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## Characterization of vertical cloud variability over Europe using spatial lidar observations and regional simulation

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# 1 **Characterization of vertical cloud variability over Europe using** 2 **spatial lidar observations and regional simulation**

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## 53 ABSTRACT

54 In this paper we characterize the seasonal and inter-annual variabilities of cloud fraction profiles in both  
55 observations and simulation since they are critical to better assess the impact of clouds on climate variability.  
56 The spaceborne lidar onboard CALIPSO, providing cloud vertical profiles since 2006, is used together with a  
57 23-year WRF simulation at 20 km resolution. A lidar simulator helps to compare consistently model with  
58 observations. The bias in observations due to the satellite under-sampling is first estimated. Then we examine the  
59 vertical variability of both occurrence and properties of clouds. It results that observations indicate a similar  
60 occurrence of low and high clouds over continent, and more high than low clouds over the sea except in summer.  
61 The simulation shows an overestimate (underestimate) of high (low) clouds comparing to observations,  
62 especially in summer. However the seasonal variability of cloud vertical profiles is well captured by WRF.  
63 Concerning inter-annual variability, observations show that in winter, those of high clouds is twice the low  
64 clouds one, an order of magnitude that is well simulated. In summer, the observed inter-annual variability is  
65 vertically more homogeneous while the model still simulates more variability for high clouds than for low  
66 clouds. The good behavior of the simulation in winter allows us to use the 23 years of simulation and 8 years of  
67 observations to estimate the time period required to characterize the natural variability of the cloud fraction  
68 profile in winter, i.e the time period required to detect significant anomalies and trends.

69 *Keywords:* lidar, clouds, simulation, Europe, natural variability

## 70 1. INTRODUCTION

71 Improving our understanding of the internal variability of the European climate is a necessary step to get more  
72 reliable prediction of the temperature evolution in the next two decades over Europe (Hawkins and Sutton 2009).  
73 Actually, using Coupled Model Intercomparison Project (CMIP3) models and despite the fact that these models  
74 may overestimate natural decadal fluctuations, Hawkins and Sutton (2009) show that the climate internal  
75 variability is the most important source of uncertainty to predict the regional temperature in the next two decades.  
76 By comparing different regions of the world, they also showed that this uncertainty is maximal over Europe  
77 compared to other regions. For prediction times of many decades (more than 2 decades), the uncertainty on the  
78 future climate prediction is no more dominated by the lack of knowledge on the inter-annual variability, but by  
79 the uncertainty of the model itself (eg. the difficulty of the model to reproduce the actual state of the atmosphere)  
80 which then becomes the dominant contribution to the uncertainty in the prediction of the future climate at  
81 regional scale. The authors suggest that using observations of the current climate state may help to better  
82 understand the inter-annual variability of the climate system and to reduce model defaults; doing so may  
83 contribute to reduce uncertainties in the prediction of the evolution of the climate at regional scale.

84 At first order, temperature over Europe is driven by the large-scale atmospheric circulation (e.g. Rojas et al. 2013;  
85 Xoplaki et al. 2004). In particular, in winter and summer, European climate is characterized by the succession of  
86 four weather regimes resulting from the relative location of three pressure centers: the Icelandic low, the Azores  
87 high and the continental highs and lows (Yiou et al. 2007; Cassou et al. 2004, 2005; Vautard 1990). Southern  
88 Europe and Mediterranean Basin climates are influenced by the North Atlantic regimes but are also under the  
89 influence of the Genoa depression and the Atlas lee depression (Reiter 1975), which drive air masses from the  
90 south.

91 However, large-scale circulation cannot explain all the variability. According to Yiou et al. (2007) temperature  
92 anomalies are more and more often influenced by other factors. Clouds are one of these factors and play a major  
93 role in the natural temperature variability due to their radiative effects. Chiriaco et al. (2014) showed the spatial  
94 correlation between the area of positive temperature anomaly which occurred during July 2006 over Western  
95 Europe and the lack of low clouds, using satellite observations. Using observations over 1984-2007, Tang et al.  
96 (2012) suggest that clouds over Europe are better indicator of summer maximum temperature changes than the  
97 proxies of soil moisture anomalies, and that the summer temperature increase is correlated with total cloud cover

98 decrease. However, clouds remain an important source of uncertainties in our understanding of climate  
99 variability (Soden and Held 2006) due to the complex cloud feedbacks with surface and boundary layer,  
100 orography and tropospheric environment (air entrainment and humidity). These processes influence the internal  
101 variability of climate response and enhance the model uncertainties. Despite these uncertainties, several studies  
102 using models have investigated the role of clouds on the present and future climate over Europe. They suggest  
103 the increasing of future summer temperature variability particularly for central Europe (Lenderink et al. 2007), a  
104 future annual precipitation increase over northern Europe (Kjellström et al. 2011) and a decrease trend of the  
105 cloud cover in summer over Europe (Tang et al. 2012). Other authors have shown how model biases in present  
106 climate influence the model response to greenhouse gases forcing. For instance, Boé and Terray (2014) show  
107 that climate models with surface temperature strongly sensitive to cloudiness simulate a larger future decrease of  
108 cloud cover than other models (over land, in summer). Lenderink et al. (2007) studied temperature variability  
109 and the surface energy budget over control period (1961-1990) and future climate (2071-2100) in an ensemble of  
110 regional climate models and show that i) temperature variability is overpredicted in the control simulations; ii)  
111 temperature variability is very correlated with surface energy budget variability in the models; iii) clouds and  
112 radiation are critical to determine the climate sensitivity on Western Europe in summer in regional models.

113 The goal of this study is to characterize and understand clouds natural variability in the current climate to better  
114 assess its effect on European climate characteristics (future work). We will focus on the vertical distribution of  
115 clouds because: (i) the clouds radiative effect depend on the vertical distribution of microphysical and macro-  
116 physical properties of clouds (Stephens 2005), and (ii) the vertical distribution of clouds is a much more robust  
117 signature of climate variability than vertically integrated variables (total cloud cover or radiative fluxes)  
118 (Chepfer et al. 2014). Three main questions are then addressed in the current paper: (1) What is the seasonal  
119 variability of the cloud vertical distribution over Euro-Mediterranean area? (2) What is its inter-annual  
120 variability? (3) Are simulations able to reproduce the amplitude of these variabilities?

121 To address these questions we used Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observations  
122 (CALIPSO) that provides very detailed vertical description of clouds' distribution (Winker et al. 2003) since  
123 2006. This study only focuses on nighttime profiles since cloud detection is more accurate than during daytime  
124 (Winker et al., 2009) (see section 2.1). Eight years (since 2006) are *a priori* not sufficient to cover the entire  
125 natural variability. Moreover, the satellite under-sampling of the Europe-Mediterranean area due to the sun-  
126 synchronous orbit could be an issue for covering the cloud variability. Then, we use a simulation in addition to

127 CALIPSO observations. Requirements for the simulation are (i) covering the Euro-Mediterranean region, (ii) a  
128 good spatio-temporal resolution in order to take into account the complex terrain and the influence of the  
129 different air masses that characterize the area of study and which influence the cloud formation, (iii) at least 20  
130 years as it is the period where the internal variability is the principal source of uncertainty in the simulation of  
131 European climate (Hawkins and Sutton 2009), iv) a nearly 'perfect' dynamics to better evaluate and estimate the  
132 fluctuations linked to clouds. A WRF (Weather Research and Forecast Model; Skamarock and Klemp 2008)  
133 simulation performed in the framework of HyMex (HYdrological cycle in Mediterranean EXperiment;  
134 Drobinski et al. 2014) and MED-CORDEX (Mediterranean COordinated Regional climate Downscaling  
135 EXperiment; Ruti et al. 2015) programs is used as it presents these characteristics: 23 years with outputs every 3  
136 hours, 20×20 km<sup>2</sup> horizontal resolution, and the nudging option towards ERA-interim reanalysis for wind,  
137 temperature and humidity above the boundary layer. Also, the same diagnosis of clouds is needed for both  
138 observations and simulation. Since WRF simulates concentration of liquid water, snow and ice which are not  
139 directly comparable with the lidar, a lidar simulator (Chiriaco et al. 2006; Chepfer et al. 2008) developed for  
140 Global Circulation Models (GCMs) in Cloud Feedback Model Intercomparison Project (CFMIP) Observation  
141 Simulator Package (COSIP, Chepfer et al. 2008) has been adapted to WRF and used in this study.

142 These tools are presented in Sect. 2. An important issue for this regional climate variability study is the satellite  
143 under-sampling effect: this issue is addressed in Sect. 3 using two different samplings of the simulation. Also,  
144 the vertical distribution of clouds has been carefully evaluated in GCMs, in particular with the GCM-Oriented  
145 CALIPSO Cloud Product (GOCCP; Chepfer et al, 2008; Cesana and Chepfer 2012) developed to evaluate clouds  
146 representation in GCMs. But it has not been evaluated in regional simulation except for some case studies (e.g.  
147 Chaboureaud et al. 2012 for some convective precipitating clouds over western Europe; Chiriaco et al. 2006 for  
148 some cirrus clouds in the Paris area). Then, the WRF/MedCordex simulation's representation of the clouds'  
149 vertical structure over Euro-Mediterranean area and in particular its ability to reproduce the amplitude of the  
150 seasonal variability is addressed in Sect. 4. In sect 5, the inter-annual variability of cloud profiles is studied in  
151 both observations and simulation: the ability of the model to detect climate extremes, and the question of the  
152 number of years required to cover the entire natural variability are addressed. Conclusions and prospects of this  
153 work are presented in Sect. 6.

154

## 155 2. TOOLS

### 156 2.1. CLOUD PRODUCTS FROM OBSERVATIONS

157 CALIOP is a two-wavelength polarization-sensitive lidar that provides high-resolution vertical profiles of clouds.  
158 This study is based on the 532-nm channel in parallel polarization analysis. This wavelength is in the visible  
159 spectrum making the signal-to-noise ratio reduced during daytime due to solar radiations, affecting the detection  
160 of daytime clouds (Winker et al. 2009). Thus, this study only focuses on nighttime profiles. It corresponds to a  
161 satellite overpass between 23 UTC and 03 UTC for this region. Results can be different for daily clouds, in  
162 particular for low clouds ( $z < 3.2$  km corresponding to  $P > 680$  hPa) that have an important diurnal cycle.  
163 CALIOP's vertical resolution is 30 m from ground to 8.2 km of altitude and 60 m above. Its horizontal  
164 resolution is 330 m. Cloud products used in this study are the GOCCP products. They are specially developed  
165 for comparison with models, in particular GCM. In GOCCP, while the original horizontal resolution of  
166 CALIOP is kept, the original vertical resolution is modified and the profiles are vertically averaged on a GCM  
167 typical vertical resolution of 480m (Chepfer et al. 2010). Here, they are adapted for comparison with regional  
168 model simulation. Two cloud products are defined:

169 -  $SR_{GOCCP}(z)$  is based on Scattering Ratio (SR) values, which highlight the contribution of particles (condensed  
170 water or aerosols) to the lidar signal (Annex 1, eq1). SR is equal to 1 in absence of clouds and aerosols.  $SR > 1$   
171 traduces the existence of particles; it is either aerosols or condensed water. The more optically thick clouds, the  
172 higher SR values. When the lidar signal is fully attenuated by optically thick clouds, the layers below are  
173 obscured and SR values at lower levels become very low ( $< 0.01$ ).  $SR_{GOCCP}(z)$  keeps the GOCCP original  
174 resolution horizontally and vertically (over 34 levels from the ground).

175 -  $CF_{GOCCP}(z)$  is the Cloud Fraction computed from SR profiles at 330 m horizontal resolution over the  $20 \times 20$   
176  $km^2$  horizontal resolution of the simulation (see Sect. 2.2) over the 34 vertical layers. Figure 1 shows the model  
177 grid and the number of CALIPSO nighttime overpasses during June-July-August (JJA) 2008 over each grid  
178 point. With this resolution, some parts are never covered by CALIPSO and the overpass number on the covered  
179 grid-boxes varies between 3 and 6 over one season but the number of observed profiles over one grid point is  
180 increased by the horizontal resolution of 330m. Cloud detection is based on SR thresholds following Chepfer et  
181 al. (2008, 2010):  $0.01 < SR \leq 1$  clear,  $1.2 < SR < 5$  (existence of particles, could be optically thin clouds or

182 aerosols), and the threshold of cloud detection is set to 5 ( $SR \geq 5$ ). These thresholds are used for each profile at  
183 each vertical level (Fig. 10 annex 1). The cloud fraction over a gridbox is then the percentage of cloudy profiles  
184 ( $SR(z) \geq 5$ ) by the total number of profiles that are not fully attenuated ( $SR(z) \geq 0.1$ ).

185 In the current study,  $SR_{GOCCP}$  and  $CF_{GOCCP}$  profiles observed throughout the period 06/2006 to 12/2011 are used,  
186 as it is the common period with the simulation (Sect. 2.2).

## 187 2.2. CLOUD PRODUCTS FROM SIMULATION

### 188 2.2.1. SIMULATION SET-UP

189 A 23-year simulation with the WRF model of the National Center for Atmospheric Research (NCAR)  
190 (Skamarock and Klemp 2008) was performed at  $20 \times 20$  km<sup>2</sup> resolution over the Mediterranean basin in the  
191 framework of COordinated Regional climate Downscaling Experiment (CORDEX; Giorgi et al. 2009. Note that  
192 MED-CORDEX is the Mediterranean focus of CORDEX; Ruti et al. 2015) and HYMEX programs (Drobninski et  
193 al. 2014). It performs a dynamical downscaling of the ERA-interim data (Dee et al. 2011) over the period 1989-  
194 2011. The model has 28 sigma-levels in the vertical. A complete set of physics parameterizations is used: the  
195 Single-Moment 5-class microphysical scheme (WSM5; Hong et al. 2004), the Kain-Fritsch convection scheme  
196 (Kain 2004), the Yonsei University (YSU) planetary boundary layer (PBL) scheme (Noh et al. 2003) and a  
197 parameterization based on the similarity theory (Monin and Obukhov 1954) for the turbulent fluxes. The  
198 radiative scheme is based on the Rapid Radiative Transfer Model (RRTM) (Mlawer et al. 1997) and the Dudhia  
199 (1989) parameterization for the longwave and shortwave radiations, respectively. The lower boundary conditions  
200 of the WRF model are provided by the land surface model (LSM) Rapid Update Cycle (RUC) (Smirnova et al.  
201 1997, 2000). Nudging above the boundary layer is used in order to avoid the small-scale variations to generate a  
202 large-scale that would diverge from the observed large-scale structures inside the limited area (Stauffer and  
203 Seaman 1990; Salameh et al. 2010; Omrani et al. 2013). Hence it allows us to compare the structure of clouds  
204 with the same large-scale environment.

205 This simulation has been already used for several studies and has been evaluated against data for temperature  
206 (e.g Chiriaco et al. 2014; Stefanon et al. 2014), precipitation (e.g Flaounas et al. 2013, 2014; Lebeaupin-Brossier  
207 et al. 2015; Vaittinada et al. 2015) and wind (Omrani et al. 2014).

208 Cloud outputs are mixing ratios of ice, snow and liquid and are interpolated on the same vertical resolution as

209 GOCCP between the ground and 16 km, which corresponds to 34 levels. They are available every 3h.

### 210 2.2.2. LIDAR SIMULATOR AND DATASETS

211 To compare WRF outputs with CALIPSO lidar observations, we use a methodology similar to the one followed  
212 by Chepfer et al. (2008) and Cesana and Chepfer (2012) comparing CALIPSO observations with a GCM. Also  
213 we define two datasets to better assess the issue of sampling. The methodology then consists in i) extracting two  
214 different datasets of WRF vertical profiles: one that uses the total grid of the simulation and another one using  
215 the satellite horizontal and temporal resolutions (see details below) ii) from these extracted profiles, computing  
216 the lidar profiles that would be observed by CALIPSO if the satellite were flying above the simulated  
217 atmosphere (Chepfer et al. 2008), iii) using the same SR bins as the ones used for GOCCP, iv) computing the  
218 cloud fraction from the simulated SR profiles as the percentage of SR ( $z$ )  $\geq 5$  per grid-box as done to produce  
219  $CF_{GOCCP}$  (see Sect. 2.1.).

220 The first dataset is obtained by extracting one profile per night, at 00 UTC, over each grid box. It corresponds to  
221 about  $2.10^6$  profiles for one season. The covered period is 1989-2011. The second dataset corresponds to the  
222 profiles that are coincident with CALIPSO tracks, at the closest time of CALIPSO overpass (i.e either 00 or 03  
223 UTC). As CALIPSO horizontal resolution is 330 m while the WRF one is 20 km, the number of CALIPSO  
224 profiles that fall into one WRF grid-box varies between 0 (Fig. 1a) and 85 (depending on the fact the track  
225 crosses the box along its diagonal or only at its corner or not at all). The WRF profiles are duplicated to obtain a  
226 CALIPSO like track with the same number of CALIPSO profiles over a grid-box, to give to each grid box the  
227 corresponding weight. However, we have tested the extraction of cloud profiles without duplicating the profiles  
228 (using only one WRF profile per grid-box that is crossed over) and the differences are very negligible for the  
229 current study. Over one season, the number of profiles of this second dataset is about  $8.10^7$  (obtained from  $5.10^4$   
230 profiles, before duplicating). The available period is 2006-2011.

231 A lidar simulator is associated to these WRF outputs. It consists on computing the lidar signal (equation 2 in  
232 annex 1) from WRF outputs considering the effective radius  $r$  and size distribution  $n(r, z)$  of each meteor (here  
233 liquid, ice and snow). Since these parameters are not direct outputs of the model and since the simulator is  
234 offline the model, they are computed from the mixing ratio of ice ( $Q_i$ ), snow ( $Q_s$ ) and liquid ( $Q_l$ ) (outputs of the  
235 model) using the microphysical equations used in the parameterization (here WSM5). WRF outputs ( $Q_i$ ,  $Q_l$ ,  $Q_s$ ,  
236 pressure and temperature) are interpolated on a regular vertical grid first and then the lidar signal is computed

237 giving SR profiles for the two datasets. Afterwards, cloud fraction profiles are computed following the same  
238 method as for observations (Sect. 2.1). The variables obtained are  $SR_{WRF+sim}^T(z)$  and  $CF_{WRF+sim}^T(z)$  for the first  
239 dataset and  $SR_{WRF+sim}(z)$  and  $CF_{WRF+sim}(z)$  for the second dataset, both over period 2006-2011 (Tab. 1). The  
240 exposant T is for the total simulation dataset.

241 Figure 2a is a randomly picked scene showing an example of the vertical profiles of  $SR_{GOCCP}(z)$  along a  
242 CALIPSO track (2009-01-19 at night) between 30°N and 50°N latitudes and 3°E and 5°E longitudes. The same  
243 scene is represented in Fig. 2b with the simulated total condensed water mixing ratios, while Fig. 2c shows  
244  $SR_{WRF+sim}(z)$ . The red shadows in Figs. 2a-c show the cloud structures, the dominant blue represents the clear sky  
245 and the dark blue indicates that the signal is fully attenuated. As the model does not simulate aerosols, the  
246 boundary layer is not properly represented in the simulated profiles (Figs. 2b and c).

247 Clouds detected by CALIOP shown in Fig. 2a are simulated by the model most of the time (Fig. 2.b and c)  
248 despite few differences (around 32°N, around 49°N). In this case study, the vertical cloud structure is less  
249 variable in the simulation than in the observations. This result of bigger and more persistent clouds in terms of  
250 occurrence and lifetime in regional simulation over Europe is expected and has been shown in previous studies  
251 with mesoscale models: e.g the evaluation of some cirrus clouds case studies (1-2 months) of the fifth-generation  
252 Pennsylvania State University-NCAR Mesoscale Model (MM5) with ground-based measurements over France  
253 (Chiriaco et al. 2006) and over Europe with Ice, Cloud and land Elevation Satellite (IceSat) lidar observations  
254 (Chepfer et al. 2007).

255 The comparison between Fig. 2b and c illustrates the importance of comparing lidar observations with a similar  
256 simulated signal:

- 257 - the lidar signal can be fully attenuated, masking lower clouds even if they are simulated: for instance at  
258 43°N, between 5 and 7 km, high mixing ratio are simulated, while  $SR(z)$  value is less than 0.01.
- 259 - a weak mixing ratio may be associated to significant SR value (e.g. between 40 and 41°N or between 46  
260 and 48°N). It's due to the fact that  $SR(z)$  is a signature of the optical depth and optical properties of  
261 clouds (type of particles, size, concentration), and not only of the particles concentration (Chepfer et al.  
262 2013).

263 In the current study, these tools are used to characterize the seasonal and interannual variabilities of the vertical  
264 structure of clouds over the Euro-Mediterranean region. As shown by Fig. 1, the orbit of the satellite does not  
265 allow us to evaluate the spatial variability of clouds at these spatial and time scales.

266 Before using the tools for inter-annual variability studies, two steps are necessary: (i) evaluate the biases caused  
267 by the satellite under-sampling (Sect. 3), and ii) evaluate the biases of the modeled clouds (Sect. 4).

268

### 269 3. BIAS DUE TO SATELLITE UNDER-SAMPLING

270 The number of satellite tracks over one season is limited in a grid-box. Figure 1a shows that CALIPSO does not  
271 overfly all the grid elements at  $20 \times 20$  km<sup>2</sup> resolution, and that the maximum number of overpasses over one  
272 grid-box is 6 in one given season. To study cloud variability at this spatial scale and resolution, it is then  
273 necessary to cumulate enough profiles temporally or spatially. This section aims at quantifying the bias in the  
274 cloud variability due to the satellite under-sampling, depending on the number of seasons cumulated.

275

#### 276 3.1. CLOUD FRACTION PROFILES

277 The effect of satellite under-sampling on cloud fraction profiles is examined here by using only the simulation,  
278 as a complete sampling is by definition not available in observations. Comparing complete sampling and satellite  
279 sampling of the same dataset (here the simulation) will allow only focusing on biases due to the sampling (so not  
280 also the biases of the simulation). Black lines (computed over around  $2 \cdot 10^6$  profiles) in Fig. 3 represent  
281  $CF_{WRF+sim}^T(z)$ , cloud fraction profiles extracted every 00 UTC over the entire domain, i.e one profile per grid box.  
282 They are compared to the red lines (computed over  $5 \cdot 10^4$  profiles) representing  $CF_{WRF+sim}(z)$ , cloud fraction  
283 profiles following the CALIPSO sampling, i.e either 0 profile over a grid box or several profiles duplicated to  
284 get the same number of WRF profiles than CALIPSO profiles in this grid box. Cloud fraction profiles are  
285 computed from the WRF/MED-CORDEX simulation using the lidar simulator (Sect. 2.2.b) and averaged over  
286 the domain and over the years (2006-2011) for each season.

287 For the four seasons (Fig. 3), CALIPSO sampling does not generate real bias on lidar cloud fractions,  
288  $CF_{\text{WRF+sim}}(z)$  being very similar to  $CF_{\text{WRF+sim}}^T(z)$ . Note that the differences are only observable for high levels ( $z >$   
289 6.5 km corresponding to pressure  $P < 440$  hPa) because of the small cloud fraction values below 6 km (Fig. 3).  
290 This aspect will be discussed in section 4. For high levels, the maximum errors of cloud fraction for one layer are  
291 around 3 % over one season (Tab. 2).

292 The absolute value of the relative errors averaged vertically over high levels are smaller in winter (5%) and  
293 spring (7%) than in fall (9%) and summer (21%). The errors may seem large in fall and summer; it is due to the  
294 small vertical shift of the profiles, which enhances the relative difference between the two profiles at levels  
295 where the cloud fraction is not maximal. The maximum of difference in cloud fraction is approximately the same  
296 for the four seasons and less than 5%. This shift of high clouds level due to satellite under-sampling may be  
297 linked to the occurrence of deep convection happening in summer and fall inducing local mesoscale clouds  
298 (Funatsu et al., 2009) that are more likely to be missed by the satellite because of their short lifetime and small  
299 spatial extent (Rysman et al., 2013).

300 For mid and low levels, the values of cloud fraction differences between  $CF_{\text{WRF+sim}}(z)$  and  $CF_{\text{WRF+sim}}^T(z)$  are very  
301 small (around 0.1 %). The average relative errors at these levels remain weak though: 6% in fall, 9% in spring  
302 and 11% in summer. Notice that in winter, the biases for mid-levels reach 13% related to a vertical shift at 6 km.  
303 The origin of the underestimate of low and mid clouds is discussed in the next section.

## 304 3.2. HISTOGRAMS OF SCATTERING RATIO

305 To go further in details and have an idea about the physical properties of clouds overestimated/ missed with the  
306 CALIPSO under-sampling, 3D SR histograms are analyzed (Fig. 4). These histograms provide detailed vertical  
307 information on cloud optical and physical properties. Two simulated SR histograms are compared: (1) the  
308 distributions of  $SR(z)$  occurrence for each altitude and SR bin when cumulating SR profiles following CALIPSO  
309 tracks ( $SR_{\text{WRF+sim}}(z)$ , Fig 4a); (2) the same but with the full resolution of the simulation ( $SR_{\text{WRF+sim}}^T(z)$ , Fig 4b).

310 In each histogram, the first bar is the percentage of fully attenuated profiles ( $0 \leq SR(z) < 0.01$ ) and the second  
311 bar corresponds to clear sky profiles ( $0.01 \leq SR(z) < 1.2$ ). The 3<sup>rd</sup> and 4<sup>th</sup> bars are for unclassified profiles ( $1.2 \leq$   
312  $SR(z) < 5$ ). The subsequent bars are for cloudy profiles ( $SR(z) \geq 5$ ). Figures 4a and b are very similar: great

313 attenuation below 7 km, many high clouds with  $5 \leq SR(z) < 20$  and some low level clouds with  $40 \leq SR(z) < 60$ .  
314 The similarities show that the total SR distribution is not significantly affected by the sampling of the satellite.

315 Figure 4c shows the absolute difference between Fig. 4a and 4b. In average, the differences due to CALIPSO  
316 sampling for  $1.2 \leq SR(z) < 20$  below 7km and for the  $SR(z)$  values  $\geq 20$  (all altitudes) are under 0.25% (grey  
317 shade). For  $5 \leq SR(z) < 20$ , the differences at high levels are around 2%. The largest differences occur for the  
318  $0.01 \leq SR(z) < 3$  (clear sky and unclassified profiles) at altitudes between 12 and 14 km.

319 Figure 4c is then separated onto seasons (Fig. 4d to f). Focusing on one season leads to greater biases due to  
320 under-sampling, as fewer profiles are cumulated. Nevertheless, differences remain very low whatever the season  
321 as for the total SR histogram, and most differences are detected whether for high levels or mid-levels in winter  
322 and spring. The greatest errors in terms of magnitude (for values passing the threshold of cloud detection  $SR(z)=$   
323 5) are detected in fall and summer consistently with Fig.3 but the differences appear on more levels in spring and  
324 winter (than fall and summer) in cloud fraction profiles (Fig. 3). Consequently, the biases of sampling detected  
325 in cloud fraction profiles (sect. 3.1) correspond mostly to SR values between 5 and 20. Also, for high clouds  
326 (above 8 km) the errors are maximal in fall in Fig.4 while it was summer in Fig 3: this is linked to the definition  
327 of the cloud fraction (cf. sect. 2.1.) explaining that the cloud fraction depends on cloudy profiles but also on fully  
328 attenuated profiles. Between 8 and 10km, there are more differences of fully attenuated profiles in fall than in  
329 summer (signal drowned in the grey shade): this explains why the errors of cloud fraction are bigger in summer  
330 even though SR shows more differences of cloudy profiles in fall. The overestimate and underestimate of SR  
331 values due to CALIPSO under-sampling depend on the levels. Around 9-10 km, where the maximum of clouds  
332 is detected, the satellite under-sampling always leads to an overestimate of SR values (as the cloud fraction in  
333 Fig3).

334 Only simulation is used in this section, despite the fact it very likely contains biases in the representation of  
335 clouds. The estimation of these biases is the purpose of the next section, and in particular concerning the  
336 seasonal cycle of clouds. Further computations of sampling errors that would have been estimated by  
337 observations are presented in table 3 and discussed in the conclusion.

338

## 339 4. CHARACTERIZATION OF THE SEASONAL CYCLE AND 340 ABILITY OF THE MODEL TO REPRODUCE IT

341 For meso-scale models, cloud biases have been detected in previous studies, but only for some case studies  
342 (Chiriaco et al. 2006; Chepfer et al. 2007; Chaboureau et al. 2002). The aim of this section is to provide an  
343 evaluation of clouds over the entire Euro-Mediterranean region and throughout several years of observations.  
344 Hence, both the mean annual and seasonal biases of the vertical distribution of clouds and the difference of  
345 amplitude of the seasonal cycle will be assessed. The average annual cycle is studied using observations and the  
346 simulation over the period June 2006 to December 2011. Only the simulation along the satellite tracks is used  
347 here in order to be consistent with observations. The seasonal variability of cloud fraction profiles is studied  
348 separately over continental Europe and over the Mediterranean Sea. Using land/sea mask, continent stands for  
349 the grid-boxes over continental Europe in the blue rectangle in Fig.1a while sea stands for the grid-boxes located  
350 over the Mediterranean Sea in the red rectangle of the same figure. Note that we also studied the variability over  
351 the Atlantic Ocean and as the results were very similar to the continent they are not shown in this paper.

352

### 353 4.1. CLOUD FRACTION PROFILES

#### 354 4.1.1. SEASONAL CYCLE

355 In this sub-section, only observations are examined to describe the seasonal cycle of vertical distribution of  
356 clouds. Pink solid lines in Fig. 5a shows  $CF_{GOCCP}(z)$  averaged over the continent from 2006/06 to 2011/12.  
357 Figure 5b is same as Fig.5a but averaged over the sea. Notice that in our case, continent is located between 40°N  
358 and 52°N while the Mediterranean Sea is between 30° and 42°. The Mediterranean Sea is under the influence of  
359 both mid latitude synoptic disturbances that affects the European climate (in particular winter; Cassou et al. 2004,  
360 Yiou et al. 2007) and the subtropical storms (Rysman et al. 2013). The average profile from observations shows  
361 that:

362 - the maximum of cloud fraction occurs around 9km (Fig. 5a and b).

- 363 - at a given level, the cloud fraction is greater over the continent than over the sea (Fig. 5a and b). This  
364 might be explained by the fact that frontal (ascending large scale motions) clouds and stratocumulus  
365 clouds (light subsiding motions and near surface instability) are more frequent over Europe (land) than  
366 over Mediterranean Sea (Cheruy and Aires 2009).
- 367 - over the continent, low clouds occurrence is equivalent to high clouds occurrence, while over the sea  
368 high clouds are a little more frequent than low clouds (Fig. 5b). This might be linked to a lack of  
369 stratocumulus and stratus clouds over the Mediterranean (Cheruy and Aires 2009).

370 In Fig. 5a and b, the pink shades represent the range of seasonal variability, when computing a profile by season,  
371 averaged from 2006 to 2011. The purpose of this representation is to estimate the envelope of variability of the  
372 vertical profile during a complete annual cycle (in observations), and evaluate its representation in the simulation  
373 (Sect. 4.2). The amplitude of the observed seasonal variability shows that the variability of high clouds is  
374 equivalent to the variability of low clouds, over the continent as well as over the sea. This seasonal amplitude is  
375 estimated to 5 % in terms of cloud fraction for high and low levels.

376 In Fig. 5c and d, the shades are envelopes containing 4 profiles of relative anomaly of each season comparing to  
377 the profile averaged over the 4 seasons and 6 years. It shows that over the continent, the observed relative  
378 variability ranges between -40% and 50% of the mean annual cloud fraction profiles (Fig. 5c). This relative  
379 anomaly has equivalent amplitude for the different vertical levels. Over the sea the amplitude of seasonal  
380 variability is greater for low levels (more or less 100 %) than for mid and high levels (more or less 60%) (Fig.  
381 5d).

382 To go further in details, the four seasons averaged over 2006 to 2011 are plotted separately in Fig. 5e (continent)  
383 and Fig. 5f (sea) for  $CF_{GOCCP}(z)$  showing that for both continent and sea, the high clouds in summer and fall (red  
384 and light blue lines) are less frequent (especially summer  $CF_{GOCCP}(z)$  around 5%) and occur at higher altitude  
385 ( $\approx 10$ km) than high clouds in winter and spring ( $\approx 8$  km with  $CF_{GOCCP}(z)$  around 9%). This result is expected  
386 because of the altitude of the tropopause that is minimal at the end of winter and maximal at the end of summer  
387 (Appenzeller et al. 1996) and it affects clouds top height (Gettelman and Forster 2002). The figures also show  
388 that the cloud fraction is weak in the mid-levels for the 4 seasons in both continent and sea and particularly in  
389 summer over the sea. The low clouds are frequent in winter over both continent and sea. Summer and fall  
390 profiles show a second maximum at low levels, even if less pronounced than winter, while spring profiles show  
391 more homogeneity from ground up to 6-7 km.

392 The great relative variability of  $CF_{GOCCP}(z)$  over the sea shown in Fig 5d around 2-3 km is explained by the  
393 small values of  $CF_{GOCCP}(z)$  in summer while they are large in winter (Fig. 5f). Note that winter clouds are under  
394 the large scale influence while summer clouds are under mesoscale influence (air-sea fluxes, topography)  
395 (Chaboureau and Claud 2006).

396

#### 397 *4.1.2. SIMULATED SEASONAL CYCLE*

398 The seasonal vertical cloud structure has been characterized with CALIPSO observations. The goal of this sub-  
399 section is to evaluate this structure in the simulation and focus on the similarities and the differences with the  
400 observed seasonal variability.

401 As for observations, the average simulated cloud fraction profile is maximal around 9km (blue line in Fig 5a and  
402 b). Nevertheless, in average, the simulation over-estimates high cloud fraction (above 6 km), for both continent  
403 ( $CF_{WRF+sim}(z=9 \text{ km}) = 30\%$  while  $CF_{GOCCP}(z=9 \text{ km}) = 10\%$ ) and sea ( $CF_{WRF+sim}(z=9 \text{ km}) = 20\%$  while  
404  $CF_{GOCCP}(z=9 \text{ km}) = 7\%$ ); the difference of the magnitude of the maximum of high clouds between observations  
405 and the simulation is larger over the continent than over the sea.

406 Under 6 km, the cloud fraction is under-estimated in the simulation over both areas. The fact that WRF  
407 overestimates the high clouds probably enhances the underestimate of low and mid clouds due to the exacerbated  
408 attenuation of simulated lidar profiles. Nevertheless, another study based on comparisons of the same simulation  
409 but with ground-based lidar located near Paris (France) shows that the model actually under-estimates the  
410 amount of low clouds, in particular in summer (Bastin et al., under review). Figures in Annex 2 also confirm that  
411 this underestimate of low clouds is not only due to lidar attenuation: the map of low clouds directly computed  
412 from the total condensed water in the simulation (hence without the lidar simulator) shows that they are almost  
413 absent in summer (less than 10%). The wrong vertical distribution of cloud layers has already been noticed with  
414 a mesoscale model for some case studies (Chaboureau et al. 2012) and the under-estimation of low clouds in  
415 summer is also a known result for GCMs in general over mid-latitudes (Cheruy et al. 2013).

416 Despite this bias, the range of seasonal variability of  $CF_{WRF+sim}(z)$  (blue shade) is by the same order of magnitude  
417 than the range of seasonal variability of  $CF_{GOCCP}(z)$  over the continent (Fig. 5a), but is larger over the sea for the

418 high clouds and narrower for mid and low clouds (Fig. 5b). It is confirmed by the relative seasonal variability  
419 profiles (blue shades in Fig. 5c and d), showing that:

420 - Above 9 km, the simulation and the observations have equivalent amplitude of relative variability (more  
421 or less ~50%): even if the result is true over both continent and sea, the simulated variability at this  
422 altitude is closer to the observed variability over the sea.

423 - For mid and low clouds, the simulated amplitude of the relative variability over continent is twice  
424 greater than the observed amplitude (more or less 100% against 50%). Over the sea, the simulated  
425 relative variability is greater than observed but the magnitude is by the same order.

426 Both observations and simulation agree that:

427 - The relative range of variability of mid and low clouds is greater than the relative range of variability of  
428 high clouds: In the simulation, this could be due whether to real seasonal variability of clouds in the  
429 model or a bias due to the very weak amount of clouds at low and mid levels. Actually, a great relative  
430 anomaly could be due to a great absolute anomaly but also to a weak cloud fraction.

431 - Mid and low levels variability is greater over the sea than over the continent, especially due to summer  
432 differences.

433 Analyzing separately the four seasons (Fig. 5 g and h) shows that despite the overestimate of high clouds and the  
434 underestimate of low clouds, some characteristics of the observed profiles are well simulated by the model over  
435 both areas: (i) a maximum at 10 km for summer and fall and at 8 km for winter and spring; (ii) less mid clouds  
436 than high clouds over both areas (Fig. 5 e and f), with an almost zero  $CF_{WRF+sim}(z)$  at these levels (except for  
437 winter where it's around 2.5 %); (iii) a second maximum at low levels except in spring. This maximum is less  
438 pronounced in simulation than observations but exists except over the sea in summer: Fig. 11 d and e (Annex 2)  
439 show that in summer no mid or low clouds are represented over the sea. This figure also gives an idea of the  
440 spatial variability of occurrence of cloud layers from the simulation, even if not evaluated against observations.

## 441 4.2. HISTOGRAMS OF SCATTERING RATIO

442 Section 4.1. showed that the simulation over-estimates the occurrence of high clouds and under-estimates  
443 occurrence of low clouds, in particular in summer, but reproduces consistently the seasonality of clouds  
444 occurrence at each level and the amplitude of relative seasonal variability over the sea (Fig. 5d). It is now

445 necessary to understand which clouds, in term of physical properties, are simulated for each season, in particular  
446 because the occurrence of clouds is not enough to evaluate the properties of clouds that can also present a  
447 significant variability. Since SR values are linked to the optical depth and optical properties of the cloud particle  
448 (type of particle, size, and concentration) (Chepfer et al. 2013), histograms of  $SR(z)$  are computed (Fig. 6)  
449 following the same method as in Fig. 4a-b, by cumulating SR profiles from 2006 to 2011 in each season (each  
450 line in the figure).

451 We first focus on  $SR(z)$  values greater than 5, passing the threshold of cloud detection. Both simulation and  
452 observations agree that the SR occurrence is dominant between 6 and 12 km and for  $5 < SR(z) < 20$  at all  
453 seasons. The overestimate of simulated high clouds (Sect. 4.1.2.) is mostly an overestimate of high clouds with  
454 low  $SR(z)$  values ( $SR(z) < 20$ ), and the occurrence of  $5 \leq SR_{WRF+sim}(z) < 20$  is around 7% when occurrence of  
455  $SR_{GOCCP}(z)$  is around 2%. In the observations, low clouds are dominated by strong  $SR(z) (>60)$ , a cloud category  
456 which is entirely missed by the simulation, whatever the season. Those clouds probably correspond to stratus  
457 clouds that appear mostly over Atlantic Ocean and continental Europe and stratocumulus clouds that appear  
458 particularly over Atlantic ocean but also over Europe and Mediterranean (Cheruy and Aires, 2009). Stratus  
459 clouds are formed when moist air near ground level starts to condensate (Khvorostyanov, 1995) while  
460 stratocumulus clouds are driven by convection and associated most of the time to strong temperature inversion at  
461 the top of the boundary layer (Cheruy and Aires, 2009). For these low clouds, only clouds with  $40 \leq SR(z) < 60$   
462 are simulated. For mid-levels, the  $SR(z) > 60$  are also missed by the simulation.

463 The first bar of each SR histogram ( $SR(z) \leq 0.01$ ) represents the full attenuation of the lidar. The number of fully  
464 attenuated profiles is greater in the simulation and the attenuated profiles occur at higher levels than the observed  
465 ones, whatever the season. This is consistent with the overestimate of high clouds and suggests that even if it's  
466 not the main reason explaining the lack of low and mid clouds, the overestimate of fully attenuated profiles  
467 increases the low and mid clouds deficit in the simulation using the lidar simulator. Also note that SR values for  
468 high clouds in the simulation are less than 20: so, the attenuation is mostly induced by the vertical extent of  
469 clouds as the optical depths of each layer are additive: typically  $SR(z) = 5$  corresponds to an optical depth of  
470 0.07 for a cirrus of 1 km depth at 10 km (Chepfer et al. 2013). This means that for clouds of 5 km depth, the  
471 optical depth is around 0.35.

472 This histogram representation also confirms that despite the differences between simulation and observations,  
473 the seasonality of  $SR(z)$  distribution is respected in the simulation:

- 474 - In winter (Fig 6a and b): the simulation creates, in comparison to other seasons, a large amount of low  
475 thick clouds and some thinner clouds, even if they are underestimated. The large amount of high clouds  
476 with  $SR(z) < 20$  is also simulated.
- 477 - Spring (Fig. 6c and d) and fall (Fig. 6g and h) histograms are very similar, with many high and optically  
478 thin clouds in both observations and simulations. The difference between spring and fall, spotted by  
479 observation as well as simulation, is that in spring there are more mid optically thin clouds.
- 480 - In summer, the minimum of mid clouds observed with GOCCP (Fig. 6e) in comparison with other  
481 seasons is also simulated.

482 To summarize, the simulation overestimates high clouds with low  $SR(z)$  values whatever the season over  
483 different layers (more than 6 layers with our vertical resolution) and this leads to a lot of attenuation of the  
484 simulated lidar signal. Low clouds as detected by a lidar are underestimated in the simulation, in particular the  
485 low very thick clouds that are present in the observations: it is both a consequence of the over-estimation of the  
486 fully attenuated profiles and a real underestimate of low clouds as it was explained in Sect. 4.1. However, the  
487 seasonality of  $SR(z)$  histograms is well captured by the model.

488 As expected, the simulated clouds' vertical structure showed high differences with the vertical structure observed  
489 by CALIPSO. This highlights the importance of the vertical structure: the overestimate of high clouds would  
490 compensate with the underestimate of low clouds when computing cloud cover and the model biases would have  
491 been smaller (computing the vertically averaged cloud fraction: simulation 5%, observations 3.7% over the sea  
492 and 8% versus 5.6% over the continent). A new important result is the ability of the model to simulate the  
493 amplitude of the seasonal variability of the cloud distribution (vertical distribution and  $SR(z)$  values distribution).  
494 This allows us to use these datasets (simulation and observations) to address the issue of the inter-annual  
495 variability of cloud vertical profile.

## 496 5. INTER-ANNUAL VARIABILITY

### 497 5.1. AMPLITUDE OF INTER-ANNUAL VARIABILITY

498 The goal of this section is to study the inter-annual variability of the vertical cloud structure over Europe with  
499 observations and evaluate the simulation's ability to reproduce it over the 6 common years. Shades in Fig. 7a

500 (continent) and b (sea) show the standard deviation (STD) of the 2006 to 2011 winter cloud fraction profiles for  
501 both simulation and observations. Figures 7c and d are the same but for summer. This STD calculation is an  
502 estimation of the inter-annual variability in a given season. The altitude where the observed variability (pink  
503 shade) is maximal or minimal is well reproduced by the simulation (blue shade), for winter over both continent  
504 and sea and for summer over the sea.

505 In winter, the range of inter-annual variability of the observed cloud fraction is greater for high levels than for  
506 mid and low levels (Fig. 7.a and b). The smaller variability of low clouds over the sea and the continent is not  
507 related to a smaller occurrence of clouds since Fig. 5e and f indicate that the cloud fraction of high clouds is  
508 equivalent to the cloud fraction of low clouds for both continent and sea, in the observations. The simulation  
509 well reproduces the behavior of the observed inter-annual variability with greater amplitude for the high levels.

510 In summer (Fig 7 c and d), the amplitude of the observed inter-annual variability shows a different behavior from  
511 winter. Over the continent, mid-level cloud occurrence is variable from one year to another, such as for low  
512 clouds. High clouds occurrence is more stable from one year to another. Over the sea, the high cloud occurrence  
513 is a little bit more variable than the mid and low ones. Simulation does not well reproduce this variability: the  
514 variability is large at high levels only for both areas. The weak variability of simulated low clouds is probably  
515 due to the small cloud fraction simulated in summer (Fig. 5g and h, red profiles).

516 Anyway, the simulation always over-estimates the range of inter annual variability for high clouds, whatever the  
517 season and the area (Fig. 7a to d). This over-estimation is enhanced when considering the total envelope of  
518 variability of clouds (maximums and minimums for each layer) traducing the behavior of extreme values (solid  
519 lines in Fig. 7e, f, g and h) instead of the mean one (shades in Fig. 7). For high clouds, the inter-annual  
520 variability is very large in the simulation, and the extreme events in terms of cloudiness are intense in winter and  
521 in summer over the continent. The relative range of the total variability of high clouds is around 25 to 50%  
522 greater than the STD in the simulation (comparing the solid blue line with the blue shade) versus 30 to 60%  
523 greater for observations (comparing the solid pink line with the pink shade) over both continent and sea for  
524 winter and summer.

525 Extremes are less intense at low and mid-levels, in particular in summer for the simulation where there are  
526 almost no clouds.

527 To compare this variability with the uncertainty linked to the satellite sampling,  $CF_{WRF+sim}^T(z)$  (i.e. with the  
528 complete simulation sampling) STD is also plotted in black line in Fig. 7a and b. The inter-annual variability of  
529 the  $CF_{WRF+sim}^T(z)$  has the same behavior than the inter-annual variability of  $CF_{WRF+sim}(z)$  with some differences  
530 that depend on the levels, the season and the area. In winter, the relative errors of estimation of the amplitude of  
531 inter-annual variability of high clouds are around 15% over the continent and 20% over the sea where it's more  
532 dependent on the altitude; the maximum of error is reached at 8km (50%). In summer, the variability is strongly  
533 modulated by local events (storms and mesoscale convective systems). These events, occurring typically at  
534 spatial scales ranging between 10 km to 100 km during less than 3 hours (Rysman et al. 2013), are more easily  
535 missed by CALIPSO considering its undersampling. It generates significant differences of variability between  
536 the two samplings (Fig. 7c and d), the total one being less variable than the CALIPSO one. It is then more  
537 questionable to study the inter-annual variability of clouds in summer from observations at this spatial resolution  
538 of 20x20 km<sup>2</sup>.

## 539 5.2. CLOUDS NATURAL VARIABILITY IN WINTER

540 The purpose of this section is to analyze the 8 years of observations and 23 years of simulation taking into  
541 account the model biases evaluated previously to try to estimate the natural variability of clouds and how much  
542 these respective datasets can be used i) to detect extremes and ii) as a referent period to detect possible trends.  
543 The simulation's full resolution (one profile per night extracted at 00UTC for each grid-box) is used. Only  
544 winter is considered: actually Sect.5.1 showed that the inter-annual variability is better simulated in winter than  
545 in summer, and is less affected by the satellite sampling.

546 Figure 8a shows the  $CF_{WRF+sim}^T$  anomaly simulated profiles averaged over the Mediterranean Sea for every  
547 winter from 1990 (December 1989, January and February 1990) to 2011 (December 2010, January and February  
548 2011) compared to winter mean 2007-2011 (blue profile in Fig. 5h; this period is used as it is the common period  
549 for observations and simulations, which is important when computing anomalies). Figure 8b is the same as Fig.  
550 8a for observations for each winter with available observations from 2007 to 2013. The anomaly is compared to  
551 winter mean 2007-2011 (pink profile in Fig. 5f). Fig. 8c and d are the same as Fig. 8a and b over the continent.

552 Despite the model biases leading to an overestimate of high clouds and an underestimate of mid and low clouds  
553 and despite the biases due to satellite under-sampling in winter (cf. 5.1), the sign of observed anomalies are most  
554 of the time reproduced by the simulation during the 5 years and at the right altitudes: the 2008 strong negative  
21

555 anomaly of high clouds over both sea and continent, the 2011 positive anomaly of high clouds over the sea and  
556 mid clouds over the continent and also the 2010 strong positive anomaly signal of high clouds over both  
557 continent and sea that are detected with CALIPSO are also simulated by the model. Some observed anomalies  
558 are poorly simulated by the model, particularly over the continent. The obvious ones are: the positive anomaly of  
559 2007 observed at 8 km (Fig. 8.d) that is negative with the simulation (Fig. 8.c) and the almost null observed  
560 anomaly of 2009 around 8km is simulated as a strong positive anomaly. Also, the variability of the maximum  
561 altitude of the anomalies are most of the time well reproduced by the model (e.g. in 2010 over the continent,  
562 altitude max= 10km, 2009 low clouds altitude max = 2km) and sometimes not (altitude max of high clouds  
563 anomaly over the sea in 2007 = 12km for observations and 9km in the simulation). Some other particular  
564 anomalies are noticed over the 22 years: e.g. the 1996<sup>th</sup> high clouds signal but at a different altitude than in 2010  
565 over the sea, the 1991-1993 enhanced low clouds signal also over the sea, the 1994 high clouds signal over the  
566 continent.

567 Some high cloud strong anomalies only appear over the continent (high clouds positive anomaly in 1994) or over  
568 the sea (negative anomaly of low clouds in 2007) while others appear over both areas (2010 high clouds  
569 anomaly). This highlights the importance of the spatial distribution of clouds that is not very discussed in this  
570 paper because it cannot be seriously evaluated. However, the separation between sea and continent is useful.

571 Winter 2010 shows a strong anomaly of high clouds occurrence, even when compared to the 22-year-time series  
572 (Fig. 8a and c). The horizontal map of high cloud fraction of winter 2010 anomaly relative to the cloud fraction  
573 averaged over the 22 winters of the simulation is shown in Fig. 12 (annex 3). As discussed by Cattiaux et al.  
574 (2010), winter 2010 is associated with a particularly cold season resulting from the persistence of the negative  
575 phase of the North Atlantic Oscillation (NAO). During the negative phase of the NAO, storms bring moist air  
576 from the Atlantic into the Mediterranean Sea and dry and cold air over northern Europe (Trigo et al. 2002) and  
577 cloud systems are frequent over the western part of the Mediterranean Basin (Chaboureau and Claud 2006): the  
578 signature of high clouds storm track over southern Europe and the Mediterranean Basin (CF anomaly around 5%)  
579 is noticeable in Fig. 12 (annex 3), as well as the advection of dry and cold air favoring clear sky (and hence  
580 negative anomaly of clouds) over Northern Europe (Trigo et al. 2002). However, winter 2010 shows an even  
581 stronger cloud fraction anomaly over the eastern part of the domain where it reaches 9%. Deeper analysis of this  
582 winter in terms of interactions between clouds and dynamics is then needed but beyond the scope of this study.

583 ERA-Interim reanalysis, which forces the simulation, contains the large-scale conditions which drive cloud  
584 anomalies: it is then rational that the simulation is able to reproduce these specific anomalies. It gives us  
585 confidence to use the complete time-period of this simulation to quantify the amplitude of cloud variability over  
586 two decades. It should allow determining how many years are needed to cover the effect of internal climate  
587 variability on cloud vertical profile, at present and without any external forcing.

588 In Fig. 9a, the mean value of CF at 9.5 km (altitude where CF is maximal in 2010) is extracted each year over  
589 the Mediterranean Sea. It gives 22 values of  $CF_{9.5}$ . The blue bars represent the standard deviation (STD) of these  
590  $CF_{9.5}$  (y-axis) estimated by varying the number of years (x-axis): the first blue bar is the  $CF_{9.5}$  STD over 5 years  
591 (from 2007 to 2011), the second over 6 years (2006 to 2011 for the simulation (blue axis)), and so on. Pink bars  
592 and pink axis are the same but for observations. Fig. 9b is the same as Fig. 9a but over the continent.

593 When considering the same years for observations and simulation, the simulated inter-annual variability at this  
594 altitude is greater than the observed one for both continent and sea. This is consistent with results in Sect. 5.1.  
595 (Fig. 7b and 7f for high clouds). The interannual variability of cloud fraction at 9.5km is greater over the  
596 continent ( $\approx 4.5\%$ ) than over the sea ( $\approx 3.8\%$ ). Also, the variability is more dependent on the years considered  
597 over the continent than over the sea: STD of  $CF_{9.5}$  ranges from 3.4% to 4% over the sea and from 4.1% to 4.9%  
598 over the continent. We notice that over the continent the STD computed over 13 and more years is around 4.1%:  
599 it varies of 0.1% between 13 and 22 years, while it varies of 0.7% between 5 and 22 years. The STD of  $CF_{9.5}$  is  
600 less dependent on the number of years when computed over 13 and more years. Over the sea this statement is not  
601 valid: the variability of STD of  $CF_{9.5}$  stabilizes between 13-20 years (0.2% compared to 0.6%) but decreases  
602 significantly between 21 and 22 years (making the STD variability between 13 and 22 years around 0.5%).

603 The blue dots are the winter 2010  $CF_{9.5}$  anomaly computed relative to the mean  $CF_{9.5}$  averaged over different  
604 time periods as indicated by the blue axis, for the simulation. The pink dots are the same as the blue dots for  
605 observations. This winter 2010 specific anomaly is confirmed in Fig. 9, where both observations and simulation  
606 agree that 2010 is a particular winter, identifying a particular event in terms of cloud occurrence since the  
607 anomaly is greater than the corresponding STD. For observations, over both sea and continent, the  $CF_{9.5}$  anomaly  
608 in winter 2010 is greater than 1.5 STD. It's up to twice STD for the simulation over the sea, in particular for time  
609 periods more than 10 years. The 2010 CF anomaly is less marked over the continent than over the sea with the  
610 simulation (Fig. 9b).

611 This takes us back to the question of spatial distribution that is not very discussed in this paper and that is  
612 necessary to understand some processes. Why the anomaly is more marked over the sea? Which feedbacks does  
613 this imply? Also, it appears that even if the simulation captures the anomaly, its intensity against the natural  
614 variability keeps uncertain despite we consider a winter nighttime situation where most of variability is driven by  
615 large scale. As discussed by Hawkins and Sutton (2009), internal variability and model uncertainties are strong.  
616 However, this study paves the way for a better assessment of clouds trends.

617

## 618 6. CONCLUSION

619

620 The goal of this study was to characterize the nighttime cloud vertical profile variability over Europe. Actually,  
621 analyzing the behavior of the seasonal and inter-annual vertical structure of clouds over the Euro-Mediterranean  
622 region is necessary to understand some climate anomalies that are not entirely explained by large scale  
623 dynamics.

624 We used CALIPSO-GOCCP observations and one WRF/MedCordex simulation that is nudged towards the  
625 reanalysis ERA-interim to reduce the biases due to dynamical effects. As the CALIPSO footprint has a small  
626 swath (the lidar essentially documents a curtain), it under-samples the area, which can impact the study. Hence,  
627 we first estimated the bias on the cloud fraction profile due to CALIPSO under-sampling in comparing simulated  
628 cloud profiles under the satellite flight track with simulated cloud profiles over the entire area. This comparison  
629 shows that biases on cloud fraction profiles due to CALIPSO under-sampling mainly depend on: i) clouds  
630 occurrence, ii) clouds spatial distribution and iii) clouds temporal variability. In particular, the bias can be non-  
631 negligible during fall and summer seasons, when convective clouds are more frequent than during the rest of the  
632 year. Nevertheless, the detailed lidar height-intensity histograms ( $SR(z)$ ) show that the effect of under-sampling  
633 mostly occurs in the non-cloudy bins. However, this assessment is based on simulation, which does not properly  
634 simulate cloud properties. As a consequence, in reality, the actual bias due to under-sampling might be different  
635 of the one estimated here. Combining the sampling errors estimated by the simulation and the model biases, an  
636 estimation of the actual sampling biases is presented in Table 3. The computation method is presented in Annex  
637 4. Actual errors of sampling averaged over the seasons are shown to be smaller than 1% in average in terms of

638 cloud fraction while the relative errors estimated are 12% for low clouds, 25% for mid clouds and 7% for high  
639 clouds. These values are comparable to the ones estimated by the model. These values can be larger for some  
640 levels (16% for high clouds, 19% for low clouds and up to 35% for mid clouds).

641 Then we addressed the three following questions in the current paper: (1) What is the seasonal variability of the  
642 cloud vertical distribution over Euro-Mediterranean area? (2) What is its inter-annual variability? (3) Are  
643 simulations able to reproduce the amplitude of these variabilities?

644 1) The seasonal variability of the cloud vertical distribution over Euro-Mediterranean area has been documented  
645 using 8 years of CALIPSO observations during nighttime. It results that clouds are more frequent (around 2%  
646 more clouds) over the continent than over the sea. This is probably linked to the geographical position of the  
647 Mediterranean as a transition region between Northern fluxes that are wet and cool and African fluxes that are  
648 dry and hot (Mariotti et al. 2015). Cheruy and Aires (2009) show that 70% of clouds population over the Euro-  
649 Mediterranean area is whether frontal (ascending large scale motions), stratocumulus or shallow cumulus  
650 (subsiding motions and near-surface instability) clouds. They also show that frontal and stratocumulus clouds are  
651 frequent over land while shallow cumulus clouds are frequent over both land and sea. Also, complex topography  
652 over land, particularly mountains, are favorable to the formation of cloud systems.

653 But the clouds are less variable from one season to another over the continent (less than 50% relative variability)  
654 than over the sea (relative variability reaches 100% in low levels). Besides, high clouds occurrence is shown  
655 (with observations) to be equivalent to low clouds occurrence in summer and winter while mid clouds are around  
656 5% smaller in terms of cloud fraction.

657 2) The cloud inter-annual variability is linked to the large scale circulation generating intra-seasonal variability  
658 and spatial distribution that are not discussed in this paper. Nevertheless, the eight years of CALIPSO  
659 observations show that winter inter-annual variability is around 4% for high clouds and 1% for low and mid  
660 clouds. Boé and Terray (2014) showed that cloud cover in 2031-2050 is expected to decrease by 3% over  
661 southern Europe and Mediterranean Sea when compared to 1961-1990. This decrease is within the range of the  
662 inter-annual variability of observed cloud fraction and weaker than the one simulated by this model and might  
663 not be totally associated to climate change.

664 3) Then we evaluated clouds vertical distribution in the WRF/MedCordex simulation. Comparisons between the  
665 observations and the “model+lidar simulator” outputs showed that the model overestimates the high clouds (20%  
666 more cloud fraction) and underestimates the mid and low clouds (5% less cloud fraction). This seems to be a  
667 persistent feature in models over Europe, which does not depend on spatial and temporal resolution of the model  
668 nor the model type (e.g Chiriaco et al. 2006 and Chaboureau et al. 2002 for mesoscale models; Cheruy et al.  
669 2013 for GCMs, Cesana and Chepfer 2012, Nam et al. , 2012; Tsushima et al. 2013). This default was pointed  
670 out using case studies observations (ground base sites and field campaigns) in the previous studies, and it is  
671 confirmed here using eight years of satellite observations by active sensors. As mentioned by Hawkins and  
672 Sutton (2009), there is room to reduce model uncertainty, in particular by improving physical parameterizations  
673 leading to this wrong vertical distribution of cloud layers. Despite these biases, the model is able to simulate  
674 realistic seasonal cycle.

675 Regarding the inter-annual variability in winter, the model simulates realistic inter-annual variability of spatially  
676 averaged CF in winter that is overestimated for high clouds (up to twice larger than observations for high clouds).  
677 The range of variability of CF in the simulation varies sometimes at different altitudes and with different  
678 intensity from the observations, particularly in summer. This is part of the model uncertainty that might be more  
679 complicated to reduce as it is also associated to clouds feedback.

## 680 REFERENCES

681 Appenzeller, C., J. R. Holton, and K. H. Rosenlof (1996): Seasonal variation of mass transport across  
682 the tropopause. *Journal of Geophysical Research, Atmospheres*, 101(D10), 15071–15078,  
683 doi:10.1029/96JD00821.

684 Bastin S., Chiriaco M., and Drobinski P. (2015): Control of radiation and evaporation on temperature  
685 variability in a WRF regional climate simulation: comparison with colocated long-term ground based  
686 observations near Paris. In revision to *Clim. Dyn.*

687 Boé J. and Terray L. (2014): Land–sea contrast, soil-atmosphere and cloud-temperature interactions:  
688 interplays and roles in future summer European climate change. *Climate Dynamics*. 42:683–699 DOI  
689 10.1007/s00382-013-1868-8

690 Cassou C., L. Terray, J. W. Hurrell, and C. Deser (2004): North Atlantic Winter Climate Regimes:  
691 Spatial Asymmetry, Stationarity with Time, and Oceanic Forcing. *J. Climate*, **17**, 1055–1068. doi:  
692 [http://dx.doi.org/10.1175/1520-0442\(2004\)017<1055:NAWCRS>2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2004)017<1055:NAWCRS>2.0.CO;2)

693 Cassou C., Laurent Terray, and Adam S. Phillips (2005): Tropical Atlantic Influence on European Heat  
694 Waves. *Journal of Climate*, 18, 2805–2811. doi: <http://dx.doi.org/10.1175/JCLI3506.1>

695 Cattiaux J., R. Vautard, C. Cassou, P. Yiou, V. Masson-Delmotte and F. Codron (2010): Winter 2010 in  
696 Europe: A cold extreme in a warming climate. *Geophysical Research Letters*, 37,20,  
697 DOI: 10.1029/2010GL044613

698 Cesana G. and H. Chepfer (2012): How well do climate models simulate cloud vertical structure? A  
699 comparison between CALIPSO-GOCCP satellite observations and CMIP5 models. *Geophysical Research*  
700 *Letters*, 39, L15704, doi: 10.1029/2012GL053153

701 Chaboureau J.P., J.P. Cammas, P. J. Mascart, J.P. Lafore, and J.P. Pinty (2002): Mesoscale model cloud  
702 scheme assessment using satellite observations. *Journal of Geophysical Research*, 107, no. D16, 4301,  
703 10.1029/2001jd000714

704 Chaboureau J. P. and C. Claud (2006): Satellite-based climatology of Mediterranean cloud systems and  
705 their association with large-scale circulation. *J. Of Geophys. Res.*, 111, D01102, doi:10.1029/2005JD006460.

706 Chaboureau J.-P., E. Richard, J.-P. Pinty, C. Flamant, P. Di Girolamo, C. Kiemle, A. Behrendt, H.  
707 Chepfer, M. Chiriaco, and V. Wulfmeyer (2012): Long-range transport of Saharan dust and its radiative impact  
708 on precipitation forecast: a case study during the Convective and Orographically induced Precipitation Study  
709 (COPS). *R. Met. S.*, 137, 236-251. DOI: 10.1002/qj.719

710 Chepfer H., M. Chiriaco, R. Vautard, and J. Spinhirne (2007): Evaluation of MM5 Optically Thin  
711 Clouds over Europe in Fall Using ICESat Lidar Spaceborne Observations. *Monthly Weather Review*, 135,  
712 2737–2753. DOI: 10.1175/MWR3413.1

713 Chepfer H., S. Bony, D. M. Winker, M. Chiriaco, J.-L. Dufresne, and G. Seze (2008): Use of CALIPSO  
714 lidar observations to evaluate the cloudiness simulated by a climate model. *Geophysical Research Letters*, 35,  
715 L20804, doi: 10.1029/2012GL053385.

716 Chepfer, H., S. Bony, D. Winker, G. Cesana, J. L. Dufresne, P. Minnis, C. J. Stubenrauch, and S. Zeng  
717 (2010): The GCM-Oriented CALIPSO Cloud Product (CALIPSO-GOCCP). *Journal of Geophysical Research*,  
718 *Atmospheres*, 115, D00H16, doi:10.1029/2009JD012251

719 Chepfer H., G. Cesana, D. Winker, B. Getzewich, M. Vaughan, And Z. Liu (2013): Comparison of Two  
720 Different Cloud Climatologies Derived from CALIOP-Attenuated Backscattered Measurements (Level 1): The  
721 CALIPSO-ST and the CALIPSO-GOCCP. *Journal of Atmospheric and Oceanic Technology*, **30**, 725-744. DOI:  
722 <http://dx.doi.org/10.1175/JTECH-D-12-00057.1>

723 Chepfer H., V. Noel, D. Winker, and M. Chiriaco (2014): Where and when will we observe cloud  
724 changes due to climate warming? *Geophysical Research Letters* 41, 8387–8395, doi: 10.1002/2014GL061792.

725 Cheruy F. and Filipe Aires (2009): Cluster Analysis of Cloud Properties over the Southern European  
726 Mediterranean Area in Observations and a Model. *Monthly Weather Review*, 137, 3161–3176. doi:  
727 <http://dx.doi.org/10.1175/2009MWR2882.1>

728 Cheruy F., A. Campoy, J.-C. Dupont, A. Ducharne, F. Hourdin, M. Haeffelin, M. Chiriaco, A. Idelkadi  
729 (2013): Combined influence of atmospheric physics and soil hydrology on the simulated meteorology at the  
730 SIRTA atmospheric observatory. *Climate Dynamics*. 40, 2251–2269, doi : 101007/s00382-012-1469-y

731 Chiriaco M., R. Vautard, H. Chepfer, M. Haeffelin, J. Dudhia, Y. Wanherdrick, Y. Morille and A.  
732 Protat (2006): The ability of MM5 to simulate ice clouds: systematic comparison between simulated and  
733 measured fluxes and lidar/radar profiles at the SIRTa atmospheric observatory. *Monthly Weather Review*, 134,  
734 897-918

735 Chiriaco M., Bastin S., Yiou P., Haeffelin M., Dupont Jean-Charles, Stefanon M. (2014): European  
736 heatwave in July 2006: Observations and modeling showing how local processes amplify conducive large-scale  
737 conditions, *Geophysical Research Letters*, 41, Issue 15, pages 5644–5652, DOI: 10.1002/2014GL060205

738 Dee D., Uppala S., Simmons A., Berrisford P., Poli P., Kobayashi S., Andrae U., Balmaseda M.,  
739 Balsamo G., Bauer P., Bechtold P., Beljaars A. C. M., Van de Berg L., Bidlot J., Bormann N., Delsol C.,  
740 Dragani R., Fuentes M., Geer A. J., Haimberger L., Healy S. B., Hersbach H., Hólm E. V., Isaksen L., Kållberg  
741 P., Köhler M., Matricardi M., McNally A. P., Monge-Sanz B. M., Morcrette J. J., Park B. K., Peubey C., Rosnay  
742 P., Tavolato C., Thépaut J. N. and Vitart F. (2011): The ERA-interim reanalysis: Configuration and performance  
743 of the data assimilation system. *Q. J. R. Meteorol. Soc.*, 137, 553–97.

744 Drobinski P., V. Ducrocq, P. Alpert, E. Anagnostou, K. Béranger, M. Borga, I. Braud, A. Chanzy, S.  
745 Davolio, G. Delrieu, C. Estournel, N. Filali Boubrahmi, J. Font, V. Grubišić, S. Gualdi, V. Homar, B. Ivančan-  
746 Picek, C. Kottmeier, V. Kotroni, K. Lagouvardos, P. Lionello, M. C. Llasat, W. Ludwig, C. Lutoff, A. Mariotti,  
747 E. Richard, R. Romero, R. Rotunno, O. Roussot, I. Ruin, S. Somot, I. Taupier-Letage, J. Tintore, R. Uijlenhoet,  
748 and H. Wernli (2014): HyMeX: A 10-Year Multidisciplinary Program on the Mediterranean Water Cycle.  
749 *Bulletin of the American Meteorological Society*, 95, 1063–1082. doi: [http://dx.doi.org/10.1175/BAMS-D-12-](http://dx.doi.org/10.1175/BAMS-D-12-00242.1)  
750 00242.1

751 Dudhia J., (1989): Numerical Study of Convection Observed during the Winter Monsoon Experiment  
752 Using a Mesoscale Two-Dimensional Model. *Journal of Atmospheric Sciences*, 46, 3077–3107. doi:  
753 [http://dx.doi.org/10.1175/1520-0469\(1989\)046<3077:NSOCOD>2.0.CO;2](http://dx.doi.org/10.1175/1520-0469(1989)046<3077:NSOCOD>2.0.CO;2)

754 Flaounas E., Drobinski P., Vrac M, Bastin S., Lebeaupin-Brossier C, Stéfanon M., Borga M. and Calvet  
755 J. C. (2013): Precipitation and temperature space-time variability and extremes in the Mediterranean region:  
756 Evaluation of dynamical and statistical downscaling methods. *Climate Dynamics*. 40, 2687–2705. DOI:  
757 10.1007/s00382-012-1558-y

758 Flaounas E., S.a Raveh-Rubin, H. Wernli, P. Drobinski, and S. Bastin (2014): The dynamical structure  
759 of intense Mediterranean cyclones. *Climate Dynamics*, 44, 2411-2427. DOI: 10.1007/s00382-014-2330-2  
29

760 Funatsu BM, Claud C, and Chaboureau JP. (2009): Comparison between the large-scale environment of  
761 moderate and intense precipitating systems in the Mediterranean region. *Monthly Weather Review*, 137: 3933–  
762 3959.

763 Gettelman A. And P.M. De F. Forster (2002): A Climatology of the Tropical Tropopause Layer. *Journal*  
764 *of the Meteorological Society of Japan*, Vol. 80, No. 4B, pp. 911-924. Doi: <http://doi.org/10.2151/jmsj.80.911>

765 Giorgi F., C. Jones, and G. Asrar (2009): Addressing climate information needs at the regional level: the  
766 CORDEX framework, *WMO Bull.*, 58 (3) (2009), pp. 175–183

767 Hawkins E. and R. Stutton (2009): The potential to narrow uncertainty in regional climate predictions.  
768 *Bulletin of American Meteorological Society*, 90, 1095–1107. doi: <http://dx.doi.org/10.1175/2009BAMS2607.1>

769 Hong, S.Y., J. Dudhia, and S.H. Chen (2004): A revised approach to ice microphysical processes for the  
770 bulk parameterization of clouds and precipitation, *Monthly Weather Review*, 132, 103–120.

771 Kain J.S. (2004): The Kain–Fritsch Convective Parameterization: An Update. *J. Appl. Meteor.*, 43,  
772 170–181. doi: [http://dx.doi.org/10.1175/1520-0450\(2004\)043<0170:TKCPAU>2.0.CO;2](http://dx.doi.org/10.1175/1520-0450(2004)043<0170:TKCPAU>2.0.CO;2)

773 Khvorostyanov V. I., (1995): Mesoscale processes of cloud formation, cloud-radiation interaction, and  
774 their modelling with explicit cloud microphysics. *Atmospheric Research*, 39, issue 1-3, pp1-67.  
775 doi:10.1016/0169-8095(95)00012-G

776 Kjellström, E., G. Nikulin, U. Hansson, G. Strandberg, and A. Ullerstig (2011): 21<sup>st</sup> century changes in  
777 the European climate: uncertainties derived from an ensemble of regional climate model simulations. *Tellus A*,  
778 *Series A*, 63A(1), 24-40, doi: [10.1111/j.1600-0870.2010.00475.x](https://doi.org/10.1111/j.1600-0870.2010.00475.x).

779 Lebeaupin-Brossier C., S. Bastin, K. Béranger, and P. Drobinski (2015): Regional mesoscale air–sea  
780 coupling impacts and extreme meteorological events role on the Mediterranean Sea water budget, *Climate*  
781 *Dynamics*, 44, 1029–1051, DOI [10.1007/s00382-014-2252-z](https://doi.org/10.1007/s00382-014-2252-z)

782 Lenderink G., A. van Ulden, B. van den Hurk and E. van Meijgaard (2007): Summertime inter-annual  
783 temperature variability in an ensemble of regional model simulations: analysis of the surface energy budget.  
784 *Clim. Change*. 81: 1,233-274. DOI [10.1007/s10584-006-9229-9](https://doi.org/10.1007/s10584-006-9229-9)

785 Mariotti A., Yutong Pan, Ning Zeng, and Andrea Alessandri (2015) Long-term climate change in the  
786 Mediterranean region in the midst of decadal variability. *Climate Dynamics*, 44, Issue 5-6, pp 1437-1456. DOI  
787 10.1007/s00382-015-2487-3

788 Mlawer J. E., Taubma J. S., Brown D. P., Iancono M. J., Clough A. S. (1997): Radiative transfer for  
789 inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave, *Journal of Geophysical*  
790 *Research*, 102, Issue D14, 16663–16682, DOI: 10.1029/97JD00237

791 Monin, A. S., and Obukhov, A. (1954): Basic laws of turbulent mixing in the surface layer of the  
792 atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, 151, 163-187.

793 Nam, C., S. Bony, J.-L. Dufresne, and H. Chepfer (2012): The ‘too few, too bright’ tropical low-cloud  
794 problem in CMIP5 models, *Geophysical Research Letters.*, 39, L21801, doi:10.1029/2012GL053421.

795 Noh Y., W. G. Cheon and S. Y. Hong (2003): Improvement of the k-profile model for the planetary  
796 boundary layer based on large eddy simulation data. *Boundary-layer meteorology*, 107, 401-427

797 Omrani H, Drobinski P, and Dubos T. (2013): Optimal nudging strategies in regional climate modelling:  
798 investigation in a big-brother experiment over the European and Mediterranean regions. *Climate Dynamics*, 41:  
799 2451-2470, doi:10.1007/s00382-012-1615-6.

800 Omrani, H., Drobinski, P., Jourdier, B., and Brossier, C.L. (2014): Investigation on the offshore wind  
801 energy potential over the north western Mediterranean Sea in a regional climate system model. *IEEE, Renewable*  
802 *Energy Congress (IREC), 2014 5th International*, DOI: 10.1109/IREC.2014.6826956

803 Reiter E. R. , (1975): *Handbook for Forecasters in the Mediterranean: Weather Phenomena of the*  
804 *Mediterranean Basin, Part 1. General Description of the Meteorological Processes*, Environmental Prediction  
805 Research Facility Naval Postgraduate School, Monterey, CA (1975), p. 344 Tech. Pap. 5-75

806 Rojas M., L. Z. Li, M. Kanakidou, N. Hatzianastassiou, G. Seze, and H. Le Treut (2013): Winter  
807 weather regimes over the Mediterranean region: their role for the regional climate and projected changes in the  
808 twenty-first century, *Climate Dynamics*. DOI 10.1007/s00382-013-1823-8

809 Ruti P, Somot S, Dubois C, Calmanti S, Ahrens B, Alias A, Aznar R, Bartholy J, Bastin S, Branger K,  
810 Brauch J, Calvet J-C, Carillo A, Decharme B, Dell’Aquila A, Djurdjevic V, Drobinski P, Elizalde-Arellano A,

811 Gaertner M, Galn P, Gallardo C, Giorgi F, Gualdi S, Harzallah A, Herrmann M, Jacob D, Khodayar S, Krichak S,  
812 Lebeaupin C, L'Heveder B, Li L, Liguro G, Lionello P, Onol B, Rajkovic B, Sannino G, and Sevault F (2015) :  
813 MED-CORDEX initiative for Mediterranean climate studies, *Bulletin of the American Meteorological Society*.  
814 doi: <http://dx.doi.org/10.1175/BAMS-D-14-00176.1>

815 Rysman J.-F., S. Verrier, Y. Lemaître, and E. Moreau, (2013): Space-time variability of the rainfall over  
816 the western Mediterranean region: A statistical analysis, *Journal of Geophysical Research Atmospheres*, 118,  
817 8448–8459, doi:10.1002/jgrd.50656

818 Salameh T, Drobinski P, and T. Dubos (2010): The effect of indiscriminate nudging time on large and  
819 small scales in regional climate modelling: Application to the Mediterranean basin. *Q. J. R. Meteorol. Soc.*  
820 136(646): 170-182, doi:10.1002/qj.518.

821 Skamarock W. C. and J. B. Klemp (2008): A time-split nonhydrostatic atmospheric model for weather  
822 research and forecasting applications. *Journal of Computational Physics*, 227:3465-3485, DOI:  
823 10.1016/j.jcp.2007.01.037.

824 Smirnova T. G., Brown John M., and B. Stanley G. (1997): Performance of Different Soil Model  
825 Configurations in Simulating Ground Surface Temperature and Surface Fluxes. *Monthly Weather Review*,  
826 125:1870–1884. doi: [http://dx.doi.org/10.1175/1520-0493\(1997\)125<1870:PODSMC>2.0.CO;2](http://dx.doi.org/10.1175/1520-0493(1997)125<1870:PODSMC>2.0.CO;2)

827 Smirnova, T. G., J. M. Brown, S. G. Benjamin, and D. Kim (2000), Parameterization of cold-season  
828 processes in the MAPS land-surface scheme, *Journal of Geophysical Research*, 105(D3):4077–4086,  
829 doi:10.1029/1999JD901047.

830 Soden B. J. and I. M. Held (2006): An Assessment of Climate Feedbacks in Coupled Ocean–  
831 Atmosphere Models. *Journal of Climate*, 19:3354–3360. doi: <http://dx.doi.org/10.1175/JCLI3799.1>

832 Stauffer DR and Seaman NL. (1990): Use of four-dimensional data assimilation in a limited-area  
833 mesoscale model. part i: Experiments with synoptic-scale data. *Mon. Weather Rev.* 118(6): 1250-1277,  
834 doi:10.1175/1520-0493(1990)118.

835 Stefanon M., P. Drobinski, F. D'Andrea, C. Lebeaupin-Brossier, and S. Bastin (2014): Soil moisture-  
836 temperature feedbacks at meso-scale during summer heat waves over Western Europe, *Clim Dyn*, 42:1309–1324,  
837 DOI 10.1007/s00382-013-1794-9

838 Stephens G. L. (2005): Cloud Feedbacks in the Climate System: A Critical Review. *Journal of Climate*,  
839 18:237–273. doi: <http://dx.doi.org/10.1175/JCLI-3243.1>

840 Tang Q., Guoyong Leng, and Pavel Ya. Groisman (2012): European Hot Summers Associated with a  
841 Reduction of Cloudiness. *Journal of Climate*, 25:3637–3644. doi: <http://dx.doi.org/10.1175/JCLI-D-12-00040.1>

842 Trigo R. M., T. J. Osborn, and J. M. Corte-Real (2002): The North Atlantic Oscillation influence on  
843 Europe: climate impacts and associated physical mechanisms. *Climate Research*, 20: 9–  
844 17. doi:10.3354/cr020009

845 Tsushima Y., Mark A. Ringer, Mark J. Webb, and Keith D. Williams (2013): Quantitative evaluation of  
846 the seasonal variations in climate model cloud regimes. *Climate Dynamics*, V 41, Issue 9, pp 2679-2696. DOI:  
847 10.1007/s00382-012-1609-4

848 Vaittinada Ayar P., Vrac M., Bastin S., Carreau J., and Gallardo C. (2015): Intercomparison of  
849 statistical and dynamical downscaling models under the EURO- and MED-CORDEX initiative framework:  
850 Present climate evaluations. *Climate dynamics*, doi:10.1007/s00382-015-2647-5

851 Vautard R. (1990): Multiple weather regimes over the North Atlantic: Analysis of precursors and  
852 successors. *Monthly Weather Review*, 118, 2056–2081. doi: [http://dx.doi.org/10.1175/1520-0493\(1990\)118<2056:MWROTN>2.0.CO;2](http://dx.doi.org/10.1175/1520-0493(1990)118<2056:MWROTN>2.0.CO;2)

854 Winker D. M., Jacques R. Pelon, and M. Patrick McCormick (2003): The CALIPSO mission:  
855 spaceborne lidar for observation of aerosols and clouds, *Proc. SPIE 4893, Lidar Remote Sensing for Industry*  
856 *and Environment Monitoring III*, 1 (March 24, 2003). doi: <http://dx.doi.org/10.1117/12.466539>

857 Winker D. M., M. A. Vaughan, A. Omar, Y. Hu, and K. A. Powell (2009): Overview of the CALIPSO  
858 Mission and CALIOP Data Processing Algorithms. *Journal of Atmospheric and Oceanic Technology*, 26:2310-  
859 2323. doi: <http://dx.doi.org/10.1175/2009JTECHA1281.1>

860 Xoplaki E., J. F. Gonzalez-Rouco, J. Luterbacher, and H. Wanner (2004): Wet season Mediterranean  
861 precipitation variability: influence of large-scale dynamics and trends. *Climate Dynamics*. 23: 63–78, DOI  
862 10.1007/s00382-004-0422-0

863 Yiou, P., R. Vautard, P. Naveau, and C. Cassou (2007): Inconsistency between atmospheric dynamics  
864 and temperatures during the exceptional 2006/2007 fall/winter and recent warming in Europe, *Geophysical*  
865 *Research Letters*, 34, L21808, doi:10.1029/2007GL031981.

866 **TABLE CAPTION**

867 Table 1: Definition of the datasets used for the study. Columns define the sampling method. Lines stand for the  
868 product used. SR is the scattering ratio measured by the lidar (annex 1) and CF is the percentage of clouds per  
869 gridbox.

870 Table 2: Maximal differences (i.e. absolute – “abs.” – errors) between simulated cloud fractions extracted with  
871 full resolution and simulated cloud fraction extracted with CALIPSO sampling [ $CF_{WRF+sim}(z) - CF_{WRF+sim}^T(z)$ ]  
872 for high, mid and low clouds. Relative (“rel.”) errors are computed as follow:  $\left| \frac{CF_{WRF+sim}(z) - CF_{WRF+sim}^T(z)}{CF_{WRF+sim}^T(z)} \right| \times 100$   
873 and represent the average value of the layer (low, mid, or high).  $CF_{WRF+sim}(z)$  and  $CF_{WRF+sim}^T(z)$  are averaged  
874 over 2006-2011 by seasons (columns).

875 Table 3: Estimation of the sampling errors in average (over the years and spatially) considering the sampling  
876 estimation evaluated by the simulation and the average model biases. “Avg” stands for errors estimated in  
877 average over the layer (low, mid or high) while “max” stands for the maximal error detected over one level.  $\alpha$   
878 values are the results of section 4.1.1. while  $\beta$  values are the results of section 3.

879 Table 4: Computing model biases over continent ( $\alpha = \frac{CF_{WRF+sim}(z)}{CF_{GOCCP}(z)}$ ) for low clouds (1st row), mid clouds (2<sup>nd</sup>  
880 row) and high clouds (3<sup>rd</sup> row) by testing different undersampling (test 1 means we extract 1 profile over 20 and  
881 test 8 means we extract all the profiles).

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883

884 **FIGURE CAPTION**

885 **Fig.1** Total area of study with an illustration of the number of CALIPSO overpass in each  $20 \times 20$  km<sup>2</sup> grid-box  
886 during one season (here JJA 2008)

887 **Fig.2** Example of a CALIPSO track in 2009/01/19 during night, around 5°E longitude. (a) SR (z) observed by  
888 CALIOP lidar; (b) the total condensed water Q(z) simulated by WRF; (c) SR(z) simulated by WRF using COSP  
889 lidar simulator

890 **Fig.3** Simulated cloud vertical profiles cumulated over the entire area of study and averaged over 2006-2011. (a)  
891 Summer (JJA), (b) fall (SON), (c) winter (DJF), (d) spring (MAM). Red profiles correspond to simulation with  
892 the sampling that follows CALIPSO tracks ( $CF_{WRF+sim}(z)$ ) and black profiles are for the entire simulation  
893 sampling ( $CF_{WRF+sim}^T(z)$ ); For each of the subplots, around  $4 \cdot 10^5$  profiles have been used to construct the red  
894 profiles while  $2 \cdot 10^6$  profiles have been used to construct the black ones.

895 **Fig.4** SR(z) histograms cumulated from 2006/06 to 2011/12 for simulation: (a) SR(z) simulation extracted  
896 following CALIPSO sampling  $SR_{WRF+sim}(z)$  in logarithmic scale; (b) same as (a) following WRF sampling  
897  $SR_{WRF+sim}(z)$ ; (c) is (a)-(b); (d) same as (c) only for summers; (e) fall; (f) winter; (g) spring. Colorbars are  
898 percentage of the normalized occurrence at each level (the sum of one line is 100%). The vertical black lines  
899 represent the SR = 5 threshold for cloud detection

900 **Fig.5** Seasonal variability of observed and simulated cloud fraction profiles. (a) Mean cloud fraction profile on  
901 2006-2011 for the simulation (blue) and observations (pink) horizontally averaged over the continent; the shades  
902 represent the envelope of the four seasons averaged profiles; (b) same as (a) over the sea; (c) Envelope of the  
903 seasonal anomaly computed relative to the mean cloud fraction profile over the continent in the simulation (blue  
904 shade) in the observations (pink shade); (d) same as (c) over the sea; (e) Mean cloud fraction profile on 2006-  
905 2011 for each season for observations horizontally averaged over the continent; (f) same as (e) over the sea; (g)  
906 same as (e) for simulated profiles; (h) same as (g) over the sea. In (e) to (h), blue is for winter, green is for spring,  
907 red is for summer, and light blue is for fall

908 **Fig.6** Same as Fig. 4a for observations and simulation in CALIPSO sampling. First column is for observations  
909  $SR_{GOCCP}(z)$ , and second column is for the simulation  $SR_{WRF+sim}(z)$ . (a) and (b) are for winter, (c) and (d) are for

910 spring, (e) and (f) are for summer, (g) and (h) are for fall. Black bars are for the threshold of cloud detection (SR  
911 =5). Colorbar is in logarithmic scale

912 **Fig.7** Inter-annual standard deviation of averaged cloud fraction profiles from 2006 to 2011. (a) For winter in the  
913 continent; (b) winter in the sea; (c) summer in continent; (d) summer in sea. Pink shade is for observations, blue  
914 are for simulation under CALIPSO sampling, and the black line for the simulation in the full sampling. (e) The  
915 shade represents +/- the envelope of the standard deviation shown in (a). The blue line is the maximal anomaly  
916 values for the simulation. Pink lines are the same as blue lines for observations. (f), (g) and (h) are the same as (e)  
917 for winter in the sea, summer in the continent and (d) summer in the sea

918 **Fig.8** CF winter anomaly compared to mean 2007-2011 for different years and altitudes, spatially averaged (a)  
919 over the Mediterranean Sea for the simulation from 1989 to 2011; (b) over the Mediterranean Sea for  
920 observations from 2007 to 2013; (c) and (d) same as (a) and (b) over the continent

921 **Fig.9** (a) Standard deviation of the simulated CF value at  $z = 9.5$  km ( $CF_{9.5}$ ) computed over the Mediterranean  
922 Sea for different number of winters (blue bars). The blue x-axis explains the time period associated to the  
923 number of years on which these standard deviations were computed (e.g. the first blue bar is the standard  
924 deviation computed over 5 years of simulation and corresponds to the winters 2007 to 2011). The pink bars and  
925 pink x-axis are the same as blue bars and blue x-axis but for observations. The blue dots are the  $CF_{9.5}$  winter  
926 2010 anomalies relative to the average  $CF_{9.5}$  computed over the different time periods. Pink dots are the same as  
927 blue dots for observations. (b) same as (a) over the continent

928 **Fig.10** Two instantaneous observed SR vertical profiles (blue around [5°E; 47°N] and red around [5°E; 43°N])  
929 in 2009/01/19 at night, and vertical black line represents the SR = 5 threshold for cloud detection. Red box in (a)  
930 represents the Mediterranean Sea area while the blue box is for Europe area

931 **Fig.11** Winter cloud fraction maps ( $CF_{WRF}^T$ : cloud fraction computed from the model without lidar simulator)  
932 averaged from 2006 to 2011 for simulation low clouds (a), mid-clouds (b) and high clouds (c). d, e and f are the  
933 same but for summer

934 **Fig.12** Winter 2010 high clouds anomaly computed with  $CF_{WRF+sim}^T$  relative to the average high cloud map of  
935 winters from 1990 to 2011

936

	<i>CALIPSO sampling: using only grid-boxes along the satellite track</i>	<i>WRF sampling: using all the domain grid-boxes</i>
<i>GOCCP observations</i> <i>06/2006 - 12/2011</i>	$SR_{GOCCP}(z)$ $CF_{GOCCP}(z)$	-
<i>WRF simulation + COSP simulator</i> <i>06/2006 - 12/2011</i>	$SR_{WRF+sim}(z)$ $CF_{WRF+sim}(z)$	$SR_{WRF+sim}^T(z)$ $CF_{WRF+sim}^T(z)$

938 *Table 1: Definition of the datasets used for the study. Columns define the sampling method. Lines stand for the*  
939 *product used. SR is the scattering ratio measured by the lidar (annex 1) and CF is the percentage of clouds per*  
940 *gridbox.*

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		<i>Summer</i>	<i>Fall</i>	<i>Winter</i>	<i>Spring</i>
<i>Low clouds</i> <i>(7 levels from the ground to 3.2 km)</i>	Maximal abs. errors	+ 0.18 %	+ 0.25 %	+ 0.75 %	+ 0.14 %
	rel. errors	11%	6%	8 %	8 %
<i>Mid clouds</i> <i>(7 levels from 3.2 to 6.5 km)</i>	Maximal abs. errors	- 0.05 %	- 0.28 %	- 1.5 %	- 0.57 %
	rel. errors	11 %	7 %	13 %	9 %
<i>High clouds</i> <i>(13 levels from 6.5 km to 13 km)</i>	Maximal abs. errors	- 3.2 %	- 2.6 %	+ 2.0 %	+ 1.5 %
	rel. errors	21 %	9 %	5 %	7 %

954 *Table 2: Maximal differences (i.e. absolute – “abs.” – errors) between simulated cloud fractions extracted with*  
955 *full resolution and simulated cloud fraction extracted with CALIPSO sampling [ $CF_{WRF+sim}(z) - CF_{WRF+sim}^T(z)$ ]*  
956 *for high, mid and low clouds. Relative (“rel.”) errors are computed as follow:  $\left| \frac{CF_{WRF+sim}(z) - CF_{WRF+sim}^T(z)}{CF_{WRF+sim}^T(z)} \right| \times$*   
957 *100 and represent the average value of the layer (low, mid, or high).  $CF_{WRF+sim}(z)$  and  $CF_{WRF+sim}^T(z)$  are*  
958 *averaged over 2006-2011 by seasons (columns).*

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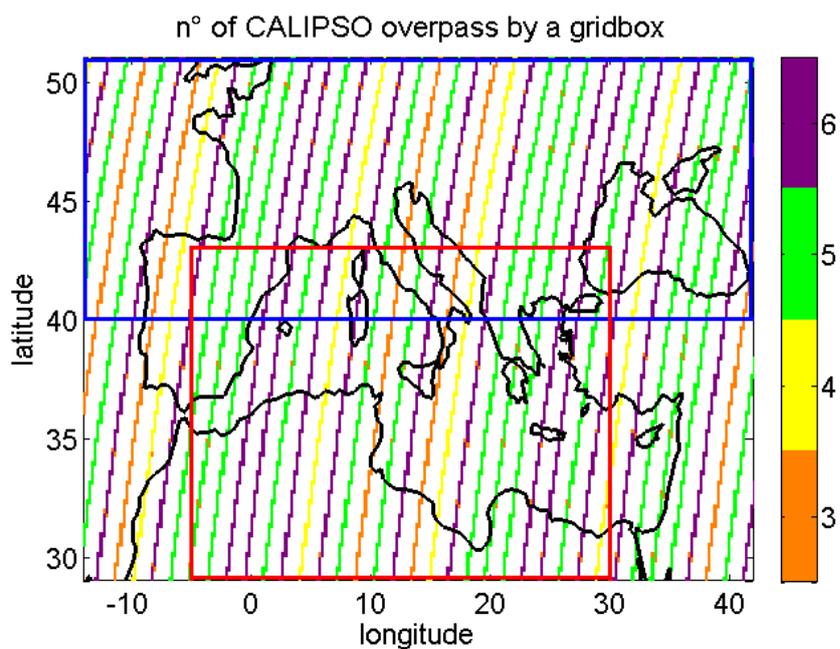
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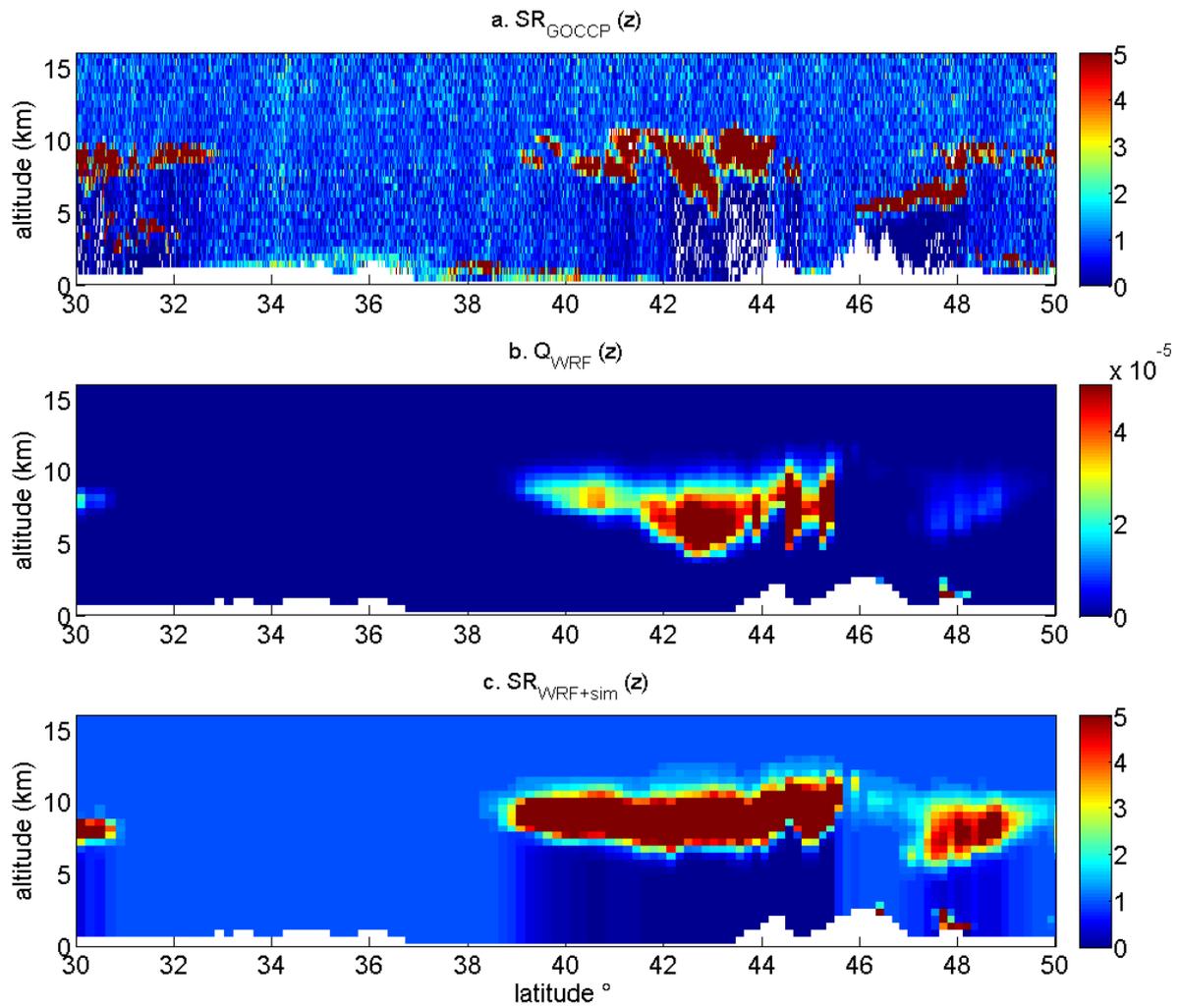
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	<i>CF Relative model bias</i> $\alpha = \frac{CF_{WRF+sim}}{CF_{GOCCP}}$	<i>CALIPSO undersampling error estimated by simulated cloud fraction</i> $\beta =  CF_{WRF+sim} - CF_{WRF+sim}^T $	$\varepsilon = \frac{\beta}{\alpha}$
<i>Low clouds</i>	0.36	Avg=0.15% ; max=0.33%	Avg=0.4% ; max=0.9%
<i>Mid clouds</i>	0.43	Avg=0.4% ; max=0.6%	Avg=0.9% ; max=1.4%
<i>High clouds</i>	2.32	Avg=1.4% ; max=2.35%	Avg=0.6% ; max=1%

965 *Table 3: Estimation of the sampling errors in average (over the years and spatially) considering the sampling*  
 966 *estimation evaluated by the simulation and the average model biases. “Avg” stands for errors estimated in*  
 967 *average over the layer (low, mid or high) while “max” stands for the maximal error detected over one level.  $\alpha$*   
 968 *values are the results of section 4.1.1. while  $\beta$  values are the results of section 3.*



969 **Fig.1** Total area of study with an illustration of the number of CALIPSO overpass in each 20×20 km<sup>2</sup> grid-box  
 970 during one season (here JJA 2008)



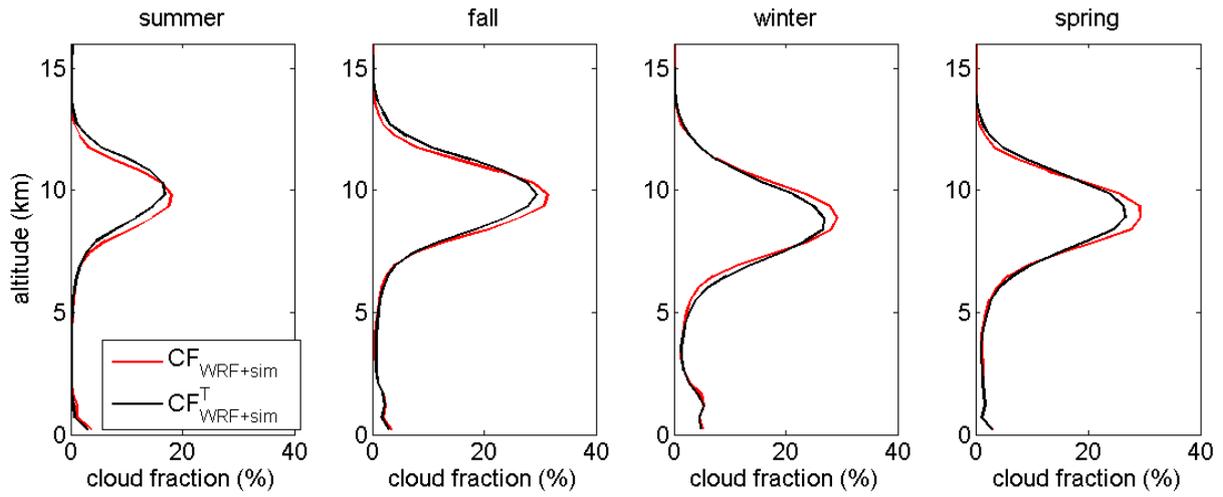
971 **Fig.2** Example of a CALIPSO track in 2009/01/19 during night, around 5°E longitude. (a) SR (z) observed by  
 972 CALIOP lidar; (b) the total condensed water Q(z) simulated by WRF; (c) SR(z) simulated by WRF using COSP  
 973 lidar simulator  
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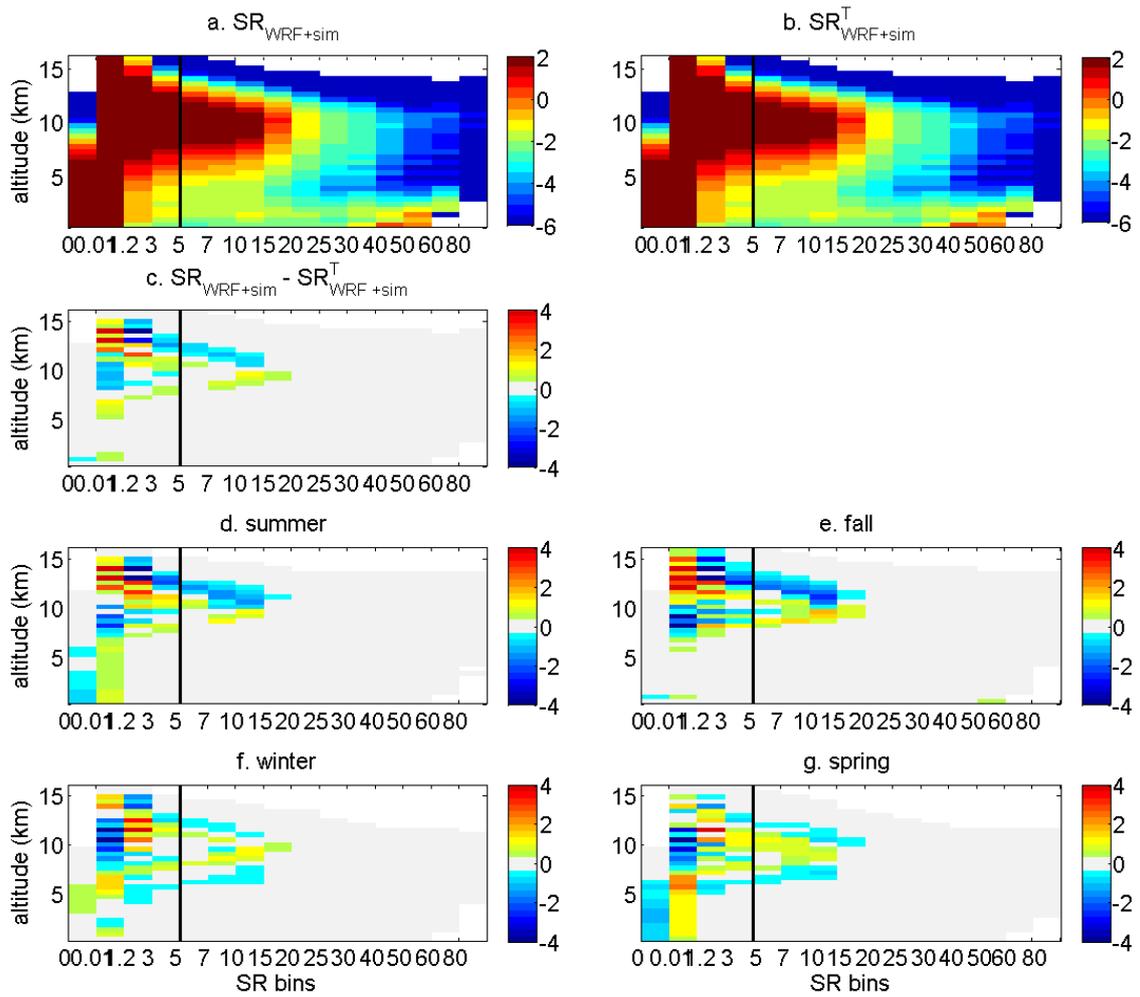
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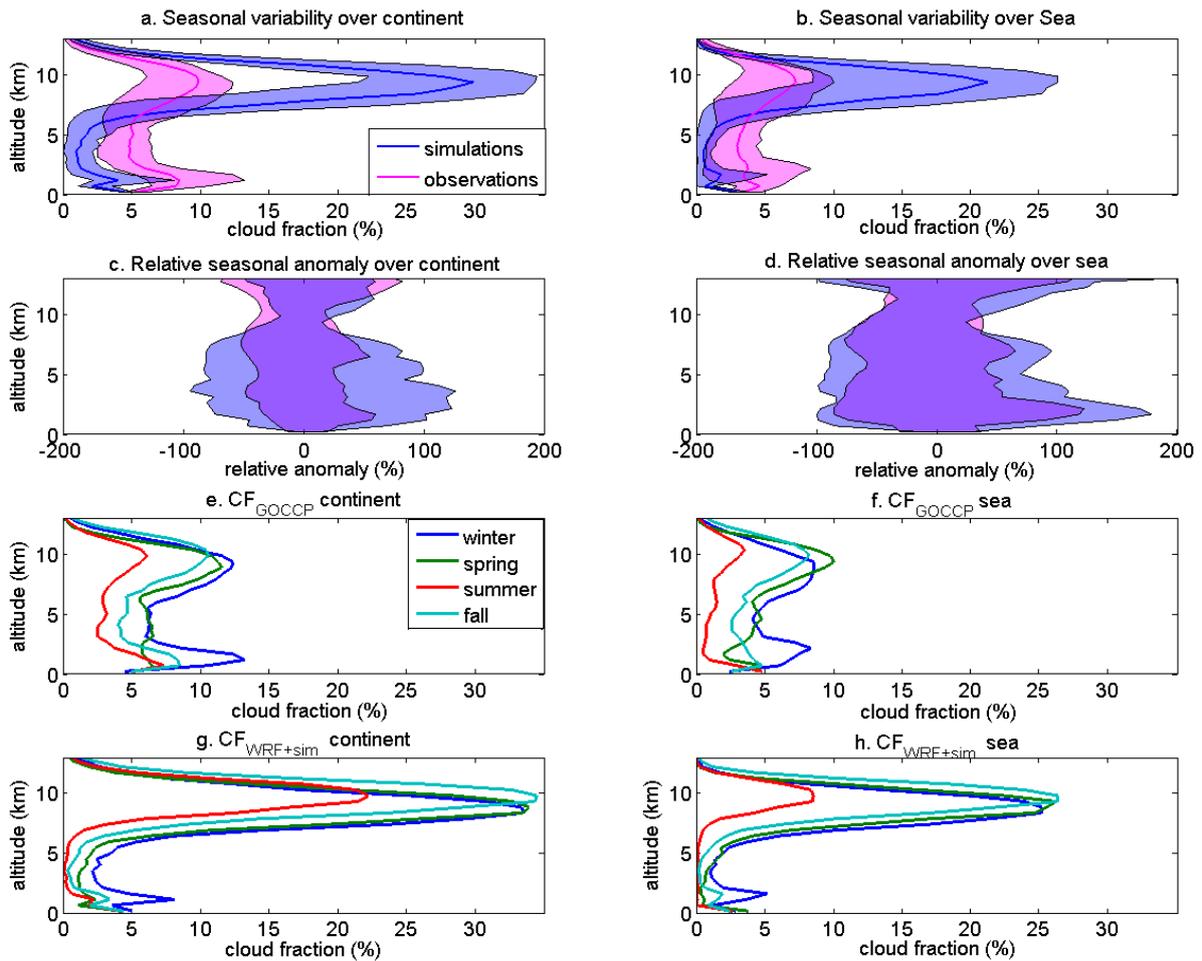
980 **Fig.3** Simulated cloud vertical profiles cumulated over the entire area of study and averaged over 2006-2011. (a)  
 981 Summer (JJA), (b) fall (SON), (c) winter (DJF), (d) spring (MAM). Red profiles correspond to simulation with  
 982 the sampling that follows CALIPSO tracks ( $CF_{WRF+sim}^T(z)$ ) and black profiles are for the entire simulation  
 983 sampling ( $CF_{WRF+sim}(z)$ ) ; For each of the subplots, around  $5 \cdot 10^4$  profiles have been used to construct the red  
 984 profiles while  $2 \cdot 10^6$  profiles have been used to construct the black ones.



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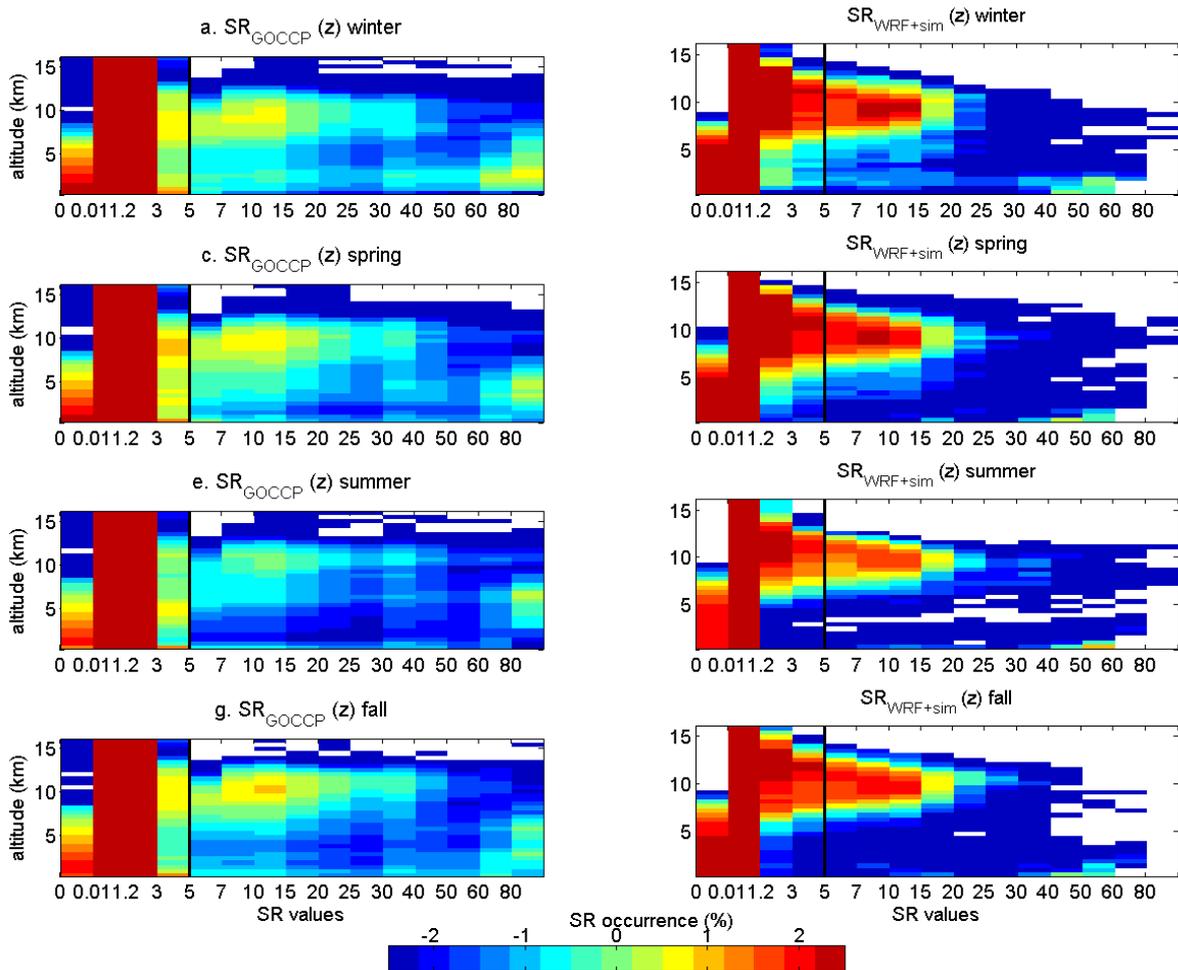
986 **Fig.4** SR(z) histograms cumulated from 2006/06 to 2011/12 for simulation: (a) SR(z) simulation extracted  
 987 following CALIPSO sampling  $SR_{WRF+sim}(z)$  in logarithmic scale; (b) same as (a) following WRF sampling  
 988  $SR_{WRF+sim}(z)$ ; (c) is (a)-(b); (d) same as (c) only for summers; (e) fall; (f) winter; (g) spring. Colorbars are  
 989 percentage of the normalized occurrence at each level (the sum of one line is 100%). The vertical black lines  
 990 represent the  $SR = 5$  threshold for cloud detection

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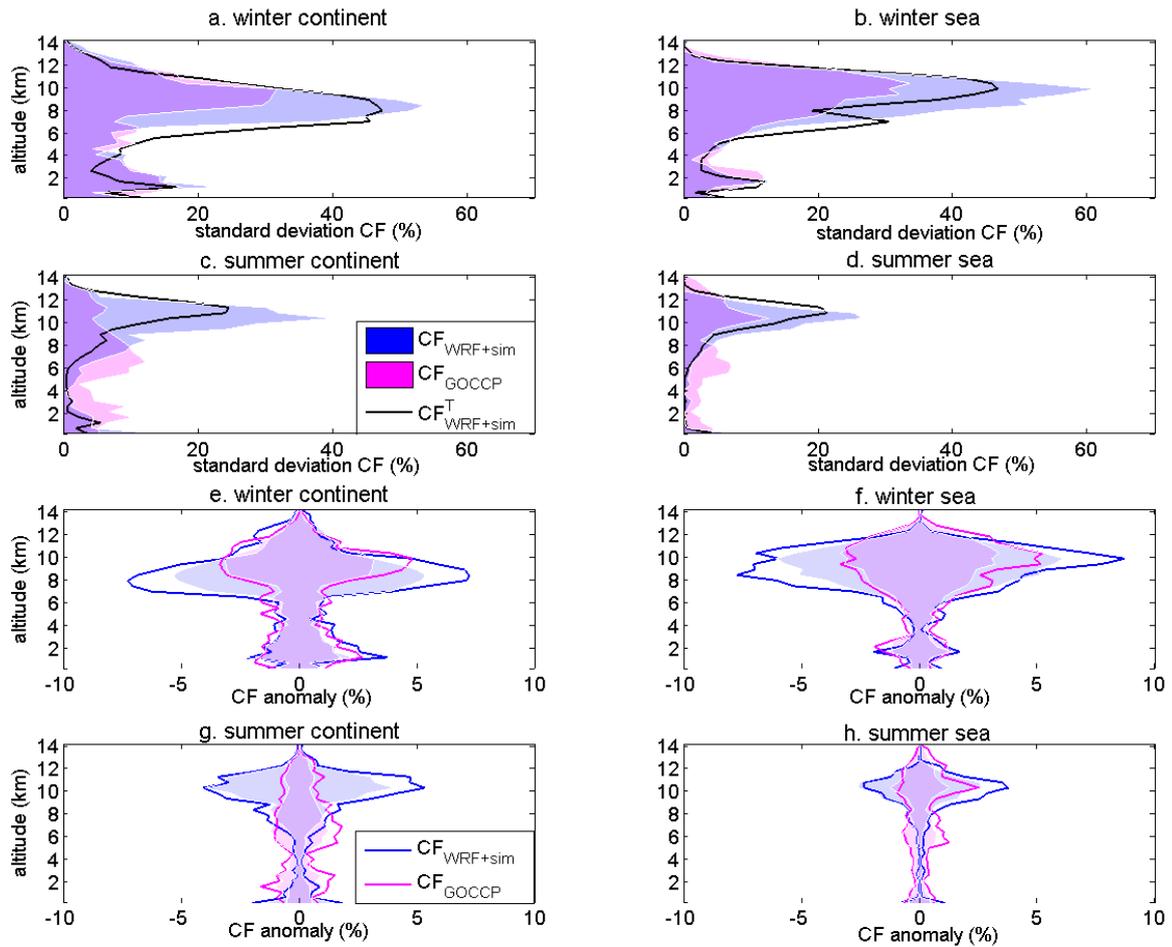
993 **Fig.5** Seasonal variability of observed and simulated cloud fraction profiles. (a) Mean cloud fraction profile on  
 994 2006-2011 for the simulation (blue) and observations (pink) horizontally averaged over the continent; the shades  
 995 represent the envelope of the four seasons averaged profiles; (b) same as (a) over the sea; (c) Envelope of the  
 996 seasonal anomaly computed relative to the mean cloud fraction profile over the continent in the simulation (blue  
 997 shade) in the observations (pink shade); (d) same as (c) over the sea; (e) Mean cloud fraction profile on 2006-  
 998 2011 for each season for observations horizontally averaged over the continent; (f) same as (e) over the sea; (g)  
 999 same as (e) for simulated profiles; (h) same as (g) over the sea. In (e) to (h), blue is for winter, green is for spring,  
 1000 red is for summer, and light blue is for fall



1001

1002 **Fig.6** Same as Fig. 4a for observations and simulation in CALIPSO sampling. First column is for observations  
 1003  $SR_{GOCCP}(z)$ , and second column is for the simulation  $SR_{WRF+sim}(z)$ . (a) and (b) are for winter, (c) and (d) are for  
 1004 spring, (e) and (f) are for summer, (g) and (h) are for fall. Black bars are for the threshold of cloud detection ( $SR$   
 1005  $=5$ ). Colorbar is in logarithmic scale

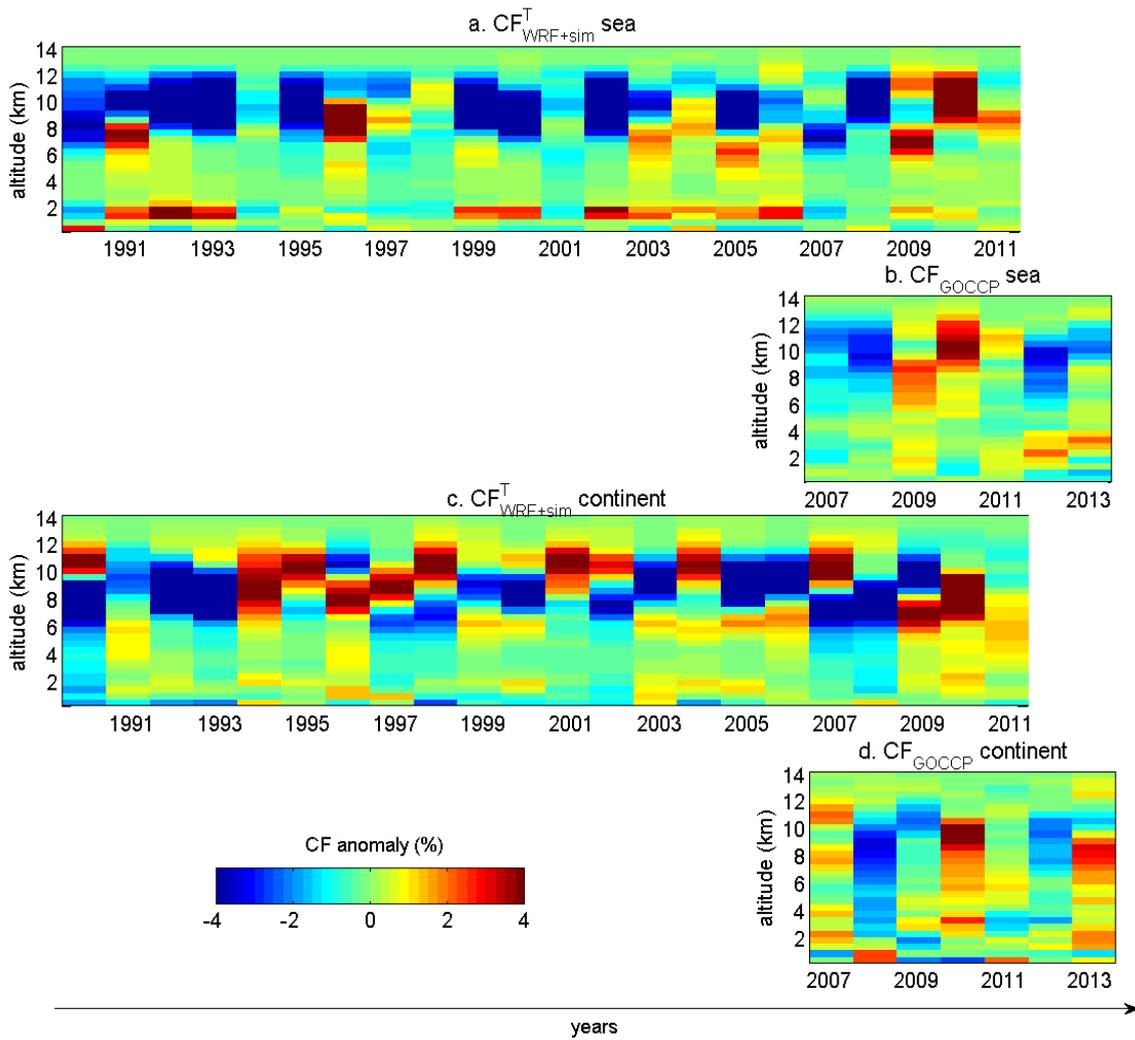
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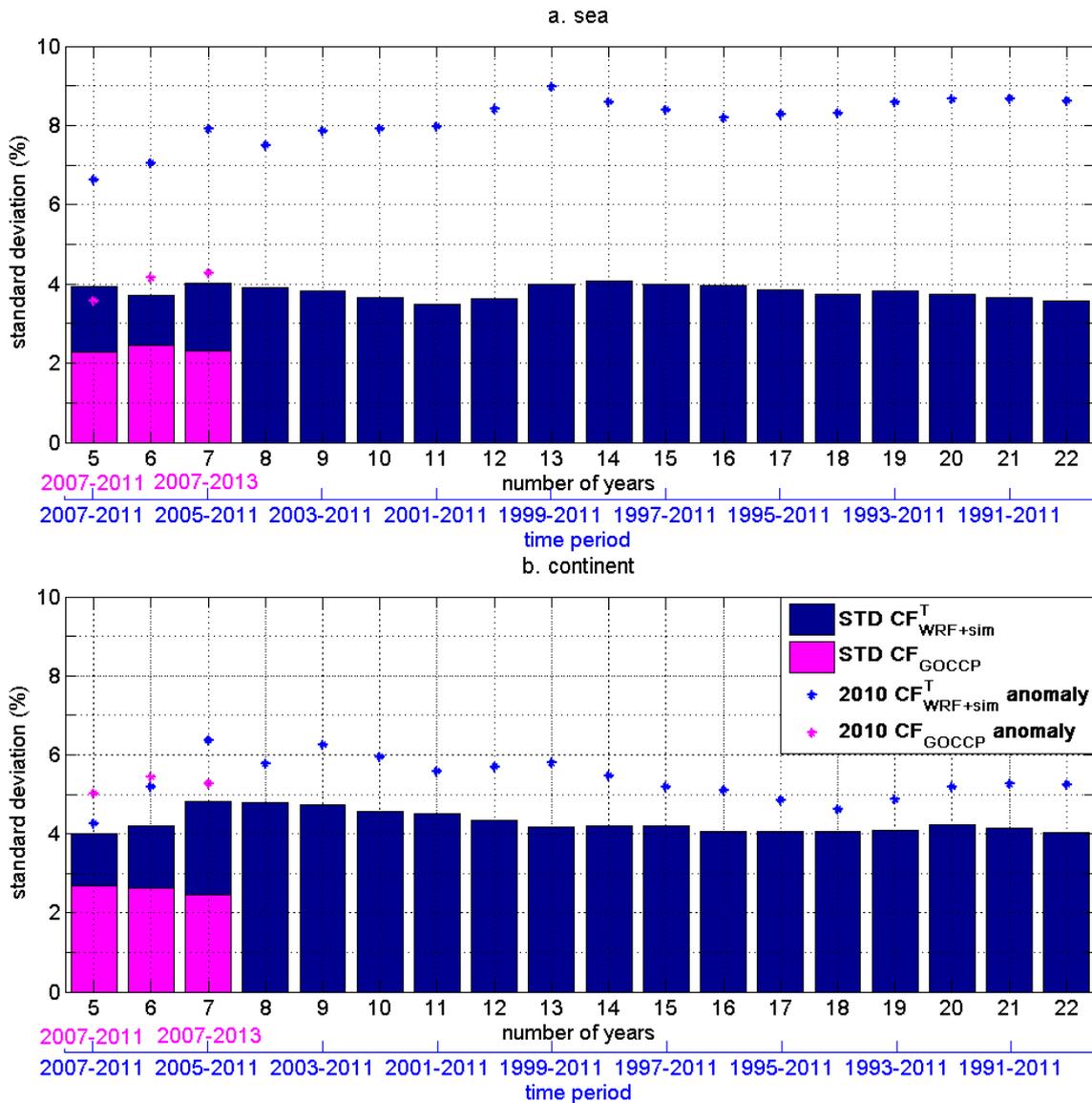
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**Fig.7** Inter-annual standard deviation of averaged cloud fraction profiles from 2006 to 2011. (a) For winter in the  
 1009 continent; (b) winter in the sea; (c) summer in continent; (d) summer in sea. Pink shade is for observations, blue  
 1010 are for simulation under CALIPSO sampling, and the black line for the simulation in the full sampling. (e) The  
 1011 shade represents  $\pm$  the envelope of the standard deviation shown in (a). The blue line is the maximal anomaly  
 1012 values for the simulation. Pink lines are the same as blue lines for observations. (f), (g) and (h) are the same as (e)  
 1013 for winter in the sea, summer in the continent and (d) summer in the sea



1014

1015 **Fig.8** CF winter anomaly compared to mean 2007-2011 for different years and altitudes, spatially averaged (a)  
 1016 over the Mediterranean Sea for the simulation from 1989 to 2011; (b) over the Mediterranean Sea for  
 1017 observations from 2007 to 2013; (c) and (d) same as (a) and (b) over the continent



1018

1019 **Fig.9** (a) Standard deviation of the simulated CF value at  $z = 9.5$  km ( $CF_{9.5}$ ) computed over the Mediterranean  
 1020 Sea for different number of winters (blue bars). The blue x-axis explains the time period associated to the  
 1021 number of years on which these standard deviations were computed (e.g. the first blue bar is the standard  
 1022 deviation computed over 5 years of simulation and corresponds to the winters 2007 to 2011). The pink bars and  
 1023 pink x-axis are the same as blue bars and blue x-axis but for observations. The blue dots are the  $CF_{9.5}$  winter  
 1024 2010 anomalies relative to the average  $CF_{9.5}$  computed over the different time periods. Pink dots are the same as  
 1025 blue dots for observations. (b) same as (a) over the continent

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1028 **Annex 1: Lidar equation**

1029 The scattering ratio SR is given by (1):

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$$SR(z) = \frac{ATB_{tot}(z)}{ATB_{mol}(z)} \quad (1)$$

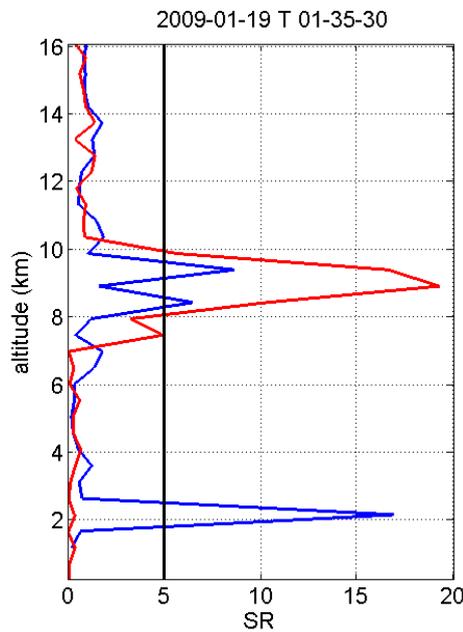
1031 Where  $ATB_{tot}$  and  $ATB_{mol}$  are respectively the attenuated backscattered signals for particles and molecules and  
 1032 for molecules only and are given by (2) and (3):

1033 
$$ATB_{tot}(z) = (\beta_{sca,part}(z) + \beta_{sca,mol}(z)) \cdot e^{-2\eta \int_{z_{TOA}}^z (\alpha_{sca,part}(z) + \alpha_{sca,mol}(z)) \cdot dz} \quad (2)$$

1034 
$$ATB_{mol}(z) = \beta_{sca,mol}(z) \cdot e^{-2\eta \int_{z_{TOA}}^z \alpha_{sca,mol}(z) \cdot dz} \quad (3)$$

1035  $ATB_{mol}$  and  $ATB_{tot}$  products are averaged vertically to obtain SR over 40 layers (Chepfer et al. 2008 and 2010).

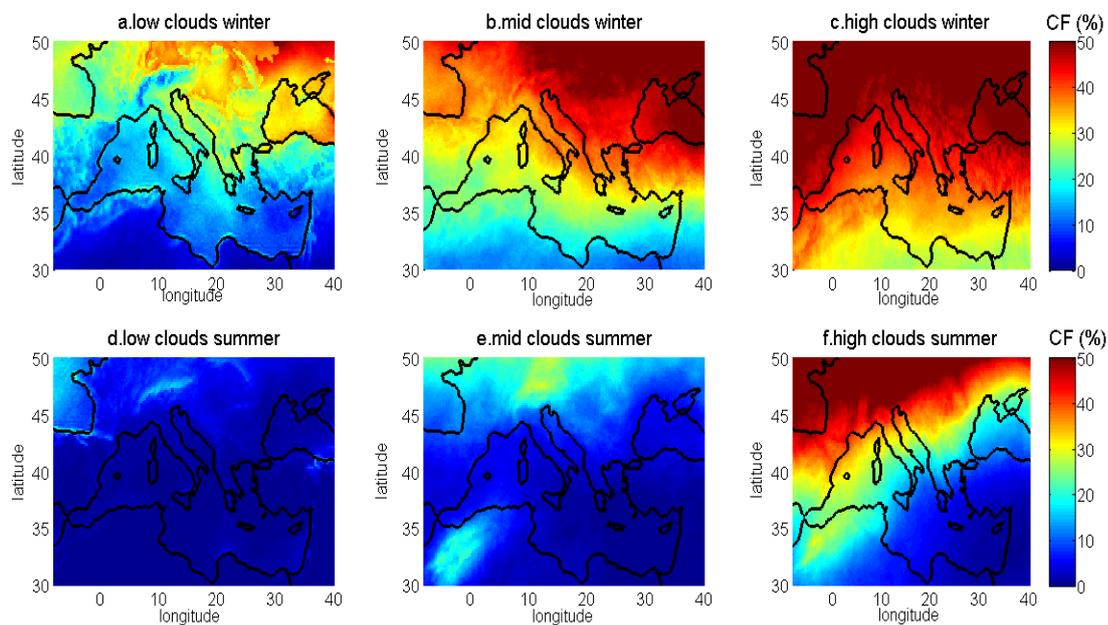
1036  $\beta_{sca,part}$ ,  $\beta_{sca,mol}$  are lidar backscatter coefficients ( $m^{-1} sr^{-1}$ ) and  $\alpha_{sca,part}$  and  $\alpha_{sca,mol}$  attenuation coefficients ( $m^{-1}$ ) for  
 1037 particles (clouds, aerosols) and molecules.  $\eta$  is a multiple scattering coefficient that depends both on lidar  
 1038 characteristics and size, shape and density of particles. It is about 0.7 for CALIPSO (Winker et al., 2003;  
 1039 Chepfer et al., 2008).



1040  
 1041 **Fig.10:** Two instantaneous observed SR vertical profiles (blue around [5°E; 47°N] and red around [5°E; 43°N])  
 1042 in 2009/01/19 at night. The vertical black line represents the SR = 5 threshold for cloud detection.

1043 Figure 10 illustrates two instantaneous SR profiles to help understand what a lidar signal looks like and how  
 1044 cloud detection is computed in this study. Above 10 km,  $SR(z)$  is around 1, indicating clear sky for both profiles.  
 1045 High clouds are detected in both profiles between 8 km and 10 km:  $SR(z)$  of the blue profile reaches the value of  
 1046 8 and while  $SR(z)$  of the red one goes up to 22. The magnitude of  $SR(z)$  depends on the cloud optical thickness  
 1047 from the Top Of Atmosphere (TOA) to the level  $z$  and the cloud microphysical properties such as the size of the  
 1048 particle or its shape. While the signal is fully attenuated for the red profile below 8 km ( $SR(z)$  is almost zero),  
 1049 the blue profile still detects low clouds around 2 km.

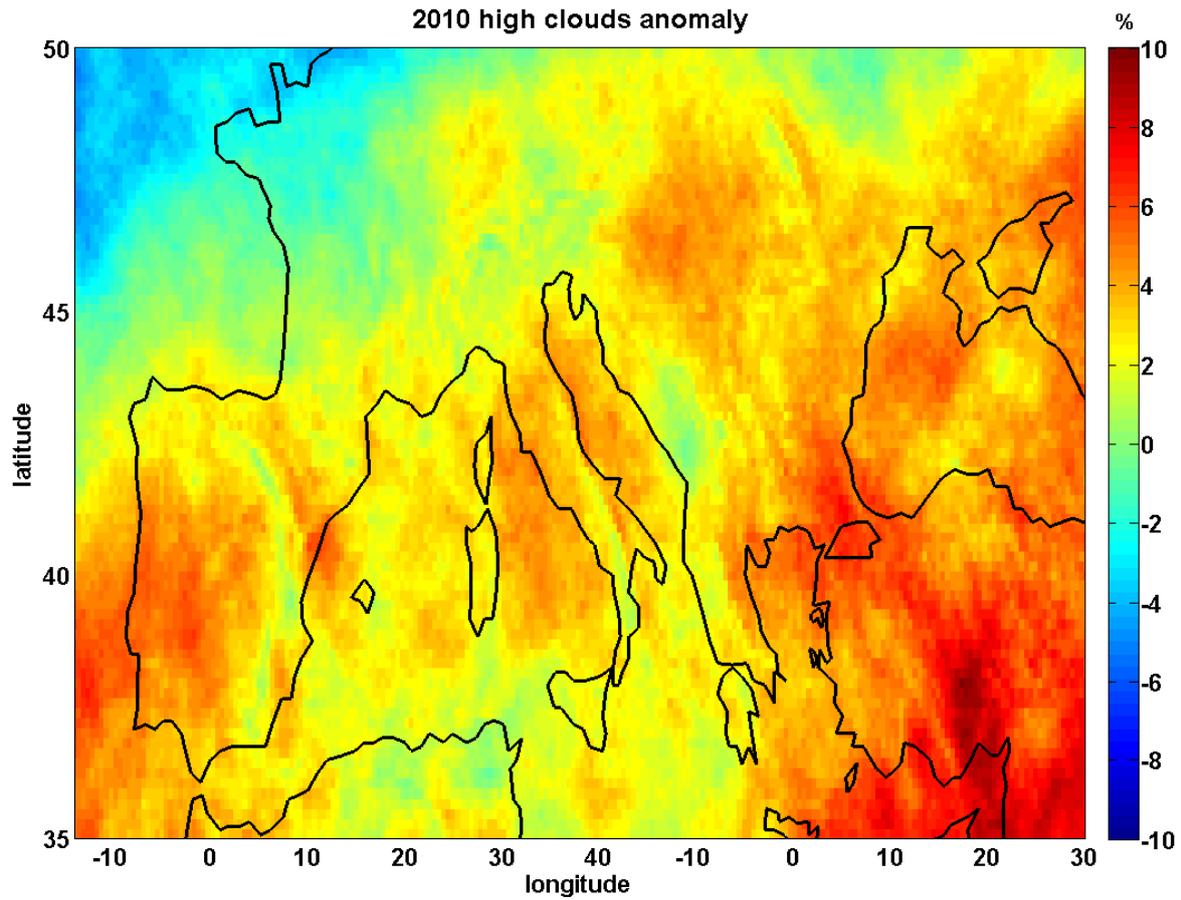
1050 **Annex 2: Simulated cloud fraction maps**



1051  
 1052 **Fig.11** Winter cloud fraction maps (cloud fraction computed from the model without lidar simulator) averaged  
 1053 from 2006 to 2011 for simulation low clouds (a), mid-clouds (b) and high clouds (c). d, e and f are the same but  
 1054 for summer

1055 Fig.11 shows that: for high clouds, a north-south gradient exists in winter with about 10% of clouds over North  
 1056 Africa and more than 50% above continental Europe, while in summer, this gradient is north-west/south-east,  
 1057 with almost no high clouds over Turkish and eastern part of Mediterranean basin. In winter, most mid and low  
 1058 clouds occur above the north-eastern part of Europe. In summer, very few mid and low clouds are simulated and  
 1059 they are mostly induced by orography.

1060 **Annex 3: Simulated winter 2010 high clouds anomaly**



1061

1062 **Fig12** Winter 2010 high clouds anomaly computed with  $CF_{WRF+sim}^T$  relative to the average high cloud map of  
 1063 winters from 1990 to 2011

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1069 Annex 4:

1070 *The CALIPSO undersampling error estimation from observed cloud fraction profiles is defined as:*

1071  $\varepsilon(z) = |CF_{GOCCP}(z) - CF_{GOCCP}^T(z)|$  with  $CF_{GOCCP}^T(z)$  a theoretical cloud fraction that we would have with a  
 1072 complete sampling (observations over all the grid-boxes every 00UTC).

1073 We define  $\alpha(z)$  as the relative model bias, so  $\alpha(z) = \frac{CF_{WRF+sim}(z)}{CF_{GOCCP}(z)}$

1074 We used a set of different samplings to test if  $\alpha(z)$  can be considered as constant, i.e independent of the number  
 1075 of profiles in the sampling. To do that, since we need both observations and simulation to test this hypothesis, we  
 1076 reduced the CALIPSO sampling using only 1 profile over 2 (test 7), 1 over 3 (test 6), and so on down to one  
 1077 profile over 20. Table 4 presents the results of these tests and indicates the  $\alpha$  values for low, mid and high clouds.  
 1078 This shows that if the number of profiles become greater than 1/15 of the CALIPSO sampling,  $\alpha(z)$  can be  
 1079 considered as nearly constant.

1080 We deduce that:  $\alpha(z) = \frac{CF_{WRF+sim}^T(z)}{CF_{GOCCP}^T(z)}$

1081 and  $\varepsilon(z)$  can be written as  $\varepsilon = \frac{|CF_{WRF+sim}(z) - CF_{WRF+sim}^T(z)|}{\alpha(z)} = \frac{\beta(z)}{\alpha(z)}$  with  $\beta(z)$  defined as the error of  
 1082 undersampling estimated by the simulation.

Sampling	Test 1 :	Test 2 :	Test 3 :	Test 4 :	Test 5 :	Test 6 :	Test 7 :	Test 8 :
	1 /20	1/15	1/10	1 /5	1/4	1/3	1/2	1/1
low	0.29	0.43	0.38	0.38	0.4	0.36	0.37	0.36
mid	0.5100	0.4700	0.4000	0.4000	0.4200	0.4200	0.4300	0.4300
high	7.9400	2.3300	2.4000	2.1900	2.3300	2.3500	2.3300	2.3200

1083 Table 4: computing model biases over continent ( $\alpha = \frac{CF_{WRF+sim}(z)}{CF_{GOCCP}(z)}$ ) for low clouds (1st row), mid clouds (2<sup>nd</sup> row)  
 1084 and high clouds (3<sup>rd</sup> row) by testing different samplings (test 1 means we extract 1 profile over 20 and test 8  
 1085 means we extract all the profiles).