 UTC on 30 December 2015. The radiosonde profile is shown in dark red, the POPRECIP simulation in solid red, and CTRL in dashed red. The shaded envelope represents the range of relative humidity across the POPRECIP PPE. Event-mean vertical profiles of snowfall fluxes (b), in-cloud liquid and ice water contents (c), and snow microphysical process tendencies for POPRECIP (d). Shaded rectangles highlight regions where observations suggest that the dominant microphysical process is vapor deposition growth (orange), riming and aggregation (blue), or sublimation (green). Note that the snow sublimation tendency is plotted with a positive sign (i.e., multiplied by -1). In panels (b) and (c), individual ensemble members are shown as colored lines. In panel (a) and (d), a shaded envelope indicates the range of values across the PPE for readability. The dark green shading in (b) represents the uncertainty on the MRR mean snowfall profile, estimated using a Monte Carlo method described in Section 5.2. In panel (b), blue lines correspond to members that correctly reproduce the dominant observed microphysical process at the correct altitude, whereas grey lines indicate those that do not.

While sublimation (resp. aggregation/riming) is always the dominant process in the model in the lowermost layer level 1 (resp. middle layer level 2) regardless of the parameters values (Fig. 11), the dominant process at the uppermost layer level 3 depends on the set of parameters chosen. In particular, autoconversion by vapor deposition tends to prevail at low values of γ_{agg} and $\tau_{snw,max}$.



More precisely, by plotting in the 2D $\gamma_{agg}-\tau_{snw,max}$ space (Fig. 11g) we see that the couple $(\gamma_{agg}, \tau_{snw,max})$ controls snowfall growth at the uppermost layer level 3. Such an analysis can thus help constraint the possible values of those two parameters to guarantee that snowfall grow mostly by autoconversion in the uppermost layer.

Note that in Fig. 10b, blue curves indicate simulations that correctly capture the dominant process at the appropriate altitude, while grey curves denote mismatches. The lack of visible clustering among blue profiles suggests that correct process representation is not tightly that important to represent the magnitude of the snowfall flux.

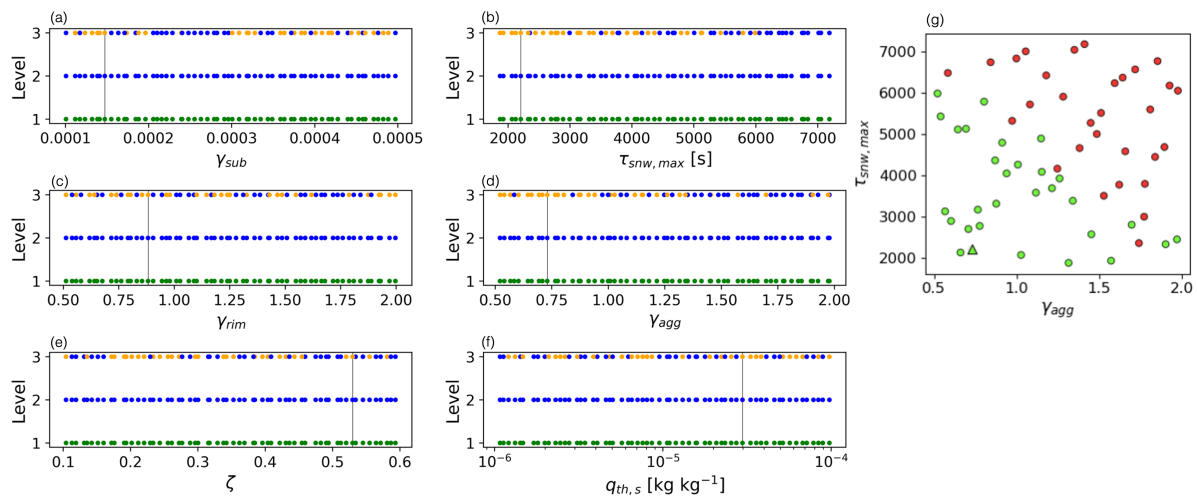


Figure 11. Left : the color dots indicate the dominant processes at each level for each of the 60 members of the PPE simulations, as a function of the different adjustable parameters. Orange denotes the crystal growth by vapor deposition or ice autoconversion in the model, blue denotes the aggregation and riming processes and green the sublimation. The black contoured triangle represent the PPE member chosen for illustration all along the analysis. Right (g) Scatter plot of the 60 members as a function of the two parameters γ_{rim} and $\tau_{snw,max}$. Green points indicate simulations with that captured the right processes at the right altitude.

7 Conclusions and discussions

This paper presented the development of the new precipitation parameterization POPRECIP in the LMDZ atmospheric general circulation model. The representation of precipitation had previously been identified as a major weakness of the model, with consequences for both surface precipitation and cloud properties, particularly in ice and mixed-phase cloud regimes. Addressing these limitations motivated the development of POPRECIP.

The new scheme introduces a more advanced representation of precipitation formation by explicitly accounting for key microphysical processes, including aggregation and riming, while also incorporating important macrophysical aspects such as subgrid considerations to account for cloud–precipitation interactions. In parallel, the numerical treatment of several existing processes was revised to improve robustness, convergence, and stability at the long time steps characteristic of climate models.



Importantly, POPRECIP provides a framework for process-oriented studies, enabling a more detailed understanding of the physical processes governing precipitation formation that could not be investigated with the previous precipitation parameteri-
655 zation.

While it does not deteriorate but also not substantially improve the representation of weakly-precipitating warm boundary layer clouds, it exhibits a more physical behaviour in single-column simulations of Arctic cold-air outbreak clouds.

The parameterization was then tested and evaluated in regional simulations of two polar case studies: an Arctic case based on airborne observations collected during the RALI-THINICE campaign, and an Antarctic case using observations from the
660 APRES3 campaign in coastal Adélie Land. For each case, PPE was conducted to assess the sensitivity of the scheme to its free parameters.

The Arctic case shows that the new parameterization reduces the overestimation of cloud LWC while having little impact on the IWC. The sensitivity analysis further demonstrates that the simulated LWC is strongly controlled by the free parameter governing the strength of riming.

665 The application to the Antarctic case highlights that the growth of precipitation remains challenging to represent. Nevertheless, in selected parametric configurations the model is able to capture the dominant microphysical processes responsible for precipitation growth and their vertical distribution. The retrieved vertical distribution of dominant microphysical processes provides a valuable constraint on the autoconversion and aggregation parameters, allowing the realism of the simulated precipitation growth pathways to be assessed.

670 Finally, the implications of POPRECIP extend well beyond polar regions. As LMDZ is a global climate model, the new precipitation parameterization affects precipitation formation across all cloud regimes. Preliminary tests indicate that introducing POPRECIP does not degrade model performance in warm-cloud cases. However, its impact on the global distribution of precipitation in a globally calibrated version of LMDZ has not yet been comprehensively evaluated and remains an important avenue for future work. Beyond this, several aspects of the precipitation scheme deserve further refinement. In particular, the
675 assumption of a constant snow particle radius is likely too simplistic and should must be revisited. The evaluation against observations also highlights remaining deficiencies, including insufficiently peaked snowfall flux profiles over Dumont d'Urville and a tendency to maintain excessive ice water contents during the RALI-THINICE cases. Addressing these limitations should further improve the representation of snowfall and cloud microphysical processes in LMDZ.

Code availability. The latest version of the LMDZ source code can be downloaded freely from the LMDZ website. The version used for
680 the specific simulation runs for this paper is the svn release 5727 from June, 27th 2025, which can be downloaded and installed on a Linux computer by running the `install_lmdz.sh` script available here: http://www.lmd.jussieu.fr/~lmdz/pub:/install_lmdz.sh.

Data availability. The LMDZ and DYNAMICO models are freely distributed under CeCILL licence at the following links <https://web.lmd.jussieu.fr/~lmdz/pub/> and <https://gitlab.in2p3.fr/ipsl/projets/dynamico/dynamico>. RALI-THINICE data are freely distributed on the



campaign website <https://ralithinice.aeris-data.fr/>. The APRES3 campaign data have been distributed on the PANGAEA data repository
685 (<https://doi.org/10.1594/PANGAEA.883562>). A description of the dataset is provided in Genthon et al. (2018).

Appendix A: Effect of the POPRECIP scheme on single column simulations of typical weakly precipitating warm boundary-layer clouds

The impact of activating the POPRECIP scheme on the representation of warm boundary-layer clouds is assessed using single-column simulations of three typical cases thoroughly described by Hourdin et al. (2019). The ARMCU case is derived from
690 observations collected on 21 June 1997 at the Atmospheric Radiation Measurement (ARM) site in Oklahoma, USA (Brown et al., 2002). It represents an idealized diurnal cycle of shallow convection over land, characterized by well-developed fair-weather cumulus clouds. The RICO (Rain in Cumulus over the Ocean; vanZanten et al., 2011) case focuses on precipitating trade-wind shallow cumulus. The SANDU case is a composite stratocumulus-to-cumulus transition case described by Sandu and Stevens (2011). It was constructed by compositing the large-scale conditions sampled along a set of individual three-day
695 Lagrangian trajectories over the northeastern Pacific during summer. The stratocumulus deck exhibits a pronounced diurnal cycle and begins to break up during the second day as the boundary layer deepens. The objective of this analysis is to determine whether activating the POPRECIP scheme degrades the representation of these three cases relative to the LMDZ configuration of Hourdin et al. (2021). This configuration corresponds to the LMDZ6A physics package retuned using a semi-automatic History Matching procedure based on a set of metrics (that is, tuning targets) on the three single-column cases complemented
700 with global radiative metrics. We perform a tuning exercise similar to that of Hourdin et al. (2021), but replacing the standard precipitation scheme with POPRECIP. The same set of tuning parameters is considered (see their Table 1), with the addition of the collection parameter γ_{col} . The simulation results are shown in Figure A1, which presents the vertical profiles of cloud fraction during the mature cloud stage for the Hourdin et al. (2021) LMDZ configuration, the reference Large-Eddy Simulations (see Hourdin et al., 2021 for details), and the simulations using the POPRECIP scheme with the ten ‘best’ parameteric
705 configurations identified by the tuning procedure. Overall, the figure indicates that the representation of warm boundary-layer clouds is very similar with and without POPRECIP. The activation of POPRECIP neither degrades nor substantially improves the simulated cloud structure. In particular, it does not alleviate the persistent underestimation of cloud fraction in the upper part of the cloud layer in the RICO case. Further evaluation using warm-cloud cases with stronger precipitation, such as congestus clouds, would be valuable, but is beyond the scope of the present study.

710 Appendix A: Summary of variables and symbols

Appendix: Symbols and variables used in the POPRECIP scheme description

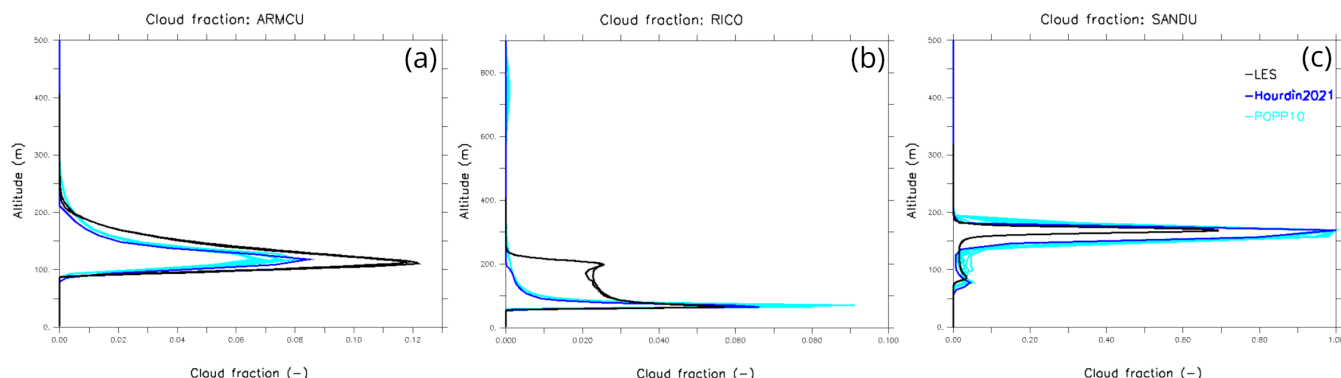


Figure A1. Vertical profiles of the cloud fraction on the ARM, RICO and SANDU single column model cases. Blue lines refer to a control LMDZ simulation with the LMDZ6A physics package – that is, why the standard precipitation scheme – adapted to a new 95 level vertical grid and calibrated in Hourdin et al. (2021). Cyan lines show the 10 ‘best’ LMDZ simulation with the POPPRECIP scheme after a full re-tuning of the model following Hourdin et al. (2021). Black lines show outputs from reference Large Eddy Simulations (see Hourdin et al., 2019).

Symbol	Description	Unit
c_p, c_{pd}, c_{pv}	Specific heat capacity of moist/dry air, water vapor	$[J\ kg^{-2}\ K^{-1}]$
$E_{rl}, E_{si}, E_{sl}, E_{ri}$	Rain–liquid, snow–ice, snow–liquid and rain–ice collection efficiencies	$[-]$
f_v	Ventilation factor	$[-]$
F_{min}	Threshold precipitation flux below which the clear-sky precipitation fraction is reduced	$[kg\ m^{-2}\ s^{-1}]$
F_r, F_s	rain, snow precipitation flux (clr/cld and grid-box-mean variants use the same notation)	$[kg\ m^{-2}\ s^{-1}]$
k	Vertical model level index	$[-]$
K_a	Thermal conductivity of air	$[J\ (mK)^{-1}]$
L_f	Latent heat of fusion	$[J\ kg^{-1}]$
m_s	Mass of one snowflake	$[kg]$
N_r, N_i, N_s	Number of raindrops, ice crystals, snowflakes	$[\#]$
p	Air pressure	$[Pa]$

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Symbol	Description	Unit
q_v, q_l, q_i	Specific content of water vapour, liquid, ice	[kg kg ⁻¹]
q_r, q_s	Specific content of rain, snow	[kg kg ⁻¹]
q_{tot}	Total specific humidity of the grid cell	[kg kg ⁻¹]
q_{sl}, q_{si}	Saturation specific humidity wrt liquid, ice	[kg kg ⁻¹]
$q_{th,r}, q_{th,s}$	Liquid/ice content threshold triggering rain/snow autoconversion	[kg kg ⁻¹]
$\Delta q_{s,max}, \Delta q_{r,max}$	Maximum snow/rain quantity that can melt/freeze	[kg kg ⁻¹]
r_r, r_s, r_i	Rain drop, snowflake, ice crystal radius	[m]
$r_{eff,r}, r_{eff,s}$	Effective radius of rain drops and snowflakes (constant)	[m]
T	Air temperature	[K]
T_{hom}	Homogeneous-freezing temperature (235.15)	[K]
T_{wi}	Ice wet-bulb temperature	[K]
T_0	Freezing point (273.15)	[K]
$\mathbf{u}$	Horizontal wind vector	[m s ⁻¹]
v_s	Snowflake fall speed	[m s ⁻¹]
w_i	Ice crystal fall velocity	[m s ⁻¹]
x_{liq}	Liquid fraction of the cloud condensate	[-]
X^{clr}, X^{cld}	Variables in clear sky, cloud	[-]
z	Height	[m]
Z_H, Z_{DR}	Radar reflectivity, differential reflectivity	[dBZ]
α_c	Cloud fraction	[-]
α_{pr}	Total precipitation fraction	[-]
$\alpha_{pr}^{clr}, \alpha_{pr}^{cld}$	Clear-sky, cloud precipitation fraction	[-]
Δz	Model layer thickness	[m]

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Symbol	Description	Unit
$\gamma_Y,$	Tuning coefficients controlling the strength of each POPRECIP process (see Table 1)	Varies
$\kappa_{eva}, \kappa_{sub}$	Exponents in the evaporation/sublimation equations	[-]
κ_{τ}	Exponent for snow autoconversion timescale	[-]
ζ	Tuning parameter for the width of total water PDF	[-]
$\rho, \rho_r, \rho_s, \rho_i$	Air, rain, snow and ice density	[kg m ⁻³]
$\tau_{rain}, \tau_{snw}, \tau_{snw,min}, \tau_{snw,max}$	Rain and snow autoconversion timescales	[s]
$\tau_{fr,imm}$	Immersion freezing timescale	[s]

Author contributions. LR ran the simulations, designed the evaluation, and wrote the first version of the manuscript. LR and EV performed the analysis. AB and EV extensively tested and debugged the parameterization and contributed to the development of LMDZ cloud scheme. LR, AB, EV, GR, MW, and ND jointly designed, developed, and implemented the parameterization, and contributed to reviewing and editing the manuscript. IM performed the warm case tuning of the LMDZ model.

Competing interests. The authors declare they have no competing interests.

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