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Use of A-train satellite observations (CALIPSO-PARASOL) to evaluate tropical cloud properties in the LMDZ5 GCM

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6 Abstract

7 The evaluation of key cloud properties such as cloud cover, vertical profile and optical depth as well as the analysis of their
8 intercorrelation lead to greater confidence in climate change projections. In addition, the use of collocated and instantaneous
9 data facilitates the links between observations and parameterizations of clouds in climate models.

10 New space-borne multi-instruments observations collected with the A-train make simultaneous and independent
11 observations of the cloud cover and its three-dimensional structure at high spatial and temporal resolutions possible. The
12 CALIPSO cloud cover and vertical structure and the PARASOL visible directional reflectance, which is a surrogate for the
13 cloud optical depth, are used to evaluate the representation of cloudiness in two versions of the atmospheric component of
14 the IPSL-CM5 climate model, LMDZ5 GCM. A model to satellite approach in applying the CFMIP Observation Simulation
15 Package (COSP) is used to allow a quantitative comparison between model results and observations.

16 The representation of clouds in the two model versions is first evaluated using monthly mean data. This classical approach
17 reveals model biases of different magnitudes depending on the model version used. These biases are an underestimation of
18 cloud cover associated to an overestimation of cloud optical depth, an underestimation of low- and mid-level tropical clouds
19 and an overestimation of high clouds. The difference between models of these biases clearly highlights the improvement of
20 the amount of boundary layer clouds, the improvement of the properties of high-level clouds and the improvement in
21 simulating mid-level clouds in the tropics thanks to the new convective, boundary layer, and cloud parameterizations
22 included in LMDZ5B compared to the previous LMDZ5A version. The correlation between instantaneous cloud properties
23 allows for a process-oriented evaluation for tropical oceanic clouds. This evaluation shows that the cloud population with
24 intermediate values of cloud cover and cloud reflectance when using monthly mean is now split in two groups of clouds,
25 one with low and intermediate values of the cloud cover and another one with cloud cover close to one. The precise
26 determination of cloud height allows us to focus on specific types of clouds (i.e. boundary layer clouds, high clouds, low-
27 level clouds with no clouds above). For low-level clouds over the tropical oceans, the correlation between instantaneous
28 values of the cloud cover and cloud reflectance reveals a major bias in the simulated liquid water content for both model
29 versions. The origin of this problem is determined and possible improvements such as considering the sub-grid
30 heterogeneity of cloud properties are investigated using sensitivity tests. In summary, the analysis of the relationship
31 between different instantaneous and collocated variables allows for process-oriented evaluations. These in turn may help to
32 improve model parameterizations and may also help to bridge the gap between model evaluation and model development.

33 1. Introduction

34

35 The evaluation of clouds simulated by general circulation models (GCMs) generally relies on monthly mean values [e.g. Yu
36 et al., 1996, Webb et al. 2001, Zhang et al. 2005], which do not contain detailed information on the transient aspects of
37 cloud behavior. Using monthly mean values might lead to a quantitative way to calibrate models in order to match present-
38 day observations with present-day simulations but these monthly means lack details on the cloud processes that create them.

39

40 The A-train jointly observes the cloud radiative properties using the passive remote sensors PARASOL (Polarization &
41 Anisotropy of Reflectances for Atmospheric Sciences coupled with Observations from a Lidar) [e.g. Parol et al. , 2004] and
42 CERES (Clouds and the Earth's Radiant Energy System) [Wielicki et al., 1996] as well as the cloud vertical structure using
43 the new generation of satellites carrying lidar instruments CALIOP/CALIPSO (Cloud-Aerosol Lidar and Infra-red
44 Pathfinder Satellite Observations) [Winker et al. 2007]. This joint observation dataset constitutes a unique opportunity to
45 perform quantitative evaluations of GCM cloudiness [Chepfer et al., 2008, Marchand et al., 2009, Zhang et al., 2010,
46 Cesana and Chepfer, 2013]. The ability of the A-train to simultaneously observe the radiative properties of clouds and their
47 three-dimensional distribution at the instantaneous time scale (typically on the order of tens of seconds) and at high spatial
48 resolution [Konsta et al., 2012] is used to provide new observational constraints to evaluate the representation of cloud
49 processes in climate models.

50

51 A major aim of physical parameterizations in GCMs is to reproduce the mean properties of cloud variables as well as the
52 relationships between these cloud variables and the dynamic and thermodynamic state of the atmosphere. If models fail to
53 reproduce these key features, they may lack the ability to properly predict cloud variations under environmental changes,
54 cloud feedbacks and the cloud response to anthropogenic forcing. A key step is thus to evaluate the mean properties of
55 clouds and to evaluate how they vary in response to a change of the physical characteristics of their environment. The
56 instantaneous analysis of cloud properties facilitates a direct relationship between observations and model
57 parameterizations. Some meteorological features occur at instantaneous time scale but are lacking when using monthly
58 mean values, such as the complex multi-layer structure of mesoscale convective systems or the inhomogeneous spatial
59 structure of marine stratocumulus. These features have an important impact on the radiative effect of clouds and therefore
60 on the magnitude of cloud feedbacks as the two are strongly correlated [Brient and Bony, 2012].

61

62 Short time scales may be considered during the development stage of a model. For instance, some recent developments in

the LMDZ atmospheric model have been undertaken in order to improve the diurnal cycle of precipitation and to solve the long-standing problem of too early precipitations in tropical regions [Rio et al., 2013]. The following questions are addressed:

- 1) Have the recent developments in the LMDZ5B atmospheric model improved the representation of clouds' physical properties?
- 2) Can the instantaneous spatial observations allow for the identification of remaining problems in the model representation of clouds? and
- 3) What role plays the cloud heterogeneities at the model's subgrid scale for the remaining problems in representing cloud in models?

The simulated cloud occurrence, vertical structure and optical depth of tropical oceanic clouds are compared with A-train observations. Monthly mean values are first considered as it is the common time scale used in climate model evaluation. Instantaneous values are then used to facilitate the interpretation of results in term of model parameterization because the statistical relationships between cloud variables at these two time scales have been shown to be significantly different [Konsta et al., 2012]. Observations provided by CERES are used to evaluate the cloud shortwave radiative albedo, observations provided by PARASOL are used to evaluate the cloud optical depth, and observations provided by CALIOP/CALIPSO are used to evaluate the cloud cover and the cloud vertical profile.

The two versions of the LMDZ5 atmospheric model, the A-train observations and the observations simulators are briefly described in Section 2. The cloud properties simulated by the model are first evaluated using monthly mean observations of top-of-the-atmosphere fluxes, reflectance, cloud cover, and vertical structure (Section 3). A more advanced process-oriented evaluation is then conducted for tropical oceanic clouds based on the correlation between instantaneous cloud properties observed with the A-train. Illustrations of how this may help to improve model parameterizations are presented in Section 4. Conclusions are given in Section 5.

2. Methodology

2.1 The LMDZ5 climate model

LMDZ5 is the atmospheric component of the IPSL-CM5 climate model [Dufresne et al., 2013]. Two versions of this atmospheric model (LMDZ5A [Hourdin et al., 2013-a] and LMDZ5B [Hourdin et al., 2013-b]) are evaluated. Both model versions are described in cited papers and only key aspects are summarized here. LMDZ5A is similar to LMDZ4 [Hourdin

et al., 2006] used in the previous version of the IPSL model [Marti et al., 2010] but it has an increased vertical resolution (from 19 to 39 vertical levels), an improved representation of the stratosphere and a modified horizontal grid (1.895° in latitude $\times$ 3.75° in longitude). LMDZ5B is a new model version that includes, in addition to LMDZ5A, many new developments on the physical parameterizations such as (i) a new scheme of the boundary layer, which combines a model of turbulent diffusion and a 'mass flux' scheme in order to represent the coherent structures of the dry or cloudy convective boundary layer [Rio and Hourdin, 2008, Rio et al., 2010] and a new low-level cloud scheme [Jam et al., 2011], (ii) a parameterization of the cold pools created by re-evaporation of convective rainfall [Grandpeix and Lafore, 2010], and (iii) a modification of the triggering and the closure of the Emanuel (1991) convective scheme based on the Available Lifting Energy for the triggering and Available Lifting Power for the closure [Rio et al., 2013]. All model results have been obtained with multi-years simulations over the period 1979-2009 in which the sea surface temperature and the sea-ice cover are prescribed to values close to observations (AMIP experiments).

104

2.2. Observations

106

Cloud cover and cloud vertical structure

108

Thanks to its vertically resolved measurements, to its high sensitivity to optically thin atmospheric layers and to its high horizontal resolution, the CALIOP/CALIPSO lidar [Winker et al. 2007] is well suited to accurately identify clear sky areas, aerosol regions [Liu et al. 2008, Vuolo et al. 2009], fractionated cloud covers such as the trade winds cumulus [Konsta et al., 2012, Medeiros et al., 2010], optically thin clouds such as the sub-visible cirrus clouds [Sassen et al. 2008, Noel et al. 2010, Martins et al. 2011], polar stratospheric clouds [Noel et al. 2008, Pitts et al. 2007, Noel et al. 2010], and to document the cloud vertical distribution of the atmosphere.

115

In order to compare the GCM results with CALIOP observations, a dedicated product called CALIPSO-GOCCP has been developed to be fully consistent with the lidar simulator [Chepfer et al. 2010]. This product consists in applying Scattering Ratio (SR) thresholds values to the 532nm lidar SR signal to detect the presence of clouds. The cloud detection (0 or 1) is done at the original horizontal Level 1 CALIOP resolution (330m along track and 75m cross-track of the satellite orbit) but on a lower vertical resolution (40 equidistant vertical levels of 480 m height). The cloud fraction is then built on a $2^\circ \times 2^\circ$ latitude/longitude grid to provide information on cloud cover, on low- mid- and high-levels cloud covers, and on vertical cloud distributions.

123

124 *The monodirectional cloud reflectance: a surrogate for the cloud optical depth*

125

126 The PARASOL instrument (POLDER-like, [Deschamps et al. 1994]) has a multi-viewing angle capability allowing for the
127 estimation of the instantaneous monodirectional reflectance of clouds. The use of this level-1 product as a surrogate of the
128 optical depth eliminates the need for many of the assumptions made during the retrieval process of the cloud optical
129 thickness [Minnis et al., 1995]. The criteria used for the selection of the viewing angle are described below. Note that Cole
130 et al. [2011] have computed and used CERES SW fluxes integrated over the diurnal cycle to perform a similar analysis but
131 the PARASOL instantaneous monodirectional reflectance provides more precise information because it contains fewer
132 assumptions than the CERES daily SW fluxes.

133

134 Above the ocean surface, the visible directional reflectance is defined as $\rho(\theta_s, \theta_v, \phi_s, \phi_v) = \pi L(\theta_s, \theta_v, \phi_s, \phi_v) / E_s \mu_s$, where
135 $\theta_s, \theta_v, \phi_s, \phi_v$, are respectively the solar zenith angle, the zenithal viewing angle, the solar and viewing azimuth angles, L is
136 the measured radiance, μ_s is the cosine of the solar zenith angle, and E_s the incident solar radiation. This directional
137 reflectance is mostly sensitive to the solar zenith angle (θ_s), to the viewing direction ($\theta_v, \phi_s - \phi_v$) and to the cloud optical
138 depth. Some viewing directions are contaminated by the specular reflection of the solar light on the sea surface (i.e.
139 sunglint), or are very dependent on cloud microphysical properties (e.g. particle shape through their optical properties). As
140 the aim is to evaluate the optical depth, the reflectance observed in a single viewing direction that is mostly sensitive to the
141 cloud optical depth and less to other parameters has been selected. The reflectance in a constant single direction has thus
142 been selected all over the globe (Fig. A) in avoiding (i) directions with $(90^\circ < \phi_s - \phi_v < 270^\circ)$ since they are sensitive to the
143 glitter reflection, (ii) the backscattering direction, which is highly sensitive to the cloud microphysical properties, and (iii)
144 the nadir direction, which is less sensitive to the optical depth than any other direction. Among the others possible directions,
145 the one at 865nm, which is the more frequently observed by PARASOL all over the globe ($\theta_v = 27^\circ \pm 2.5^\circ, \phi_s - \phi_v = 320^\circ \pm 2.5^\circ$)
146 was selected. All the directional reflectance values of spatial resolution of $6 \times 6 \text{ km}^2$ measured by PARASOL in this direction
147 are then projected onto a $2^\circ \times 2^\circ$ grid. The calibration of PARASOL is described by Fougnie et al. [2007]. The calibration
148 accuracy is within 1.5% for the 865nm channel. The spatio-temporal sampling of PARASOL and CALIPSO observations is
149 presented in Annex A.

150 Due to the difference of viewing angles and of pixel sizes between PARASOL and CALIPSO, cloudy and clear-sky
151 properties cannot be separated at the pixel scale. The clear-sky and cloud properties are computed at the $2^\circ \times 2^\circ$ scale
152 following the methodology proposed by [Konsta et al., 2012]. The monodirectional reflectance R averaged over each grid

cell depends on the clear-sky reflectance (CSR), on the cloud cover (CC) and on the cloud reflectance (CR) in this grid cell according to the relation

$$R = CC \times CR + (1-CC) \times CSR. \quad (2.1)$$

We assume that for each horizontal $2^\circ \times 2^\circ$ grid cell, the fraction CC of the highest values of monodirectional reflectance observed at the pixel level correspond to cloudy conditions and the fraction (1-CC) of the lowest values correspond to clear-sky conditions. In practice, the cloud cover CC is first determined from CALIPSO observations daily and for each $2^\circ \times 2^\circ$ grid cell. Then, the cloudy monodirectional reflectance CR is computed as the grid cell average of the fraction CC of the highest values of monodirectional reflectance observed by PARASOL at the pixel level. The clear-sky monodirectional reflectance CSR is the grid cell average of the fraction (1-CC) of the lowest values. Note that the cloud reflectance used here is different from the total reflectance generally used and which contains the contribution of clear-sky surrounding clouds. Except stated otherwise, the results and figures based on CALIPSO and PARASOL observations are for the two-year period 2007-2008.

2.3. PARASOL and CALIPSO simulators

By taking into account the effects of cloud overlap together with the specificities of the satellite, the model-to-satellite approach allows for a consistent comparison between the satellite observations described above (Section 2.2) and model outputs. The PARASOL and CALIPSO simulators have been designed for this purpose. The first step of the CALIPSO and PARASOL simulators consists in sub-gridding the model outputs (i.e. temperature, pressure, cloud cover, cloud condensate and effective radius of cloud droplets and ice crystals). The model vertical profiles are converted to an ensemble of subgrid-scale profiles by dividing each grid cell into a few tens of subcolumns generated randomly using the Subgrid Cloud Overlap Profile Sampler [Klein and Jakob 1999; Webb et al. 2001]. In each subcolumn, the cloud cover is assigned to be 0 or 1 at every model level with the constraint that the cloud condensate and cloud cover averaged over all subcolumns is consistent with the grid-averaged model diagnostics and the cloud overlap model assumption.

The monodirectional reflectance, the total cloud cover and the vertical cloud distribution as it would be seen by PARASOL and CALIPSO respectively are then computed in each subcolumn (see Chepfer et al. 2008 for CALIPSO and Annex B for PARASOL) and averaged over each model grid cell. Due to the large and highly variable reflection of solar light on ground surfaces, the PARASOL simulator is not used above continents.

Above oceanic regions, the cloud monodirectional reflectance CR is computed for every grid cell at each time step, using the same definition as for the observations (Eq. 2.1) :

$$CR = \frac{R - (1 - CC) \cdot CSR}{CC} \quad (2.2)$$

where the monodirectional reflectance R and the cloud cover CC have been previously computed by the simulator. The clear-sky reflectance has a fixed value in the simulator ($CSR=0.03$), which is consistent with the observations under clear-sky conditions [Konsta et al., 2012].

To simplify the post-processing of simulator outputs, the monodirectional reflectance is computed by the PARASOL simulator at every time step for a constant solar zenith angle corresponding to the A-train overpass at each latitude instead of the zenith angle corresponding to the local time in the GCM grid (Annex B). Sensitivity tests indicate that the cloud cover and reflectance computed by the PARASOL and CALIPSO simulator with the LMDZ model are almost insensitive to the frequency call of the simulator (every 1.5, 3, 6 hours), to the number of sub columns (if greater than 20 in each grid cell), and to the use of day or night time outputs (2% for the cloud cover and 2.5% for the reflectance; Figures not shown). In this study the simulator is called every 3 hours and 100 subcolumns are used.

As the reflectance is not commonly used for cloud description, an approximate relationship between cloud reflectance and cloud optical thickness is given. The cloud reflectance (CR) can be converted in cloud optical thickness ($C\tau$) using the coarse approximation of spherical particles when $T > 0^\circ\text{C}$, non-spherical particles when $T < 0^\circ\text{C}$ and $\theta_s = 30^\circ$ (see Annex B, Figure B.b). However, this cloud optical thickness should not be considered as the real cloud optical thickness but only as an approximation given for convenience. The rigorous comparison between models and observations should be made using the monodirectional reflectance. The same assumption to convert PARASOL cloud reflectances into cloud optical thicknesses are made for both observed and simulated data in order to make sure that this conversion does not introduce an artifact when comparing models and observations.

PARASOL and CALIPSO simulators are included in COSP (CFMIP Observational Simulator Package) [Bodas-Salcedo et al. 2011], which also includes other simulators such as the ISCCP [Klein and Jakob 1999; Webb et al. 2001] and the CloudSat [Haynes et al. 2007] simulators. The COSP simulator is available to the community via the Cloud Feedback Model Intercomparison Project (CFMIP) web page (cfmip.net) and the monthly and instantaneous datasets used here,, which are consistent with the COSP outputs, are available via the “CFMIP Observations for Model evaluation” website (climserv.ipsl.polytechnique.fr/cfmip-obs.html).

3. Assessment of cloud properties using monthly mean statistics

212 The basic properties of clouds simulated by climate models are usually evaluated by comparing monthly mean values of the
213 observed TOA fluxes, cloud cover, cloud optical depth and cloud top height (ie. Zhang et al. 2005, Klein et al. 2013). In this
214 section, this approach based on monthly mean statistics is followed to evaluate the two versions of the LMDZ climate
215 model.

216 **a) Cloud geographical distribution**

217 An aim of the improvements in the LMDZ5B model compared to LMDZ5A was to improve the cloud cover [Hourdin et al.,
218 2013-b] and this has been achieved in many regions. The cloud cover (CC) simulated by LMDZ5B is in better agreement
219 with observations than the CC simulated by LMDZ5A over the North Pacific and the North Atlantic, over the warm pool
220 (where there are high convective clouds), along the East coast of the oceans and over the trade wind regions (where cumulus
221 clouds dominate) but the CC values are still too low (Annex C). When considering the zonal mean values, LMDZ5B still
222 underestimates the cloud cover in the tropics even though the bias is reduced by a factor of two compared to LMDZ5A.

223 High-level clouds feature a better vertical distribution in LMDZ5B than in LMDZ5A with a lower cloud cover but this CC
224 is still too high (Annex C). LMDZ5B is able to simulate mid-level clouds even though they are still too few. Low-level
225 clouds are also better simulated in LMDZ5B with a larger cloud cover but the low-level clouds are too low and they are too
226 concentrated in one single layer. At middle and high latitudes, LMDZ5B simulates relatively well the large vertical extent of
227 the frontal clouds associated with storms. The cloud fraction is improved in LMDZ5B compared to LMDZ5A but the cloud
228 reflectance is not. Cloud reflectance and therefore cloud optical thickness are strongly overestimated by both model versions
229 almost everywhere and in particular over the subtropical oceans and in mid and high latitudes (Annex C).

230 **b) Tropical oceanic clouds in dynamical regimes**

231
232 Based on their geographical distribution the cloud properties discussed in the previous sections can be summarized in the
233 tropics using dynamical regimes [Bony et al., 2004]. Figure 1 shows the monthly mean cloud cover (Fig. 1a), cloud
234 reflectance (Fig. 1b) and SW albedo (Fig. 1c) as a function of the monthly mean vertical velocity at 500hPa (ω_{500}), as well
235 as the PDF of ω_{500} (Fig. 1d) over the tropical ocean. In the convective regions ($\omega_{500} < 0$), the cloud cover simulated by
236 LMDZ5B is closer to observations than the cloud cover simulated by LMDZ5A whereas in subsidence regions where
237 $\omega_{500} > 20$ hPa/day, the cloud cover is underestimated in both model versions (Fig. 1a). The cloud reflectance is strongly
238 overestimated in both model versions, this overestimation being smaller in convective regions where $\omega_{500} < -40$ hPa/day
239 (Fig. 1b). Despite these large discrepancies, both model versions simulate reasonably well the SW albedo in subsidence

240 regions ($\omega_{500} > 0$, Fig. 1c), which dominates in the Tropics (65% of the surface of the entire tropical belt). Even though both
241 model versions reproduce well the mean value and the overall geographical pattern of the albedo (Figures not shown), they
242 do not succeed in simulating properly the cloud cover and the cloud reflectance.

243 **c) Correlation between cloud top pressure and cloud optical thickness over the tropical ocean**

244

245 The joint histograms of cloud top pressure (CTP) and cloud optical thickness ($C\tau$) obtained with ISCCP data [Rossow and
246 Schiffer, 1991] have been widely used for atmospheric studies [e.g. Jakob and Tselioudis 2003, Rossow et al. 2005] and for
247 model evaluation [e.g. Webb et al, 2001, Williams and Tselioudis, 2007, Klein et al., 2013]. They are computed using two
248 methods. In the first method, the ISCCP-D2 data and ISCCP simulator outputs are directly used (Fig. 2d-f). In the second
249 method, the CALIPSO and PARASOL data and simulator outputs are used and CTP is defined as the pressure of the highest
250 level where the instantaneous CALIPSO-GOCCP cloud cover in each $2^\circ \times 2^\circ$ grid cell is greater than 0.1 (Fig. 2a-c).
251 Sensitivity tests show that results are not very sensitive to the value of this threshold when it varies from 0.1 to 0.3. The
252 cloud optical thickness $C\tau$ is computed from the cloud reflectance. The 2D histogram features these instantaneous CTP and
253 $C\tau$ data for each grid cell and averaged over time.

254
255 The histogram based on ISCCP data (Fig. 2d) is very different from the one based on CALIPSO-PARASOL data (Fig. 2a).
256 The 2D histogram over the tropical ocean obtained with the observed CALIPSO-GOCCP and PARASOL data (Fig. 2a)
257 features two distinct populations, the first one in the high troposphere and the second one in the low troposphere. The
258 histogram based on ISCCP data features a single cluster in the middle of the troposphere. This difference is a direct
259 consequence of the much-improved measurement of cloud height by active sensors (CALIPSO) compared to passive ones
260 (ISCCP), as already shown by Chepfer et al. [2008]. A more advanced comparison between ISSCP and merged CloudSat
261 and CALIPSO histograms is described in Mace and Wrenn [2013].

262
263 Similar features are also visible in the model results where two populations of low and high clouds are featured when using
264 the CALIPSO and PARASOL simulators (Fig. 2.b-c) and one single cluster of high clouds is featured when using the ISCCP
265 simulator (Fig. 2.e-f). The main biases of the models (e.g. too many high clouds, too few low clouds, too high cloud optical
266 thickness) also appear when using ISCCP observed and simulated data. However low clouds can be evaluated using
267 CALIPSO but they do not occur when the ISCCP simulator is applied. Similarly, the optically thin clouds can be evaluated
268 using CALIPSO but they are not detected in ISCCP observations nor simulated by the ISCCP simulator.

269 4. An advanced “process oriented” evaluation of tropical clouds description taking full advantage of the A-train 270 capability

271

272 4.1 The added value of evaluating the model using instantaneous cloud properties

273 For climate change and climate variability studies, it is important to characterize and assess how cloud properties vary as a
274 function of atmospheric variables describing the clouds environment. For example in climate feedbacks studies, changes ΔE
275 of the cloud environment variables are assumed to depend on the change ΔT of the surface temperature (e.g. Colman and
276 McAvaney, 1997, Soden and Held, 2006, Bony et al., 2006) and cloud properties are therefore assumed to vary as a function
277 of the surface temperature. Within this framework, various studies (e.g. Bony et al., 2004, Webb et al., 2006) have analyzed
278 how the cloud radiative effects vary in response to a change of the surface temperature at inter-annual time scales or under
279 climate change situations. This approach has been successful to improve our understanding of cloud feedbacks but the
280 temporal scale used in those studies (in general monthly mean variables) does not allow for the understanding of the direct
281 relationship between these results and the model parameterizations. Indeed, cloud properties vary instantaneously in
282 parameterizations (or with small time constants) when variables describing the environment (e.g. atmospheric stability,
283 humidity) vary. These dependencies between cloud and environment characteristics are highly non-linear so that the
284 relationship between these instantaneous variables may be very different from the relationship between their monthly or
285 seasonal mean values. The A-Train offers new possibilities to analyze the correlation between instantaneous cloud cover and
286 cloud reflectance in more detail.

287

288 Assuming for simplicity reasons that the clear-sky reflectance over ocean is negligible, the variation ΔR of the reflectance R
289 (eq. 2.1.) depends on how the cloud cover (CC) and the cloud reflectance (CR) vary:

$$290 \Delta R \approx CR \cdot \Delta CC + CC \cdot \Delta CR.$$

291 The variation ΔR will be very different if the variations ΔCC and ΔCR have the same or the opposite sign. The CC and the
292 CR variables estimated at the same time and location, as well as their joint variations are analyzed with a focus on the case
293 where they vary in the same or opposite way. An evaluation of how cloud properties vary with the environment is more
294 constraining for the models than an evaluation of the mean state and this may be one way to increase the confidence in
295 model results.

296

297 The relationship between cloud cover and optical thickness over the tropical oceans is shown in Figure 3. In observations,
298 this relationship has been shown to be significantly different when using monthly or instantaneous values [Konsta et al.,
299 2012]. When using monthly mean data (Fig. 3, lower line), the observations show an almost linear relationship between the

two variables: as the cloud cover increases, the cloud optical depth increases too (Fig. 3-d). The two versions of the model show very different behaviors from what is observed: the cloud optical depth remains almost constant when the cloud cover varies. LMDZ5A does not simulate the highest values of cloud cover and the simulated cloud reflectance is too high on average (Fig. 3-e). LMDZ5B simulates clouds that have a large cover and a reflectance consistent with observations, but clouds with a small cover have a too large reflectance (Fig. 3-f).

The use of instantaneous data gives a more accurate picture of the relationship between cloud cover and cloud optical depth and, for the models, a very different picture from what is observed (Fig. 3, upper line). For the observations, there is a tendency of increasing cloud reflectance with increasing cloud cover as it is the case for monthly mean values. But the instantaneous data reveal two separate cloud populations: one with a low cloud cover ($CC < 60\%$) and a low reflectance ($CR < 0.2$) and another one with a cloud cover close to 1 and a cloud reflectance value ranging from 0.1 up to 0.9 (Fig. 3-a). Intermediate cloud reflectance (0.2-0.4) and cloud cover (0.5-0.8) values occur when using monthly mean values but are much less frequent when using instantaneous values. These intermediate values therefore do not correspond to actual clouds. Both model versions simulate clouds with a low cover but their reflectances are much too high (Fig. 3-b,c). LMDZ5A simulates few clouds with a cover close to one whereas LMDZ5B simulates many clouds with a cover close to one, as observed. However, the cloud reflectance-cloud cover relationship for both model versions does not show an increase of the cloud reflectance with cloud cover, as this is the case in the observations. The LMDZ5B model even shows an opposite relationship.

The instantaneous correlation between cloud variables constitutes a key test to improve the confidence in how cloud properties simulated by climate models vary in changing situations. Fig. 3 clearly shows that this test is highly challenging. Instantaneous cloud properties are analyzed below.

4.2. Evaluation of tropical clouds using instantaneous clouds properties

4.2.a Optical thickness of clouds and their vertical distribution

The PDF of the cloud reflectance (CR) observed by PARASOL, and simulated by the model and by the PARASOL simulator (Figure 4a), confirms that both model versions strongly overestimate the cloud reflectance. The cloud population is divided into three different classes according to their optical thickness: (i) the optically thin and intermediate clouds ($CR < 0.2$, i.e. $C\tau < 3.41$) corresponding to 55%, 29% and 25% of the clouds for the observations, the LMDZ5A and the LMDZ5B model respectively, (ii) optically thick clouds ($0.2 < CR < 0.5$, i.e. $3.41 < C\tau < 11.42$) corresponding to 34%, 50% and 53% of the clouds for the observations, LMDZ5A and LMDZ5B respectively, and (iii) very thick clouds ($CR > 0.5$, i.e.

330 $C\tau > 11.42$) corresponding to the less populated class: 11%, 21% and 22% for the observations, LMDZ5A and LMDZ5B
 331 respectively. The mean vertical profile of the cloud fraction observed from CALIPSO-GOCCP and simulated by the model
 332 and by the simulator is shown on Fig.4b-d for the three cloud classes. Note that the cloud reflectance is a vertically
 333 integrated value and characterizes the whole atmospheric column, while the cloud fraction is a local value at each vertical
 334 level. For optically thin and intermediate clouds (i.e. when $C\tau < 3$ or $CR < 0.2$) the lidar traverses clouds and provides
 335 information for the whole atmosphere whereas for thicker clouds ($CR > 0.2$) the lidar signal is attenuated and does not
 336 provide information below the higher thick clouds.

337
 338 Observations show that optically thin and intermediate clouds ($CR < 0.2$, i.e. $C\tau < 3.4$, Fig.4b) are mainly low-level clouds
 339 with low values of cloud fraction ($CF \approx 0.15$). For the same range of optical thickness, both versions of the model simulate
 340 high-level clouds with intermediate values of the cloud fraction ($CF \approx 0.27$). LMDZ5A does not simulate any low-level
 341 clouds while LMDZ5B simulates some but too few of them ($CF < 0.05$). For optically thick clouds ($0.2 < CR < 0.5$, Fig.4c),
 342 both model versions correctly simulate high-level clouds with a mean cloud fraction ($CF \approx 0.17$) and altitude close to the
 343 observations. Both model versions simulate low clouds with a correct fraction but a too low altitude, and they fail to
 344 simulate mid-level clouds. Lastly, optically very thick clouds ($CR > 0.5$, Fig.4d) are mainly high-level clouds with a large
 345 cloud fraction ($CF \approx 0.37$) with some mid-level clouds ($CF \approx 0.2$) and only a few low-level clouds. The two model versions
 346 simulate optically very thick clouds with a very different cloud vertical structure, with much less high-level clouds
 347 ($CF \approx 0.15$ for LMDZ5A and 0.06 for LMDZ5B), almost no mid-level clouds and a fraction of low-level clouds ($CF \approx 0.1$)
 348 close to observations, also smaller.

349
 350 In summary, the fraction of mid- and high-levels clouds increases with the cloud reflectance in the observations but
 351 decreases in the models. The fraction of low-level clouds in the observation and in the models does not show a tendency of
 352 change with cloud reflectance, and the altitude of the simulated clouds is generally too low.

353 **4.2.b. Focus on high-level clouds**

354 In ascent regions, $2^\circ \times 2^\circ$ grid cells with no high-level clouds ($P < 440$ hPa) are rarely observed (15% of the time, Fig. 5-a) and
 355 rarely simulated by both models (10 to 20% of the time). The cumulative distribution function (CDF) of high-level cloud
 356 cover regularly increases (their frequency of occurrence is almost constant) until it becomes close to one. For these very
 357 cloudy conditions, there is a rapid increase in the CDF. The frequency of occurrence is much larger (typically 5 to 10 times)
 358 when the cloud cover is close to 1 than when it is smaller. The cloud cover of high-level clouds is larger than 95% in more
 359 than 25% of the situations. LMDZ5A simulates too few of these high-level clouds with a large cover but simulates too many

high-level clouds with a too low cover compared to observations. On the contrary, LMDZ5B simulates too many high clouds with a large cover and the high-cloud cover is larger than 95% in 40% of the situations. Situations with almost no high-level clouds are much more frequent in subsidence regions ($\approx 40\%$ of the cases, Fig. 5-b) than in ascent regions and situations with a large cover of high-level clouds are rare. Both model versions simulate the observed general behaviors of clouds although LMDZ5B is closer to observations. The normalized cloud cover of high-level cloud is now used to show how the different height categories (here, high clouds) are divided among all observed clouds [Stubenrauch et al., 2012]. The normalized cloud cover of high-level cloud is defined as $NCC_high = CC_high/CC$, where CC_high is the cloud cover of high-level clouds and CC is the total cloud cover [Konsta et al., 2012]. In ascent regions, the observed normalized high-cloud cover regularly increases with the total cloud cover (except for very small cloud cover) and reaches values close to one in fully overcast situations (Fig. 5-c). When the total cloud cover is small the high-level clouds have a small contribution to the total cloud cover, which means that mid- and low-level clouds dominate. On the contrary, high-level clouds dominate when the cloud cover is close to one. In LMDZ5A, high-level clouds always dominate even when the total cloud cover is small. This model fails to simulate enough mid- and low-level clouds in ascent regions. This bias is also in LMDZ5B but to a much lesser extent. In subsidence regions, the observed normalized high cloud cover is small and increases with the cloud cover (Fig. 5-d). Low-level clouds dominate, as expected. Both model versions simulate a too large normalized cover of high clouds, which is consistent with a too small value of the low-level cloud cover in these regions.

4.2.c. Focus on boundary layer clouds

Low-level clouds over the tropical oceans are now examined. To do so, only the atmospheric columns where low-level clouds are dominant are considered. These columns are defined as the $2^\circ \times 2^\circ$ grid cells where the normalized cover of low-level clouds ($P > 680$ hPa) is greater than 90% ($CC_low/CC > 90\%$). Figure 6 presents the relationship between the cloud cover and the cloud reflectance in these situations. The observed clouds may be organized in two groups (Fig. 6-a). In the first group, clouds have a small cover and a small reflectance. Further analysis shows that these clouds are present all over the tropics, with the most dominant population confined in the trade cumulus regions. Their properties are consistent with those of small cumulus clouds. The second group is composed of clouds with cover close to one and with a large reflectance ($0.3 < CR < 0.6$), i.e. a large optical thickness ($5.5 < C\tau < 17$). They are mainly located on the east coast of the tropical oceans and their properties are consistent with those of stratocumulus clouds. The results shown in Fig 6-a are broadly consistent with those obtained by Cole et al. [2011] using CERES and MODIS observations (their Fig. 7). The LMDZ5A model results show very different characteristics (Fig. 6-b) with most of the low-level clouds having a too large cloud cover and a too large reflectance, i.e. a too large optical thickness. The LMDZ5B model simulates two clusters of low-level clouds, one with small cloud cover values and another one with a cloud cover close to one, which is more consistent with observations.

390 However, when the cloud cover is small, the reflectance is much too large and increases when the cloud cover decreases,
 391 unlike in the observations.

392

393 This relationship is further examined by focusing on specific cloud regimes, following the methodology proposed by
 394 Medeiros and Stevens [2011] that allows for a separation in stratocumulus and shallow cumulus regimes. The stratocumulus
 395 regime corresponds to a regime where clouds have the largest values of cloud cover and cloud reflectance both in the
 396 observations and in the model. The shallow cumulus regime corresponds to clouds with smallest values of cloud cover and
 397 cloud reflectance (Figures not shown). For the stratocumulus regime, the observed relationship between cloud cover and
 398 cloud reflectance is quasi linear. The same relationship is also examined using the ISCCP 3hourly observations on the $2^\circ \times 2^\circ$
 399 grid and the ISCCP simulator (Fig. 6d,e,f). The ISCCP cloud optical thickness is converted into cloud reflectance by
 400 assuming a fixed solar zenith angle $\theta_s = 30^\circ$ and spherical particles. The results from ISCCP are consistent with those
 401 obtained with CALIPSO-PARASOL, both for the observations and the models. However, the confidence in CALIPSO-
 402 PARASOL data is higher because the detection of low-level clouds is much better compared to ISCCP, as noted above. In
 403 CALIPSO-PARASOL data, cumulus clouds are found to have larger values of cloud cover compared to ISCCP
 404 observations. The decrease of cloud reflectance with cloud cover observed with CALIPSO-PARASOL is not evident with
 405 ISCCP.

406

407 According to theory and observation, the cloud optical thickness increases with the cloud top height for low-level clouds of
 408 same base height: as the cloud grows vertically, there is more water to condensate and the cloud optical depth increases.
 409 This relationship is a function of many phenomena (e.g. turbulent mixing, precipitation efficiency) that are not accurately
 410 known. This motivated many field campaigns [e.g. Coakley et al., 2005, Siebesma et al., 2003]. The analysis of this
 411 relationship on the global scale is performed using the CALIPSO-PARASOL observations. Figure 7 shows the mean cloud
 412 reflectance as a function of the cloud top pressure when low-level clouds are dominant. The cloud top pressure is defined as
 413 the first layer going downward from the 680hPa where the cloud cover is greater than 0.1. Both model and observations
 414 show that the cloud optical depth increases with cloud top altitude, as expected. However, the cloud optical thickness
 415 simulated by the models is two to three times larger than the observed one.

416

417 The poor representation of the low-cloud properties by the models may have important consequences for climate change
 418 studies. Indeed, low-level clouds cover most of the tropical ocean and are the main source of spread in climate sensitivity
 419 estimates [Bony and Dufresne, 2005; Vial et al., 2013]. In addition, the amplitude of the low-level cloud feedback depends

420 on the cloud radiative effect [Brient and Bony, 2012] and an error in the later may impact the value of the former.

421

422 **4.3. From model evaluation to model improvement**

423

424 The analysis of monthly mean values confirm that the LMDZ5 model, as many other models, simulate low-level clouds
425 with a too low cloud cover and a too high optical thickness (Nam et al. 2012, Klein et al. 2013). The use of instantaneous
426 values further highlights the fact that the mean values as well as the variation of the cloud optical thickness with the cloud
427 cover in the LMDZ model were biased. Key deficiencies in the model parameterizations can be identified and improved
428 using the diagnostics presented above. Presenting new parameterizations for low-level clouds is far beyond the scope of this
429 paper but the proposed diagnostics are very relevant for future model developments. The major discrepancy in models
430 compared to observations identified above with increasing optical thickness as the cloud cover decreases, (Fig. 6) is further
431 analyzed in the LMDZ5A model.

432

433 Many factors affect both the cloud cover and the cloud reflectance but a sensitivity analysis shows that the main driver of
434 the erroneous relationship between these two variables in the LMDZ5A model is the liquid water content and not the micro-
435 physic properties of clouds such as the cloud droplet size. The vertical integrated cloud water amount (or liquid water path)
436 for low-level clouds increases as the cloud cover increases, like the cloud reflectance increases as the cloud cover increases,
437 and the liquid water path is strongly correlated with the cloud reflectance (Figure 8-a).

438

439 The cloud fraction and liquid water content in the LMDZ5A model are diagnosed from the large-scale value of the total
440 (vapor + condensed) water Q_t , the moisture at saturation Q_s , and the subgrid scale variability of the total water using a
441 generalized log-normal Probability Distribution Function (PDF) defined by three statistical moments (mean, variance,
442 skewness) (Bony and Emanuel, 2001 ; Hourdin et al., 2006). This parameterization was originally developed for convective
443 clouds and then applied for all cloud types. Off-line calculations show that the increase of cloud reflectance when
444 decreasing cloud cover (Fig. 6-b) cannot be explained under the conditions that generally exist when low-level clouds are
445 present. This increase of cloud reflectance when the cloud fraction decreases is due to the deep convection scheme, which
446 was activated quite frequently, without producing any deep convective clouds but affecting the PDF of the total water
447 content at low level and therefore the cloud properties there. Sensitivity simulations confirm this hypothesis and show that
448 the cluster of points in the upper left part of Fig. 6-b with high cloud reflectance and low cloud cover no longer exists when
449 the deep convection scheme is switched off. However, the observed increase of reflectance with cloud cover is not

450 reproduced by the model.

451

452 An implicit assumption of most (if not all) PDF parameterization approaches is that the sub-grid cloud cover is
453 homogeneous (i.e. is either 0 or 1) in the vertical in each atmospheric layer. This assumption may be relevant when the sub-
454 grid total water content is far above or far below the moisture at saturation but it may be questionable when the two values
455 are similar. A very simple test is therefore performed to check the sensitivity of the cloud characteristics to this assumption.
456 The equation for the liquid water content is assumed to be unchanged and clouds are assumed to only cover a vertical
457 fraction of the atmospheric layer when the sub-grid humidity is close to the saturation and that this vertical fraction varies
458 linearly from 0 when $Q_t = Q_s - \Delta Q$ to 1 when $Q_t = Q_s + \Delta Q$. In practical terms, the only change is to use $Q_s - \Delta Q$ instead of Q_s
459 in the cloud cover formula. For the vertical spread of the humidity ΔQ , the same value as the one used to characterize the
460 horizontal spread of humidity (Hourdin et al., 2006) is used. The typical values of $\Delta Q/Q_t$ are about a few percent and are
461 comparable with the change of Q_s due to the vertical gradient of temperature within an atmospheric layer. These values of
462 the vertical spread of humidity are assumed to be realistic even though this choice is arbitrary and only relevant for a
463 sensitivity test. The impact of this change on the cloud cover and on the relationship between cloud cover and cloud
464 reflectance is very large for all clouds (Fig. 8-b to compare with Fig. 3 -b) and for conditions where low-level clouds are
465 dominant (Fig. 8-c to compare with Fig. 6-b). Above all, the cloud reflectance increases with the cloud cover, which is
466 consistent with observations and opposite to the results obtained with the original parameterization. A direct consequence is
467 a modification of the SW flux at the TOA of about 10 W/m^{-2} . The geographical distribution of the simulated cloud
468 reflectance is much closer to the observed values and the cloud cover is more realistic with more mid- and low-level clouds
469 simulated, which is closer to the observed vertical structure of clouds. Although much more work is required to develop a
470 new parameterization, the sensitivity test presented here illustrates the fact that the diagnostic is accurate enough to help
471 identifying the origins of a major problem in the simulated low-level cloud properties and to show the direct effect of a
472 specific modification of the model parameterizations.

473

474 Various studies have already analyzed how the effect of sub-grid heterogeneity of cloud properties may affect the
475 parameterization of cloud radiative properties [e.g. Barker and Wielicki, 1997; Li et al. 2005] or autoconversion rate [e.g.
476 Kawai and Texeira, 2012; Boutle et al., 2014], but very few have analyzed the effect of sub-grid heterogeneity of cloud
477 properties on the parameterization of cloud cover even though this possibly large effect has been recognized [Pincus and
478 Klein, 2000]. Based on results obtained with large eddy simulation (LES) models, Neggers et al. [2011] found that this
479 effect is large for cumulus cloud type, which is consistent with the present hypothesis. To our knowledge few (if any)

480 atmospheric models consider this phenomena in their parameterization. The possible role of this simplification on the
481 tendency of models to simulate too few and too bright clouds deserve further investigation and is beyond the scope of this
482 study.

483 **5. Summary and conclusion**

484

485 An evaluation of cloud characteristics (e.g. cloud cover, cloud vertical distribution and cloud optical depth) simulated by
486 two versions of the LMDZ5 GCM using observations from PARASOL and CALIPSO has been presented. Model and
487 observations have been compared using “observations simulators”, which allow direct comparisons between modeled and
488 level 1 observed data by taking into account the spatial scale differences between model and observations and by avoiding
489 most of the *a priori* hypothesis usually made in retrieval algorithms. This evaluation was performed using both monthly
490 mean and instantaneous values and it has been shown how the latter allows for further analysis and may be used to help
491 improve cloud parameterizations.

492

493 The comparison between the two versions of the LMDZ5 model and PARASOL/CALIPSO observations using the monthly
494 mean climatologies clearly shows an improvement in the representation of cloud cover and cloud vertical distribution of
495 LMDZ5B compared to LMDZ5A. This improvement consists of an increase of the cover of the boundary layer clouds
496 especially in the trade wind regions, an improvement of the altitude and, to a less extent, of the fraction of the high-level
497 clouds and the simulation of the large vertical extent of the frontal clouds associated with storms in the middle and high
498 latitudes. Although reduced, some model biases in LMDZ5A are still present in LMDZ5B and are generally shared with
499 many climate models, such as the lack of mid-level clouds in the mid latitudes [Chepfer et al., 2008, Zhang et al., 2005], the
500 presence of optically too thick high-level clouds all around the globe [Zhang et al., 2005], and the lack of boundary layer
501 clouds all around the tropical belt (in particular in the trade wind regions). Both versions of LMDZ5 model simulate the SW
502 albedo well although they strongly overestimate the cloud optical depth. This suggests some compensating error between
503 the cloud cover and the cloud optical depth and this compensation is different for the two model versions.

504

505 When using the instantaneous relationship between the cloud properties, new features appear and more precise conclusions
506 may be drawn. Over the tropical oceanic regions, the following results are obtained:

507 1. Observed clouds are grouped in two clusters, one where clouds have a low to mid cover ($CC < 60\%$) and a low reflectance
508 ($CR < 0.2$) and another one with a cover close to 1 and a reflectance that ranges from 0.1 to 0.9. The two model versions
509 reproduce these two clusters but with a lack of fully overcast situations ($CC = 1$) in LMDZ5A. In both model versions, there

510 are too few clouds with a small cover and their reflectance is strongly overestimated. There is a general tendency of
511 increasing cloud reflectance with increasing cloud cover in observations and the models do not reproduce this tendency.
512 Note that the relation between cloud fraction and cloud optical thickness are very different when using monthly mean or
513 instantaneous values both in observations and in the models.

514 2. The CALIPSO-GOCCP data allow for a detailed description of the vertical distribution of clouds. In the observations, the
515 fraction of mid- and high-levels clouds increases with the all clouds reflectance, whereas it decreases in the models. In
516 ascent regions, there are frequent situations where grid cells are fully overcast with high-level clouds ($CC_{high} > 95\%$): 25%
517 in observations, 10% and 40% in LMDZ5A and LMDZ5B respectively. In observations high-level clouds co-exist with low-
518 and mid-level clouds whereas the multi-layer clouds are much less frequent in models.

519 3. The tropical low-level cloud properties can be grouped in two clusters. The first one corresponds to cumulus-type clouds
520 and the second one corresponds to stratocumulus-type clouds. Observations show that the cloud optical depth increases with
521 cloud cover and none of the model versions reproduce this general trend; they may even feature an opposite trend. Both
522 models underestimate the low-level cloud cover but overestimate their reflectance. They do not produce enough low-level
523 clouds and the low-level clouds they produce are too thick. These two biases partly compensate each other when
524 considering the SW albedo. On average, the altitude of the low-level clouds simulated by the models is too low.

525 4. The relationship between instantaneous cloud cover and cloud reflectance for low-level clouds may be directly compared
526 to what is expected from the parameterizations in these conditions. This diagnostic allows for the identification of key
527 deficiencies in the parameterization and their origin as well as for possible improvements. More precisely, part of the
528 problem is due to intermittent triggering of the convection and another part of the problem is due to a too large in-cloud
529 liquid water amount. For the latter, the origin of the problem is believed to be the assumption of current parameterization
530 that the cloud covers is homogeneous all along the vertical in each atmospheric layer. Sensitivity tests were performed and
531 have shown that suppressing this assumption may have a large impact, leading to potential improvements in the cloud
532 characteristics. More work is required to go from these sensitivity experiments to establishing useful new parameterizations.

533

534 Multi-instrument missions like the A-train offer the possibility to observe many properties of the clouds and their
535 environment, which allows for a better evaluation of the climate simulated by atmospheric models. Beyond the separate
536 analysis of each variable, the analysis of their joint variations allows for a deeper analysis and a better understanding of the
537 dominant physical processes driving cloud properties. This is the case when using monthly mean values and using
538 instantaneous data is even more powerful. Even if the instantaneous observations are less precise than averaged values and
539 even if the collocation procedure may lead to the rejection of many observations, the analysis of their joint variations allows

540 for a more precise evaluation of cloud properties, it may highlight new features or problems, it may facilitate the link
541 between observations and parameterizations. It may also help to bridge the gap between model evaluation and model
542 development.

543

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552

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732 **Annex**

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734 **Annex A: Spatio-temporal sampling of CALIPSO and PARASOL observations**

735

736 (i) Temporal resolution:

737 The two instruments follow the same sun-synchronous A-train orbit, which passes over each location twice a day at about
738 1:30 AM and 1:30 PM local solar time. As PARASOL collects measurements during daytime, only the daytime CALIPSO
739 data are considered. The two instruments fly over the same orbit so they document the same cloud parcel simultaneously at
740 about 1:30 AM local solar time.

741 The incomplete sampling of the diurnal cycle has negligible impact (less than 1%) on the results [Chepfer et al., 2008].

742

743 (ii) Spatial resolution

744 As a PARASOL pixel (6x6km) is much larger than a CALIOP/CALIPSO one (330m along-track, 75m cross-track), one
745 value of the directional reflectance is associated to at least 18 lidar profiles. To overcome these differences, the CALIOP
746 cloud cover and the PARASOL reflectance are processed independently on a statistical basis and then compared to daily
747 mean values on a $2^\circ \times 2^\circ$ grid (several hundreds of km^2). To test the impact of the sampling over seasonal mean results on a
748 $2^\circ \times 2^\circ$ grid, two PARASOL reflectance datasets were built in the same viewing direction ($\theta_v = 27^\circ, \phi_s - \phi_v = 320^\circ$): the first
749 dataset includes all reflectance values measured by PARASOL, and the second dataset includes only the reflectance
750 measured along the CALIPSO ground track. The maximum distance between a PARASOL and a CALIOP pixel in the first
751 dataset is 50km. The number of measurements is about 30% lower in the second dataset. Maps of $2^\circ \times 2^\circ$ mean directional
752 reflectances and variances (not shown) are similar for both datasets although the second one is noisier, thus suggesting that
753 both PARASOL datasets (collocated or not with CALIOP) can be analyzed. The similarity between the two datasets also
754 shows that the few PARASOL pixels collocated with CALIOP ($6 \times 6 \text{ km}^2$) are representative of all PARASOL pixels
755 included in the $2^\circ \times 2^\circ$ grid cell.

756 Similarly, it is thus reasonable to consider that the CALIOP dataset (even with a 330mx75m resolution) when averaged over
757 several months is statistically representative of the monthly/seasonal cloud cover within a GCM grid cell.

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Annex B: Sensitivity of the PARASOL monodirectional reflectance to the atmosphere's composition

(i) Optical properties

The cloud particle optical properties (e.g. single scattering albedo, scattering phase function, extinction coefficient) depend on the wavelength, the particle size and its shape. As the absorption phenomena is negligible in both ice and water at 864nm [Warren et al 1984 ; Hale and Querry, 1973], the single scattering albedo is close to one regardless of the size and shape of the particles. As the radius of cloud particles are always larger than the wavelength considered here, the scattering phase function is not very sensitive to the droplet size but it is sensitive to the particle shape. A spherical shape assumption, which is typical of liquid water computed with the Mie theory and a non-spherical shape, which is typical [Chepfer et al. 2002] of ice crystals whose optical properties are computed with Geometric Optic enhanced with Finite Differential Time Domain [Yang et al 2000 and 2001] are used. As shown in Fig. C.a, their scattering phase functions differ significantly for scattering angles close to backscattering (180°), haloes (22° and 44°) and rainbow (140°), and also between 90° and 130° , which corresponds to the viewing and solar zenith angle selected for PARASOL data in the Tropics. Complementary computations (not shown) indicate that the scattering phase function (at this wavelength) weakly depends on the particle size compared to the influence of the shape. On the contrary, the particle extinction coefficient is directly dependent on the particle size: it is proportional to the scattering efficiency (close to 2 as the particles are large than the wavelength) multiplied by the particle cross section, which is expressed as the function of the particle size.

(ii) Radiative transfer computations

The directional reflectance is computed using a doubling-adding radiative transfer code [DeHaan et al. 1986]. The cloud particles optical properties such as the single scattering albedo and the truncated scattering phase function developed in Legendre polynomial are introduced in the radiative transfer code. The Rayleigh scattering is also taken into account in the computation even though its contribution to the total directional reflectance remains small (τ is about 0.013 for the whole atmospheric column). As the studied viewing direction is off-glitter, the ocean is described as a lambertian surface with a constant plane albedo of 0.03. The directional reflectance is then computed as in Chepfer et al. [2002] for various cloud optical depths and solar zenith angles.

Figure C.b shows that changes of reflectance values due to solar zenith angle variations are less than 0.1 in the tropical regions (30°S - 30°N , $18^\circ < \theta_s < 60^\circ$) for a given phase function. It reaches a maximum of 0.15 between the ITCZ and the

791 higher observable latitudes ($\theta_s > 60^\circ$). Thus variations of the latitudinal reflectance larger than 0.15 (0.1 in the Tropics)
 792 cannot be attributed to variations of θ_s . They are due to changes in the atmosphere composition (clouds). The sensitivity of
 793 the reflectance to the cloud particles scattering phase function is maximum at high latitudes / high solar zenith angle (0.13)
 794 and slightly reduces in the tropics (0.1).

795

796 (iii) *PARASOL simulator*

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798 The PARASOL simulator is initiated with the mixing ratios of in-cloud liquid and ice water content in each model grid cell
 799 that are then converted into sub-grid mixing ratios using SCOPS. In each subcolumn, the total cloud optical depth (τ_{tot})
 800 is the sum of the subcolumn ice (τ_{tot_ice}) and liquid (τ_{tot_liq}). These are computed assuming that the cloud particles
 801 are spherical with a radius equal to the effective radius predicted by the model. For five solar zenith angles ($\theta_s = 0^\circ, 20^\circ,$
 802 $40^\circ, 60^\circ$ and 80°) and knowing the total cloud optical depth, two directional reflectance values are then computed for
 803 each day and for each solar zenith angle assuming that the cloud is entirely composed of liquid water (Refl_liq) or ice
 804 water (Refl_ice). These reflectance values are derived from a bilinear interpolation over pre-calculated look-up tables
 805 containing results of radiative transfer computations (Annex B) for the cloud particle's shape assumption (spherical and
 806 non spherical) made in the model. The subgrid directional reflectance is then computed as follow: $Refl =$
 807 $(Refl_liq * \tau_{tot_liq} + Refl_ice * \tau_{tot_ice}) / \tau_{tot}$. The directional reflectance obtained for each subgrid is then averaged
 808 over each GCM grid cell for each day and for each θ_s . After the simulations have been performed, the five
 809 monodirectional reflectances corresponding to the five solar zenith angles from the simulator's outputs are used to
 810 interpolate linearly the monodirectional reflectance depending on the monthly mean value of the solar zenith angle at
 811 each grid point. The simulated monodirectional reflectance is then directly comparable to the observations.

812 **Figure B:**

813 a) Scattering phase function for spherical and non-spherical particles. Monodirectional reflectance simulated as a
 814 function of the solar zenith angle for spherical and non-spherical particles in the viewing direction ($\theta_v = 27^\circ \phi_v = 320^\circ$).

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836 **Annex C: Traditional global monthly mean evaluation of cloud properties**

837 a) Cloud cover

838 On average, cloud cover is underestimated over the tropical regions and is broadly consistent with observations in the mid
839 and high latitudes (Fig. C1). In the Tropical Western Pacific, along the ITCZ and the SPCZ, the cloud cover simulated by
840 LMDZ5A is about 60-70% whereas observations indicate a cloud cover ranging from 80% to 100%. In regions where the
841 cloud cover is low, such as in the trade wind cumulus region, observations indicate a cloud fraction between 40 and 60%
842 whereas the simulated cloud cover is only about 20 to 50%. Although LMDZ5B underestimates the averaged cloud cover in
843 the tropics, the bias is reduced by a factor close to 2 compared to LMDZ5A. The improvement is very significant in almost
844 all fully overcast regions (e.g. warm-pool, east Pacific and Atlantic) and even with an overestimated cloud cover.

Figure C1 : Geographical distribution of the total mean cloud cover over the ocean averaged over the period 2007-2008 (a) observed with CALIPSO-GOCCP during day time, (b) simulated with LMDZ5A and the lidar simulator, (c) simulated with LMDZ5B and the lidar simulator; (d) zonal mean of the same quantity observed (red line) and simulated (LMDZ5A: black line and LMDZ5B: black dotted line).

845

846 b) Cloud vertical profile

847 The zonal mean vertical distribution of the observed CALIPSO-GOCCP cloud fraction clearly highlights the well-known
848 links between cloud characteristics and large circulation of the atmosphere (Fig. C2-a). The altitude of the higher clouds
849 follows the tropopause height and decreases from the equator to the poles. The LMDZ5A model with the lidar simulator
850 produces a cloud fraction of high-level clouds that is too large almost everywhere and the altitude of these clouds is too
851 high, in particular over the polar region in the southern hemisphere (Fig. C2-b). In the tropics, LMDZ5A strongly

852 underestimates the cloud fraction at low and middle altitudes. Although this feature is amplified by the masking effect of
 853 high clouds on the lidar signal (thick high level clouds, with typical $C\tau > 3$, attenuate the signal and mask low- and mid-level
 854 clouds that might exist below them), this underestimation already occurs with the cloud cover simulated by the model (i.e.
 855 without using the lidar simulator, cf Chepfer et al., 2008). At higher latitudes, the model cannot simulate the large vertical
 856 extent of the frontal clouds associated with storms. Instead, it simulates two separate groups of low- and high-level clouds.
 857 This zonal mean vertical distribution of clouds is improved in LMDZ5B (Fig. C2-c). In the tropics, boundary level clouds
 858 are present although they are too low and too concentrated in one single layer. At middle and high latitudes, the model
 859 almost simulates the continuous vertical structure of the cloud fraction.

Figure C2: Zonal mean cloud fraction profile averaged over the period 2007-2008, a) observed from CALIPSO-GOCCP, b) simulated with LMDZ5A and the lidar simulator and c) simulated with LMDZ5B and the lidar simulator.

c) Cloud reflectance

In the trade wind regions, the observed cloud reflectance (typical value of 0.15) is only slightly higher than the clear-sky value (approximately 0.03), indicating that clouds are optically thin. This is not the case for the two model versions (Figure C3-b,c). They strongly overestimate the cloud reflectance almost everywhere, in particular over the subtropical oceans and

in the mid and high latitudes. The models cannot reproduce the contrast between the higher values (≈ 0.3) of cloud reflectance observed along the ITCZ and the Eastern Pacific ocean and the lower values (< 0.2) over the tropical trade wind cumulus region. They simulate high cloud reflectances (> 0.2) over the tropics. On average, the cloud reflectance, and therefore the cloud optical thickness, simulated by the models over the ocean is too high almost everywhere.

Figure C3: Same as Fig. C1 for the monodirectional cloud reflectance over ocean, on average for the period 2007-2008 (a) observed with PARASOL, (b) simulated with LMDZ5A and the PARASOL simulator, (c) simulated with LMDZ5B and the PARASOL simulator; (d) zonal mean of the same quantity observed (red line) and simulated (LMDZ5A: black line and LMDZ5B: black dotted line).

860 List of Figures

861 1. (a) Cloud cover, (b) cloud reflectance (approximative optical depth on the right y-axis, see Sect. 2.3), (c) SW

862 albedo as a function of the vertical velocity ω_{500} and d) PDF of ω_{500} , observed (coloured lines), simulated with LMDZ5A
 863 (black line), simulated with LMDZ5B (black dotted line). For the cloud cover and the cloud reflectance the simulator is
 864 used for model results. All the data are monthly means over the tropical oceans. The monthly mean vertical velocity are
 865 from the ERA interim reanalysis [Simmons et al. 2007] and the short wave planetary albedo is estimated from CERES -
 866 EBAF (Clouds and the Earth's Radiant Energy System) [Loeb et al. 2009].

867 2. Cloud top pressure versus optical thickness (PC-tau) histograms, computed over the tropical oceans (a,d) observed,
 868 (b,e) simulated with the LMDZ5A model and the simulator and (c,f) simulated with the LMDZ5B model and the simulator.
 869 The upper line (a,b,c) corresponds to CALIPSO-GOCCP and PARASOL data (2007-2008) and simulator and the lower line
 870 (d,e,f) corresponds to ISCCP-D2 data (1983-2007) and simulator respectively. The color bar represents the number of points
 871 at each grid cell (PC-tau) divided by the total number of points in the histogram.

3. 2D histograms of cloud reflectance and cloud cover over the tropical oceans (a,d) observed with PARASOL and
 CALIPSO GOCCP, (b,e) simulated with LMDZ5A and the simulator, and (c,f) simulated with LMDZ5B and the simulator.
 The upper line (a,b,c) corresponds to instantaneous values and the lower line (d,e,f) corresponds to monthly mean values.
 The color bar represents the number of points at each grid cell (cloud cover-cloud reflectance) divided by the total number
 of points.

4. (a) PDF of cloud reflectances and (b,c,d) vertical profile of the cloud fraction CF3D for three classes defined by the
 grid average cloud reflectance (b) $CR < 0.2$, (c) $0.2 < CR < 0.5$, (d) $CR > 0.5$. Red line corresponds to observed values from
 PARASOL and CALIPSO-GOCCP respectively, black line corresponds to simulated values with LMDZ5A and the
 simulator and black dotted line corresponds to simulated values with LMDZ5B and the simulator. The data are
 instantaneous values and are taken only over the tropical oceans.

5. Cumulated distribution function of the cloud cover for high level clouds (upper row) and average relationship
 between total cloud cover and the normalized cloud cover of high level cloud (lower row, see text) in ascent (left column)
 and subsidence (right column) regions, observed with CALIPSO-GOCCP (red line), simulated with LMDZ5A and the
 simulator (black line), and simulated with LMDZ5B and the simulator (black dotted line). The cloud cover are
 instantaneous values and the ascent and the subsidence regions are defined as regions over the tropical ocean where the
 monthly mean value of ω_{500} is respectively negative and positive.

6. 2D histograms of instantaneous cloud reflectance and cloud cover over the tropical ocean for situations where low
 level clouds dominate ($CC_{low} > 0.9 * CC$) (a, d) observed, (b, e) simulated with LMDZ5A and the simulator, and (c, f)
 simulated with LMDZ5B and the simulator. The upper line (a,b,c) corresponds to PARASOL observations, CALIPSO

observations and simulator respectively and the lower line (d,e,f) corresponds to ISCCP. The color bar represents the number of points at each grid cell (cloud cover-cloud reflectance) divided by the total number of points.

7. Instantaneous mean cloud reflectance as a function of cloud top pressure for mainly low-cloud situations (using the criterion: $CC_{low} > 0.9 * CC$) over the tropical ocean, observed with PARASOL and CALIPSO-GOCCP (red line), simulated with LMDZ5A and the simulator (black line), and simulated with LMDZ5B and the simulator (black dotted line).

CTP is defined as the highest level of low clouds where the local cloud cover is greater than 0.1.

8. 2D histograms of instantaneous of (a) cloud liquid water path versus cloud reflectance simulated with LMDZ5A for conditions where low-level-clouds dominate ($CC_{low} > 0.9 * CC$) (b) cloud reflectance versus cloud cover with the modified parameterization in LMDZ5A (see text) for all clouds and (c) for conditions where low-level-clouds dominate ($CC_{low} > 0.9 * CC$). The color bar represents the number of points at each grid cell divided by the total number of points.

Figures

Figure 1: (a) Cloud cover, (b) cloud reflectance (approximative optical depth on the right y-axis, see Sect. 2.3), (c) SW albedo as a function of the vertical velocity ω_{500} and d) PDF of ω_{500} , observed (coloured lines), simulated with LMDZ5A (black line), simulated with LMDZ5B (black dotted line). For the cloud cover and the cloud reflectance the simulator is used for model results. All the data are monthly means over the tropical oceans. The monthly mean vertical velocity are from the ERA interim reanalysis [Simmons et al. 2007] and the short wave planetary albedo is estimated from CERES - EBAF (Clouds and the Earth's Radiant Energy System) [Loeb et al. 2009].

Figure 2: Cloud top pressure versus optical thickness (PC-tau) histograms, computed over the tropical oceans (a,d) observed, (b,e) simulated with the LMDZ5A model and the simulator and (c,f) simulated with the LMDZ5B model and the simulator. The upper line (a,b,c) corresponds to CALIPSO-GOCCP and PARASOL data (2007-2008) and simulator and the lower line (d,e,f) corresponds to ISCCP-D2 data (1983-2007) and simulator respectively. The color bar represents the number of points at each grid cell (PC-tau) divided by the total number of points in the histogram.

Figure 3: 2D histograms of cloud reflectance and cloud cover over the tropical oceans (a,d) observed with PARASOL and CALIPSO GOCCP, (b,e) simulated with LMDZ5A and the simulator, and (c,f) simulated with LMDZ5B and the simulator. The upper line (a,b,c) corresponds to instantaneous values and the lower line (d,e,f) corresponds to monthly mean values. The color bar represents the number of points at each grid cell (cloud cover-cloud reflectance) divided by the total number

of points.

Figure 4: (a) PDF of cloud reflectances and (b,c,d) vertical profile of the cloud fraction CF3D for three classes defined by the grid average cloud reflectance (b) $CR < 0.2$, (c) $0.2 < CR < 0.5$, (d) $CR > 0.5$. Red line corresponds to observed values from PARASOL and CALIPSO-GOCCP respectively, black line corresponds to simulated values with LMDZ5A and the simulator and black dotted line corresponds to simulated values with LMDZ5B and the simulator. The data are instantaneous values and are taken only over the tropical oceans.

Figure 5: Cumulated distribution function of the cloud cover for high level clouds (upper row) and average relationship between total cloud cover and the normalized cloud cover of high level cloud (lower row, see text) in ascent (left column) and subsidence (right column) regions, observed with CALIPSO-GOCCP (red line), simulated with LMDZ5A and the simulator (black line), and simulated with LMDZ5B and the simulator (black dotted line). The cloud cover are instantaneous values and the ascent and subsidence regions are defined as regions over the tropical ocean where the monthly mean value of ω_{500} is respectively negative and positive.

Figure 6: 2D histograms of instantaneous cloud reflectance and cloud cover over the tropical ocean for situations where low level clouds dominate ($CC_{low} > 0.9 \cdot CC$) (a, d) observed, (b, e) simulated with LMDZ5A and the simulator, and (c, f) simulated with LMDZ5B and the simulator. The upper line (a,b,c) corresponds to PARASOL observations, CALIPSO observations and simulator respectively and the lower line (d,e,f) corresponds to ISCCP. The color bar represents the number of points at each grid cell (cloud cover-cloud reflectance) divided by the total number of points.

Figure 7: Instantaneous mean cloud reflectance as a function of cloud top pressure for mainly low-clouds situations (using the criterion: $CC_{low} > 0.9 \cdot CC$) over the tropical ocean, observed with PARASOL and CALIPSO-GOCCP (red line), simulated with LMDZ5A and the simulator (black line), and simulated with LMDZ5B and the simulator (black dotted line). CTP is defined as the highest level of low clouds where the local cloud cover is greater than 0.1.

Figure 8: 2D histograms of instantaneous of (a) cloud liquid water path versus cloud reflectance simulated with LMDZ5A for conditions where low-level-clouds dominate ($CC_{low} > 0.9 \cdot CC$) (b) cloud reflectance versus cloud cover with the modified parameterization in LMDZ5A (see text) for all clouds and (c) for conditions where low-level-clouds dominate

($CC_{low} > 0.9 * CC$). The color bar represents the number of points at each grid cell divided by the total number of points.