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Comparison of rainfall profiles in the West African monsoon as depicted by the TRMM-PR and LMDZ climate model

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Abstract

Vertical rainfall profiles obtained with TRMM-PR 2A25 standard products are compared with rain profiles deduced from LMDZ model with two parametrization schemes: Emanuel (1991) and Tiedkte (1989). We focus on the low layers of the atmosphere over West Africa during the monsoon season (JJAS). The precipitation decrease above 4km is systematically not represented in rainfall profiles generated by Emanuel's parametrization scheme. However, Emanuel's shows a decrease similar to the observation from 4km down to the surface, especially in the Sahel (proper depth of the layer dominated by re-evaporation). As to Tiedkte's scheme, it describes best the downward increase in the upper levels of the atmosphere, whereas the downward decrease in the lower levels begins too low when compared to the observation. Tiedkte's parametrization shows an overestimation of liquid water production over Ocean and over the Guinean region and a slightly too strong re-evaporation in Sahara and Sahel. The zonal distribution of vertical rain profiles is then biased with this model scheme compared to the 2A25-PR product. On the other hand, although Emanuel's scheme detects too much re-evaporation over Sahara and underestimates liquid water production over Ocean compared to PR observation. It shows a good meridional distribution of these parameters. This is especially true in Sahel where Emanuel's scheme gives the best representation of re-evaporation.

1. Introduction

Understanding precipitation systems and evaluate their contribution to the water cycle in West Africa is an important research issue. Rainfall over this region is controlled by the low levels advection of moisture from the Gulf of Guinea. This advected moisture flow is the signature of the African Monsoon that drives the North-South rain gradient between the coast and the Sahara. This moisture gradient also leads to specific water vapor profiles and surface-atmosphere exchanges as we move from the coast to the North. This induces a regional modulation of the rain re-evaporation. Wherever this re-evaporation occurs, it induces a mesoscale downdraught driven by the adiabatic cooling from thermal exchanges (Zipser, 1969; Leary and Houze, 1979). This adiabatic cooling cancels at least partially the deep convective heating in regions of large-scale ascent (Bister and Emanuel, 1997). Additionally, the re-evaporation of rainfall profiles is also responsible for the generation of cold pools of air or wakes, involved in the self organization and propagation of the squall lines in West Africa.

Many efforts have been made to improve representation of convective rainfall in climate models since it remains a critical issue for the simulation of the current climate (IPCC 2001). In particular, precipitating downdraught and re-evaporation processes mentioned above were shown to be a key issue for the representation of the mean tropical climate (Hourdin et al 2006, Braconnot et al. 2007) and its variability (Lin et al. 2006) in coupled ocean-atmosphere models. As storms propagate from initiation (genesis) to decay, they leave behind them footprints of condensation rate and evaporation rate at various levels of the troposphere. These features have an important impact on the circulation in tropics (e.g. Mapes and Houze 1993).

One way to understand dynamics and thermodynamics of these processes and their associated microphysics is to identify local variations of the vertical structure of precipitation systems (e.g., Bart and Dejene 2005, Schumacher and Houze 2006, Houze 1981, Zipser and Lutz 1994) as measured since January 1998 by the Tropical Rainfall Measuring Mission Precipitation Radar (TRMM-PR). Appropriate computation of statistics for vertical profiles of precipitation will help to infer some of the regional and seasonal characteristics of the re-evaporation processes in the lower layers of the atmosphere.

The present study aims to characterize and compare the vertical profiles of precipitations obtained respectively with the TRMM PR radar and two different parameterizations schemes used in the LMDZ climate model. This study focuses mostly in the assessment of differences that might be found between the PR and the two sets of parametrisation in terms of vertical rain structure. This limited ambition is due to the difference in the definition of rain in the observation and the model and in the possible quantitative bias that exists in the PR data due to sampling.

A short presentation of the TRMM-PR specifications and data characteristics will be given to help understand the specificities of satellite rain observations. The sampling method is an important issue for this work. To characterize the main climatological structures, long time series are necessary. Three datasets are built to characterise the most relevant one for our study. Description of LMDZ and main characteristics of the two parametrizations schemes of the convection are given. These two descriptions are critical to analyse the results presented afterwards and frame the comparison. Although the representation of rain and evaporation in the lower atmospheric layers in the model are

quite different from the observation which in turns is not unbiased, this comparison gives a good insight on the possible strength and weaknesses of each parameterization.

2. Data and Methods

a. TRMM data and Analysis methods

In this study we use the Tropical Rainfall Measuring Mission-Precipitation Radar (TRMM-PR) data (June to September) from 1998 to 2006. A detailed description of TRMM-PR characteristics can be found in Kummerow et al. (1998). Nevertheless an important aspect of TRMM-PR that must be emphasized here is its sampling. With a swath of about 230km, the PR visits the same location only once or twice a day and at approximately the same local time every 47.5 days (Negri et al. 2002), except at the equator where the sampling is about once every other day. Hence, one full day of PR orbit is not enough to cover the tropical belt without gaps and also a single orbit cannot completely sample large systems. Since regional distribution of rainfall in West Africa is characterised by an important variability, it is necessary to build long range statistics of precipitation systems in West Africa. In the present study 9 years (1998 to 2006) of PR-2A25 data are used to retrieve average rain rate profiles from version 6 products. These rain rates are obtained after correction of attenuation on the measured reflectivity profiles (Meneghini et al. 2000, Iguchi et al. 2000) and using a dynamically adjusted Z-R relationship (Iguchi et al. 2000). Because our purpose is to compare observations and models and their respective definition of convective and stratiform rainfall differ, we consider only the total rainfall. All orbits overpassing West Africa during the considered

period are kept. In order to match the model spatial resolution, the statistics are built by averaging the PR data in a 1°x1° grid.

Although many attempts were made to assess the quality of the PR rain profiles it is difficult to get a definite answer since these statistics are the only one of their kind. One can nevertheless state that two main sources of uncertainties should be kept in mind when performing the comparison with the model-based profiles. First, the attenuation correction although assumed to be quite robust (Seto and Iguchi 2007) can always be questioned locally especially over land due to soil moisture effects. This should nevertheless not affect the surface rain estimates by more than 1 to 2% on average. Second, the minimum sensitivity of the PR (about 15 to 17 dBZ) does not necessarily give a good sensitivity to the ice phase aloft specially the content of the smaller ice particles.

Due to the specific downward looking geometry of PR, ground clutter echoes are affecting the edges of the swath. This shadow zone affects the radar bins up to almost 2km at the very edge of the swath, leading to a strong dependence of the number of averaged point as a function of altitude between the surface and 2km altitude. Additionally, topography can increase this effect and some of the lower bins might then be affected by ground clutter leading to biased averages. To minimize this effect, three datasets based on different sampling strategies were tested: the first one is the full dataset i.e. it is made of all the pixels for each PR-scan (49 pixels) in order to sample the largest possible number of systems; the second is the “10-pixel sampling” dataset, made with the 10 central pixels of each scan. The third dataset is the “central pixel sampling” which is elaborated with the unique central pixel of each scan. The interest of this third dataset is

to be able to go down as low as possible before being contaminated by ground clutters and with minimal mainlobe contamination. Fig. 1 shows a meridional cross- section of the averaged percentage of discarded pixels with respect to the number of pixels at 4 km, on land (Fig. 1.a) and on ocean (Fig.1.b) for the 3 datasets. These transects (Fig.1.a, Fig.1.b) are used by many authors for studying the West African monsoon features and also by models intercomparison projects like AMIP and ALMIP.

The so-called "lost pixels" are the pixels eliminated after quality control i.e. eliminated pixels by 2A25 algorithm filtering. As expected, the dataset made only of the central pixel goes closer to the ground than the other datasets. Over land the 10% contour of lost pixel is just below 1.5 km altitude between latitude 0 and 22°N, and just below 2 km between latitude 22°N and 24°N. Around 1.5 km, less than 4% of the pixels are lost in the Equatorial region while less than 1% are lost in Sahel and between 1% and 4% are lost in Sahara. The percentages are quite similar with the 10-pixel sampling dataset (small differences are found in Sahara and Sahel) but a substantial difference can be found when compared to the full dataset. With the latter, the 10% contour of lost pixels is located around 2km altitude and 30% at 1.5 km highlighting the ground surface echo contamination expected on the edges of the swath. Over ocean, most profiles remain valid down to 1 km altitude. The 10-pixel sampling exhibits a similar pattern but all contours are slightly higher in altitude. When keeping all the pixels, the 10% contour of lost pixel is at around 1.75 km altitude. Since we are interested by re-evaporation in the lowest atmospheric layers, we will keep only the 10-pixel sampling hereafter for the comparison, because it offers a good trade-off between the sampling and the contamination by ground clutter.

Fig. 2.a and 2.b illustrate the zonal average of the vertical precipitation profiles. The average rain intensities are generally weaker over land than over ocean but the rain signature goes higher in altitude (up to 7.5 km for land, only 5km for ocean) and are more widely spread in latitude (from 0°N to 18°N for land, only from 2°N to 14°N for ocean). Although, this unexpected heavier rain over ocean could be a problem in 2A25 v6, it is also possibly due to the comparatively small oceanic domain that was selected and that encompasses one of the region with the heaviest rain observed in Western Africa: west of the Fouta-Djallon. Fig. 3 shows a zonal average of the gradient of vertical retrieved rainfall rate from 4km to ground for each of the three datasets and for both oceanic and continental regions. Once again, only the levels with less than 10% of lost pixels are kept. One can notice areas with either downward decreasing (negative values) and/or downward increasing (positive values) rainfall. The Guinean gulf (Eq. to 5°N) and the Guinean regions (5°N to 8°N) are generally characterized by a downward increasing rainfall while Sahel (8°N to 18°N) and Sahara (18°N to 24°N) present a downward decreasing rainfall. This decrease in rain amount as one goes closer to the ground is likely to be the combination of both re-evaporation in the stratiform regions and water accumulation in convective cells. The oceanic regions are characterized in every case by a downward increasing rain probably due to a very moist air ascent coupled with shallower precipitation.

b. Model and Experiments design

LMDZ model used here is the Laboratoire de Météorologie Dynamique second generation global climate model; the Z of LMDZ is standing for Zoom capability. The

time integration is done using a leapfrog scheme, with a periodic predictor/corrector time-step. The monthly sea surface temperature and sea-ice boundary conditions constructed at Program for Climate Model Diagnosis and Intercomparison (PCMDI) (Taylor et al. 2000) are first interpolated on the LMDZ grid and then to daily values using cubic-splines (Hourdin et al. 2006). For the applications presented here, a uniform resolution of 3.75° in longitude and 2.5° in latitude with 40 levels vertical resolution is used, the time step is 3min. Rain is calculated as combination of the precipitation from the large scale condensation and rain condensate from convection at each pressure level. The rainfall profiles are obtained from averaging. In this study, the moist convection is parametrised by two schemes of parametrisation: the Tiedtke's (1989; hereafter TDK) scheme, used in previous versions of LMDZ and the Emanuel's (1991, 1993; hereafter EM) scheme which was shown to improve significantly the large-scale distribution of tropical precipitation and in turn the simulation of the mean climate and its variability in both atmospheric (Hourdin et al. 2006) and coupled ocean-atmosphere simulations (Braconnot et al. 2007). These schemes are based on a "mass flux" representation of the convection updraughts and downdraughts as well as of the induced motions in the environmental air.

In the TDK scheme, one convection cloud consists in a single saturated updraughts. Entrainment and detrainment between the cloud and the environment can take place at any level between the free convection level and the zero-buoyancy level. There is also one single downdraught extending from the free sinking level to the cloud base. The mass flux at the top of the downdraught is a constant fraction of the convective mass flux at cloud base. This downdraught is assumed to be saturated and is kept at saturation by

evaporating precipitation. The version used here is close to the original formulation of TDK (1989) and relies on a closure in moisture convergence. Triggering is a function of the buoyancy of lifted parcels at the first grid level above condensation level.

In the EM scheme, the foundations of the convective systems are regions of adiabatic ascent originating from some low-level layer and ending at their Level of Neutral Buoyancy (LNB). Shedding from these adiabatic ascents yields, at each level, a set of draughts which are mixtures of adiabatic ascent air (from which some precipitation is removed) and environmental air. These mixed draughts move adiabatically up or down to levels where, after further removal of precipitation and evaporation of cloud water, they are at rest at their new levels of neutral buoyancy. In addition to those buoyancy-sorted saturated draughts, unsaturated downdraughts are parameterized as a single entraining plume of constant fractional area (here 1 percent of the grid cell) driven by the evaporation of precipitation. The version of the EM's scheme used here is close to EM (1993). Closure and triggering take into account both tropospheric instability and convective inhibition.

EM's and TDK's mass flux schemes thus differ by several fundamental aspects. The triggering depends on atmospheric stability in both schemes (the max in the closure formula for EM's) but the closure does so only in EM's scheme. Also the ratio of the downdraught to updraught mass fluxes is limited for TDK (0.3 at the downdraught top) but not for EM (it can be occasionally greater than 1). Those differences were shown to produce significant differences for the simulated climate (Hourdin et al. 2006)

3. Results

a. Comparison of vertical rainfall profiles

First, Fig. 4 shows the average vertical profiles classified by regions typical of the different climate systems of West Africa. Following Geerts and Dejene (2005), we interpret the vertical variation of rain rate along a profile as the signature of some dynamic, thermodynamic and microphysics processes occurring to the water mass during its fall from 7.5km to the lowest available measure near the ground. Namely these processes are in a large part the rain production in the convection and the evaporation that occurs below the freezing level in the stratiform regions. The drop size distribution evolution along the profile is taken into account by the dynamic adjustment of the Z-R relationship (Iguchi et al. 2000).

The observed average vertical rainfall profile in continental Guinean regions (10°W-10°E; 5°N-10°N) corresponds mostly to equatorial forests (Fig. 4. a) and shows an almost constant value from the freezing level to the ground. This is somewhat what is expected in this region where convection is not very intense and relative humidity is extremely high in the low levels. A rather weak updraft will produce precipitation very early during the ascent and evaporation is almost non existent. Above the supposed freezing level (approximately 4.5 km) the amount of precipitation decreases rapidly due to the shallow character of convection that does not transport large amounts of supercooled water or ice phase aloft. It is possible that this decrease is amplified in the observed profiles by the comparatively lower sensitivity of the radar to the ice and specially the small ice particles. The black square at the surface in Fig. 4 shows the average surface rain over the same area obtained from CMAP data. The mere averaging

of PR rainrate cannot be fundamentally compared to CMAP estimates, but it provides a reference in terms of rain intensities. So, the CMAP data is assumed to be a qualitative and indirect validation of rainfall statistics built with TRMM-PR. EM's profile is not too far from observation below 4.5km but show a different concavity above. Below the freezing level, the vertical gradients are close to TRMM's but with a slightly more marked downward increase. TDK exhibits a rather different behaviour, the location of the main features and their intensities are quite different, although it is closer to observation in shape. This is particularly noticeable at the sharp transition at 2.25 km that occurs in the observation near the freezing level. Once again, the downward increase is slightly more pronounced below the transition region above the transition, the profile decreases rapidly with increasing altitude in a fashion that is close to the observation. The three profiles offer very different intensities at most altitudes.

In the semi-arid Sahelian region 10°W-10°E, 10°N-17°N (Fig. 4.b), the situation is quite different. The TRMM-PR rainfall profile decreases downward consistently from the bright band position (4.25 km) to 1.25 km. In our opinion, this is due to the combined effect of both the evaporation that reduces the water content as one gets closer to the ground in the stratiform regions of the cloud but also to the condensation/accumulation level that is near 4 km in the more convective regions. One must notice that the latter has a stronger effect on the profile shape (vertical gradient of rain rate) than the evaporation itself. The rain intensity above the freezing level remains high when compared to the previous case which is the likely signature of more intense convection. Although not having the same intensity, the TRMM-PR profile and the TDK scheme profiles show some similarities in terms of shape both in their upper and lower parts. Their intensity is

also closer than in the previous case and the transition region position on the TDK profile is somewhat higher than in the previous case. Between 5.75 km and 7.5 km, the TDK and PR profiles are merged. Below 5.5 km, TDK's profile is always below the observation in terms of intensities. The general features of the EM's profile are not so different but for the freezing level which is not well defined: it seems to stretch between 4.75 km and 3.5 km. On the other hand, below 4 km the downward decrease is well characterised with a rate similar to both observation and TDK. The rainfall rate at all levels of the EM profile is higher than in PR and TDK profiles. The comparison with CMAP data suggests qualitatively that TRMM-PR underestimates rainfall at the surface in the Sahelian region.

Under Saharan latitude (Fig. 4.c) the situation is again different. The observation does not show a very marked freezing region. From TRMM-PR observation, it seems that the Saharan rain is not very intense in average which is the signature of a rather weak convection. This leads to a weaker water accumulation near the freezing level while evaporation might eventually remain strong. This situation is induced by the African monsoon flow which drives a precipitation gradient between Sahara and Sahel. The Sahel region is characterised by rains mainly due to strong and well organised convective systems. The freezing level on both TDK and EM profiles is not very well characterised either, but it seems to be in a higher position than in the observations from PR. Both profiles follow the same variation as the TRMM-PR profile with a downward decrease from 4 km although it is slightly stronger in EM's scheme. TDK profile is close to PR profile in terms of intensities while EM is always stronger. TDK reproduces best the decrease with altitude in the upper levels, but this decrease seems to start at a lower altitude.

It should be noticed that the observation show rather different features for the three different regions with a consistent evolution along the South-North transect. The two parameterizations show somewhat a similar series of changes from the Guinean coast to Sahara.

b. Comparison of Vertical Gradient seen by observations and model's schemes

This part deals mostly with the low levels (below 4km) in terms of re-evaporation and liquid water production. Following Hirose and Nakamura (2000, 2004), the "Index of Vertical Gradient" (IVG) was computed to quantify the vertical gradient of rainfall rate. These authors computed their index between 2 and 3.5 km. Similarly, the IVG is defined as the first derivative of the rainfall rate between altitude H1 and H2 but the results from section 2 on rainfall profiles comparison compelled us to choose 2.25 and 1.5 km as reference altitudes (Fig. 6). IVG is expressed in $mm.day^{-1}km^{-1}$ and the sign of this gradient suggests the importance of the production/evaporation processes along the profile. The PR data are first averaged on a 2°x2° grid matching the model resolution before the IVG is computed. Assuming that attenuation of reflectivity is well corrected in the 2A25 product (e.g., Bolen and Chandrasekar 2000) before the calculation of the rainfall rate, positive values of IVG correspond to evaporation processes in the stratiform rains and accumulation of raindrops due to updraughts in the more convective cells. On the other hand, negative values are the signature of liquid water production in the lower levels. Fig. 5 and Fig. 6 present respectively the map and the zonal average of IVG as obtained from observation and from LMDZ with both TDK and EM parameterizations. The zonal variation of IVG is studied over the continental region; we have performed a

zonal mean between 5°W and 5°E. The choice of this interval is meant to minimize the influence of the relief, i.e. the influence of rain mostly driven by orography.

The IVG computed from TRMM-PR shows negative values over ocean (between 15 °N and the equator) and in the Guinea regions (Coast and south-west of West Africa) where water production is expected when going from the freezing level to the ground. Strong negative IVG are also found around relief regions like Fouta Djallon (south east of Senegal) in Cameroon hills, in Hoggar Mountain and in the Darfur region (south east of Tchad); it is interpreted as low-level liquid water production from shallow precipitation or systematic events with low levels convergence in these regions. The PR shows also quite consistently positive IVG over most of the rest of the continental regions where both re-evaporation and water accumulation is expected to be the strongest (Fig. 5.a). The maximum of re-evaporation is located in the central Sahelian region (12°N-15°N) with a rate around +0.4 $mm.day^{-1}km^{-1}$. The minimum is slightly below +0.1 $mm.day^{-1}km^{-1}$ and is detected over the Sahara region. Zonal average of IVG obtained with TRMM-PR observations illustrated in Fig. 6, shows that re-evaporation processes appear over the continental areas between 7°N and 24°N and liquid water production processes are dominant below 7°N.

IVG computed with TDK's scheme exhibits a pattern quite close to observation. Differences are noticed with the region of negative IVG spreading too much over the continent up to 13°N. This northward extension is clearly visible also in the zonal averages (Fig. 6). In addition, positive IVG signatures are a bit too intense and extend too far (up to 20°N) when compared to observation (Fig. 5b). The maxima of IVG are observed around 16°N in Mali and around the Darfur plateau. With EM's scheme,

negative IVG are observed over most of the continent and somewhat over ocean in the south-west corner of the domain. The magnitude of the IVG is very well represented in the Sahel (9°N-18°N) (Fig. 5c) but notably overestimated in the Sahara. This over estimate is likely associated with a much too high production of precipitation in the middle troposphere (as shown in Fig. 4.c). The regional spots of negative IVG seen by TRMM-PR in the Gulf of Guinea, in the East Atlantic Ocean and around mountain regions, are well depicted by EM's scheme. However, a small patch continental negative IVG is observed over the west side of the Guinean regions and the south-western side of the domain over ocean (between 26°W and 10°W).

Analysing Fig. 6, it appears that both EM and TDK schemes follow the observation over the landmasses. Differences can be seen for instance with the northward shift of the IVG=0 $mm.day^{-1}.km^{-1}$ TDK's scheme around 13°N (Fig.6.a) and 15°N (Fig.6.b). This is likely due to the strong overestimated amplitude of the minimum when compared to the observations.

On the other hand, EM scheme follows quite well the zonal variations of IVG seen by TRMM-PR observations (Fig. 6) with only a little shift of about 1.5° to the North. In the Sahel region, the amplitudes of EM's scheme and the observations are practically identical with only minor differences where EM scheme depicts more evaporation than the observation (Fig.6.a). The Saharan latitudes are characterised by a higher rate of atmospheric re-evaporation on EM scheme with a maximum near 21°N which is not present in the observations. The latter show a very low evaporation rate in Sahara. EM's parameterization seems to depict a good meridional structure of IVG except in Sahara where it is overestimated.

Liquid water production processes as well as re-evaporation processes obtained from the observations are roughly depicted by the two schemes of LMDZ models but with local differences. EM's scheme seems to be the best for representing the average meridional structure of re-evaporation processes in mid-Sahelian latitudes where it follows very well the TRMM-PR observations. It slightly underestimates the water production in equatorial regions, and extends the negative IVG region to 8.5°N where observations place it at most at 7°N. Evaporation in the Sahara region is overestimated by EM's scheme with a maximum rate located in the upper Sahara. On these same zonal averages TDK's scheme strongly overestimates all the processes when compared to the observations; it depicts liquid water production until 13°N with a strong intensity. The evaporation rate in the upper Sahel (15°N-17°N) is also overestimated.

4. Summary and conclusions

The main objective of this paper was to study and compare vertical rainfall profiles during West African monsoon season computed from TRMM-PR data and from two different convection parameterizations scheme in the LMDZ model (Tiedtke and Emanuel).

The average rain profiles obtained with the two schemes present both similarities and differences when compared to TRMM-PR profiles in both the upper and lower layers. The rainfall rate profiles (averaged between 10°E and 10°W) seen by Emanuel's scheme fails systematically to represent the upward decrease over 4km. However, it catches the downward decrease seen in the observation from 4km to the surface especially in Sahel (proper depth of the layer dominated by re-evaporation in stratiform regions/water

accumulation in convection). As for Tiedtke's scheme, it provides a better description of the upward decrease in the upper level, whereas the decrease observed in the low levels starts too low (about 2 km in Sahel) compared to the observation.

The index of vertical Gradient was computed from observation and compared to both parameterizations to assess how re-evaporation associated with convection and water production processes are reproduced by the LMDZ climate model in the low levels of the atmosphere. The horizontal structure of IVG has shown the influence of the low level processes which influence the convective activity and how the two model parameterizations translate it. The Tiedtke's scheme seems to be able to reproduce the general structure of both negative (mostly over ocean) and positive (mostly over land) IVG but with a signal that is generally too strong in amplitude and shows some local discrepancies at the ocean-land transition. Emanuel's scheme gives a much better range of intensities but its general pattern seems further from the one given by the observations. Nevertheless, because the intensities are better represented the match is quite good. However, Emanuel's schemes tends to show a slightly too strong signal over the Sahara. When comparing the zonal averages the results are somewhat different. Liquid water production processes as well as re-evaporation processes obtained with observations are roughly depicted by the two schemes of LMDZ models but with local differences. It seems that Emanuel's scheme offers a much better meridional gradient on average, both in terms of structure and intensities, than Tiedtke's scheme.

The behavior of parameterizations used here (as said in Hourdin et al (2006)) is probably related by the fact than the updraught in Tiedtke’s parameterization is an entraining plume. Both its vertical extension and intensity are thus sensitive to the humidity of the

free troposphere. While in Emanuel's scheme, the adiabatic updraft does not entrain air from the free troposphere so that the cloud top is always at LNB. A dry free troposphere can reduce convection by modifying the humidity of the mixed draughts, but not limit its vertical extension.

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FIG 1: Zonal percentage of lost pixels from 4km to ground at latitudes with the three datasets: (right) full dataset, (middle) 10-pixel sampling dataset and (left) only the central pixel sampling. a) over land, b) over Ocean

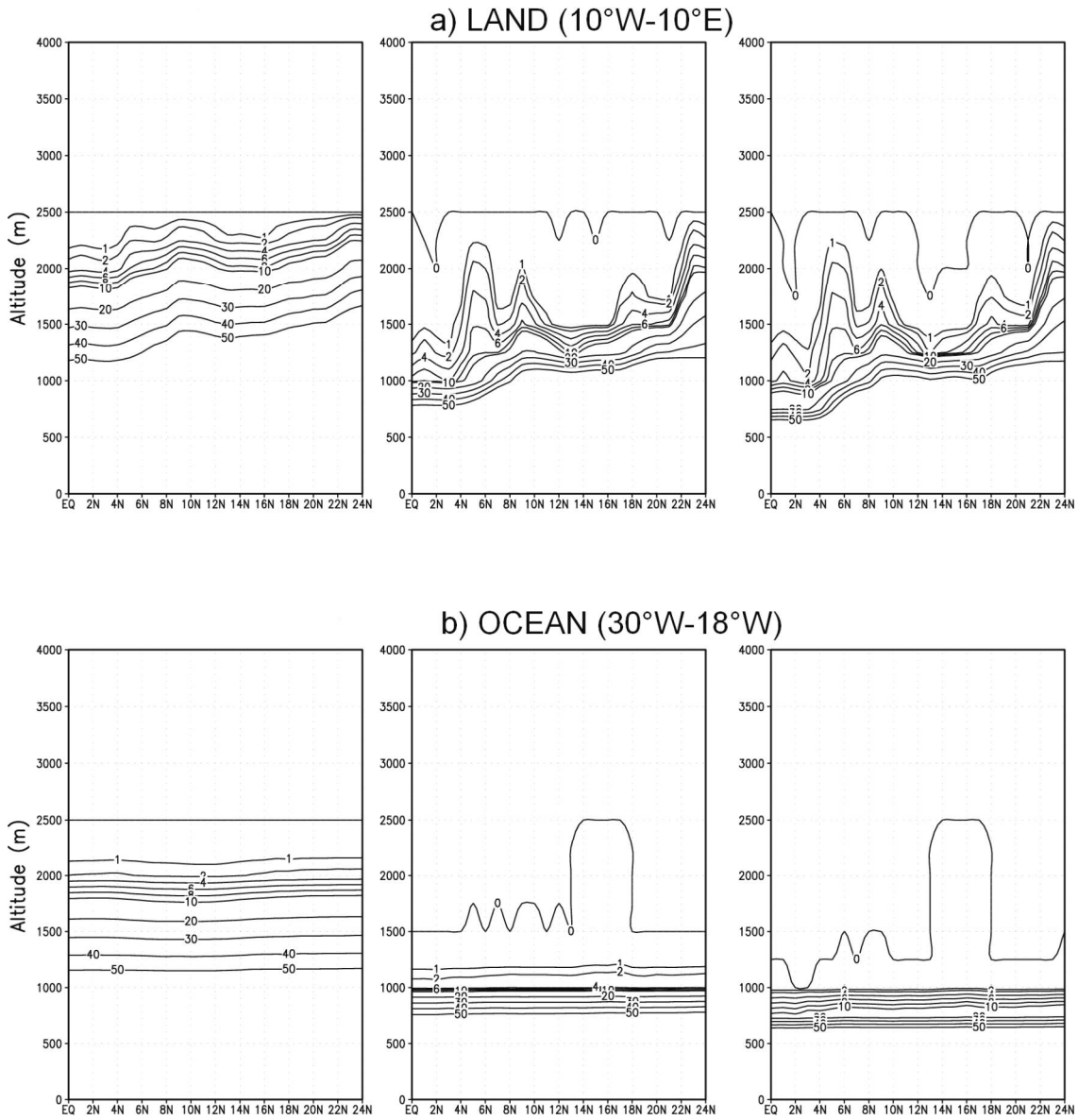
FIG 2: Average zonal vertical precipitation (mm hr^{-1}) contours obtained from PR (threshold 10% of lost pixel) from 4 km to ground using only the central pixel sampling. Solid lines indicate percentage of lost pixel from 4 km to ground, similarly to Fig. 1.

FIG 3: Same as Fig. 1 but for the zonal average of the vertical gradient of rainfall rate mm $hr^{-1}km^{-1}$ from 4km to ground. Dashed contours indicate regions of downward decreasing rain rate and solid contours regions of downward increasing rain rate.

FIG 4: Mean vertical rain profiles over land (10°W-10°E) obtained by TRMM-PR and LMDZ output. TRMM-PR in solid lines and LMDZ output: Emanuel scheme bold dash line and Tiedkte scheme dash line. The black squares at 0km altitude indicate precipitation at surface obtained with CMAP data.

FIG 5: Distribution of IVG computed between 2.25km-1.5km altitude from observation (using 10 pixel sampling dataset) and model. LMDZ output is from climatological reference of 1981. TRMM-PR is the average over JJAS from 1998 to 2006

FIG 6: Meridional IVG variation averaged between 5°W and 5°E. TRMM-PR in solid line (using 10-pixel sampling dataset) and LMDZ output: Emanuel scheme as bold dashed line and Tiedkte scheme as short-dashed line.



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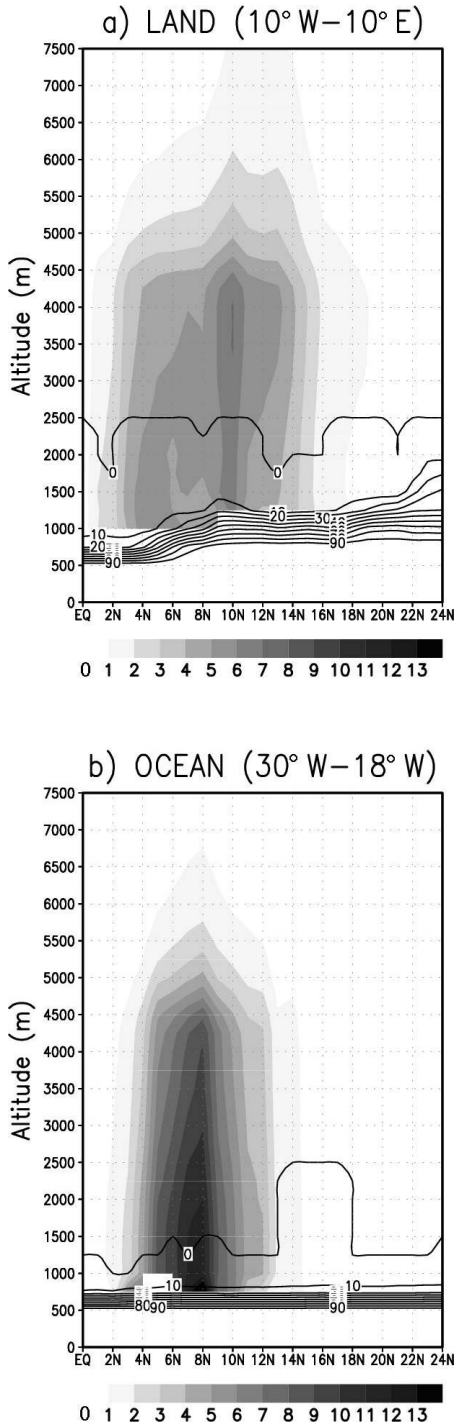

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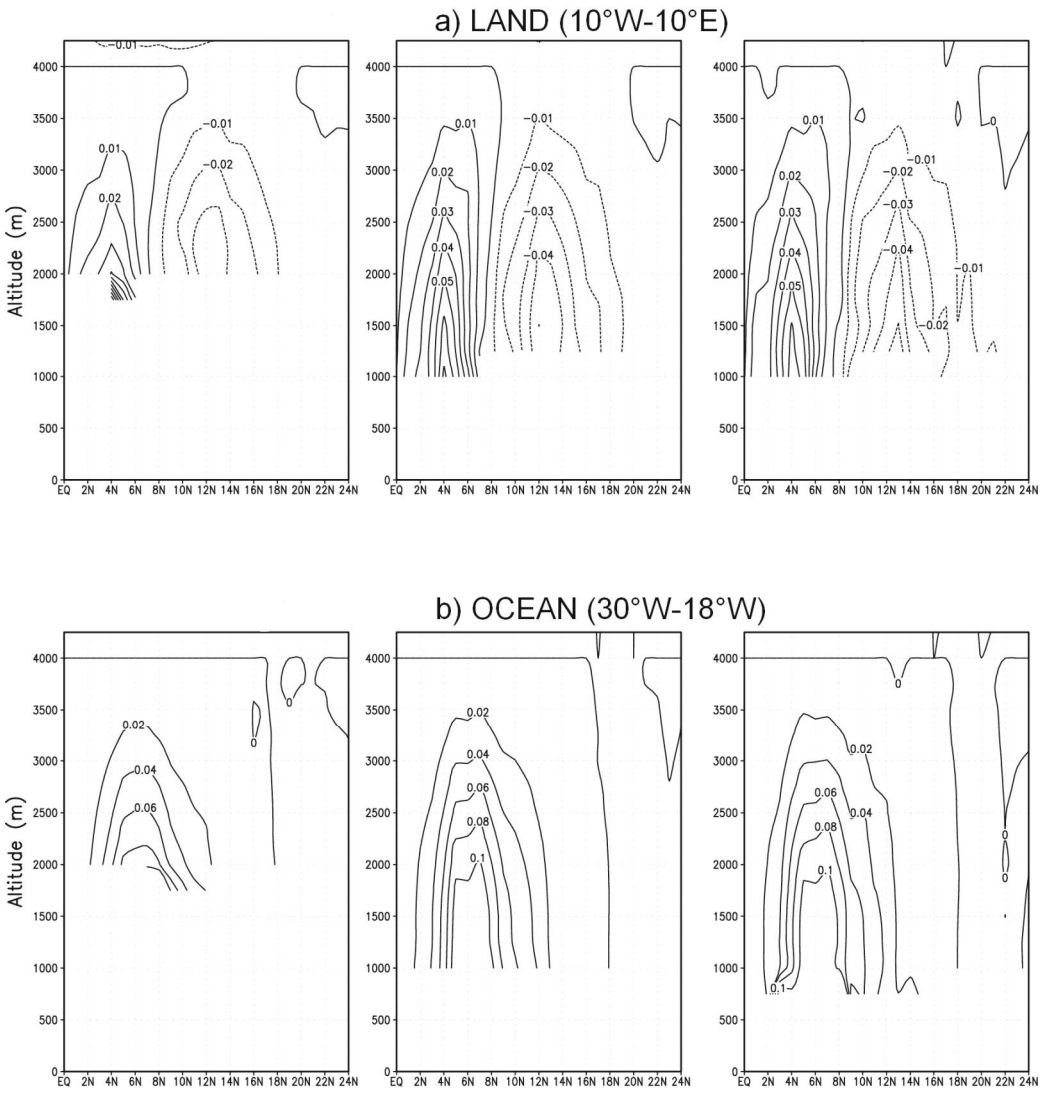

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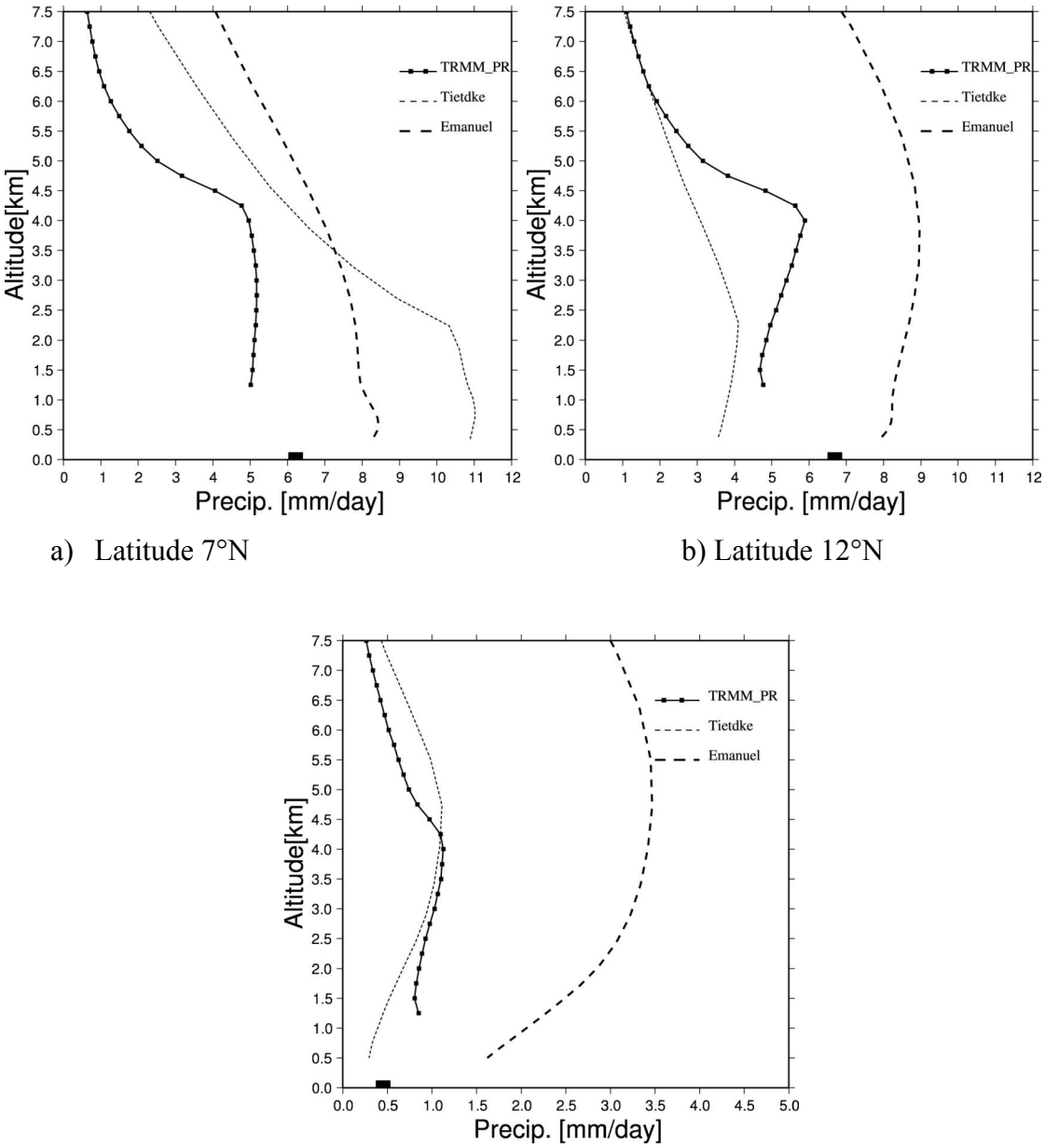

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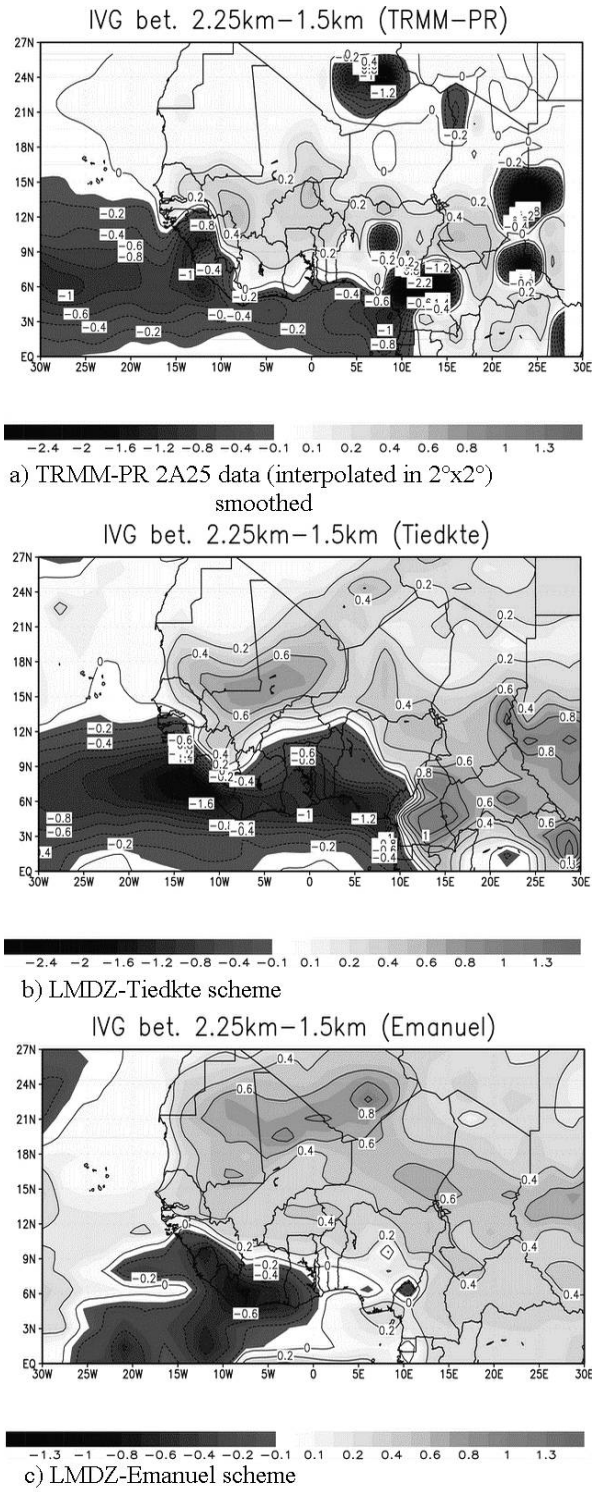

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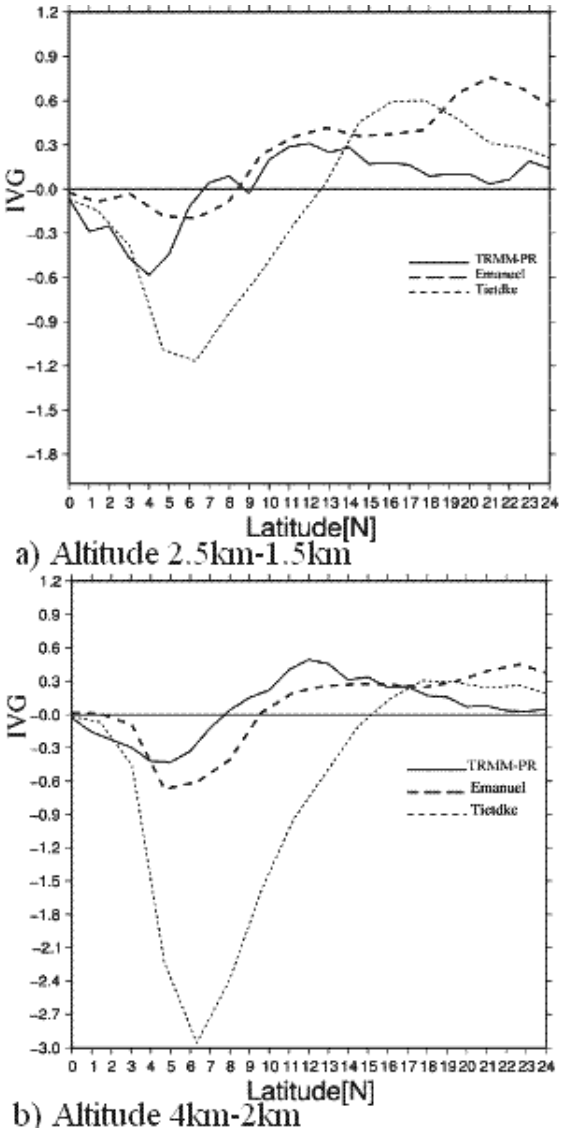

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