Treatment, analysis of SCMs runs and report on the RCE simulations

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1 Continental RCE case

Simulating moist processes over land requires the coupling of the atmospheric model with a Land Surface Model (LSM). The complexity of such a coupled model makes it difficult to tackle the problem of the various biases observed in climate models over land (such as the too early maximum of convective precipitation during the day or the warm bias over mid-latitude continents (ref ?)). The purpose of this subtask is to provide a simplified framework where the interactione between surface conditions and moist processes may be analysed more easily.

1.1 Motivations

The specificity of the atmospheric processes over land lies in the high value of the surface roughness (about an order of magnitude larger than over ocean), in the large range of possible surface moisture values and in the important role of the diurnal cycle of surface conditions. These three features make the boundary layers over land very different from the oceanic ones. Especially they exhibit a large range of nature and depth, from stable layers few hundred metres deep (at night) to convective layers three to five kilometres deep (over dry areas, during daytime). Many of these specific features can be explained by local conditions and processes : they are more or less independent from synoptic or continental scale processes. For instance, the boundary layers over Sahel can be mainly described as a boundary layer over a semi-arid area, even though the monsoon flux and the AEJ are important features of the Sahel climate. Consequently one may expect numerous features of moist processes over land to be independent from synoptic or continental scale processes. In order to focus on such features, the present case is devoted to the study to Radiative-Convective Equilibrium over land.

Instead of attempting to deal with the full physics of land surface - convection coupling, the present subtask defines a simplified framework in which only a part of the problem is adressed : we assume that the main cause for the diurnal cycle of moist processes is the thermal response of the ground to the SW forcing ; thus we replace the surface hydrology by a prescribed aridity coefficient β (equal to the ratio of the actual evaporation to the potential evaporation) and reduce the LSM to a mere heat diffusion model. Hopefully, such a device should keep the main features of the diurnal cycle of moist convection while removing the complexity of LSMs. The first results presented in subsection (1.3) show that many features of moist processes over land are indeed simulated.

1.2 Case specifications

In this land-RCE case, the atmospheric model is isolated from all external interaction except the sun and the soil. The land surface model is devoid of any hydrology. Instead, the ratio β of evaporation to potential evaporation is fixed. Then some temperature has to be fixed : we nudge the soil temperature at a depth such that the diurnal cycle is present.

Even with such a drastic simplification there remain a large family of possible cases. First, of course, the aridity coefficient β and the target soil temperature may vary over a large range of values. Second, the inertia ($I = \sqrt{\lambda\rho C}$, where λ is the thermal conductivity, ρ the density and C the specific heat capacity) (I is also called the effusivity) may vary depending on the soil properties and on the soil moisture. Obviously, on a given soil, β and I are strongly correlated ; however, we shall keep the inertia $I = 2000$ KW^{-1} constant in the present case-study. Then, the astronomical conditions may be varied : the case depends on the latitude, the day of the year and whether one keeps the seasonal cycle or simulates a fixed day of the year. The large scale wind may be chosen arbitrarily, thus providing a range of surface fluxes and of vertical shear. Finally, the domain size may play a role if scale aware parametrizations are used : it may be similar to a climate model grid cell (say 100×100 km^2) or to the whole inter-tropical band (ca 10^{14} m^2).

Since our purpose is to study the workings of parametrizations, we restrict ourselves to a very limited set of cases : the day of the year is fixed to march 21 (although it may be interesting to study the role of the length of the night) ; the wind is held uniform at 10 ms^{-1} ; the latitude is set to 10N and the domain size to 10^{14} m^2.Finally, the RCE case represents a 2-parameter family of cases. The two parameters are :

- Aridity coeficient β (between 0 and 1).
- Target temperature in the ground.

In order to implement the case a Fortran subroutine (named "surf_land_x") is provided which simulates the soil evolution and its coupling with the atmosphere. This subroutine should be called in lieu of the standard module coupling the atmospheric model with the land surface. Subroutine "surf_land_x" simulates the dry soil evolution and its coupling with the atmosphere. It calls subroutines "soil_x" and "fluxs_x" which compute the soil temperatures and the surface fluxes respectively.

Surface fluxes obey the sign conventions of Fig. 1. Energy conservation at the surface reads :

$$\Phi_{\mathrm{SW,net}} + \Phi_{\mathrm{LW,net}} + \Phi_{\mathrm{lat}} + \Phi_{\mathrm{sens}} + \Phi_{\mathrm{ground}} = 0 \tag{1}$$

An implicit scheme is applied to the heat conduction equations in the ground. The discretized equations are solved through a standard LU decomposition, which yields two loops : (i) the upward loop starts from the bottom boundary condition (zero flux condition) and uses the present temperature profile to compute the coefficients necessary for the new temperature profile computation ; after this loop the surface temperature is computed ; (ii) the downward

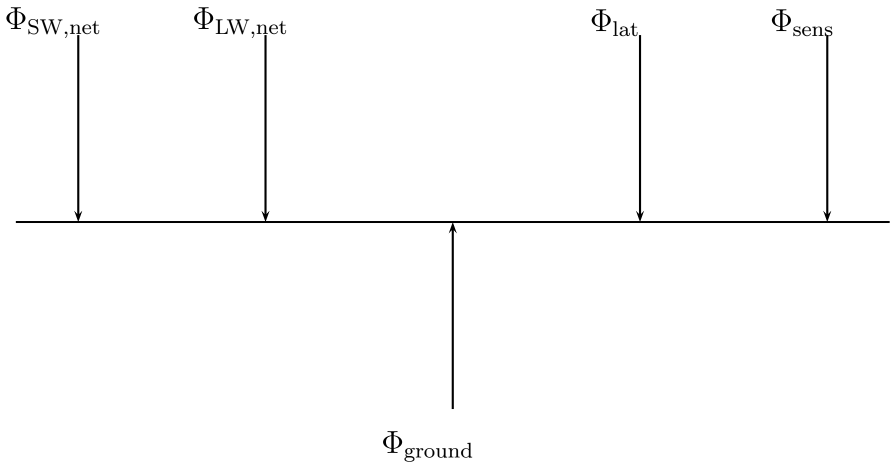

FIG. 1 – Flux sign convention

loop starts from the surface temperature and computes the new temperature profile thanks to the coefficients determined during the upward loop.

The coupling between the boundary layer and the soil models is done by solving the system of Eq. (1) together with the PBL surface equations :

$$\begin{aligned}
\Phi_{\mathrm{sens}} &= \rho C_p C_D (1+\sqrt{u^2+v^2})(T_1 - T_s) \\
\Phi_{\mathrm{lat}} &= \rho L_v C_D \beta (1+\sqrt{u^2+v^2})(q_1 - q_{\mathrm{sat}}(T_s)) \\
C_p T_1 &= A_T \; + \; B_T \Phi_{\mathrm{sens}} \delta t \\
L_v q_1 &= L_v A_q \; + \; B_q \Phi_{\mathrm{lat}} \delta t
\end{aligned} \tag{2}$$

and with the soil surface equations :

$$\Phi_{\mathrm{ground}} = F_g \; + \; G_g T_s \tag{3}$$

where A_T, B_T, F_G and G_G are coefficients computed during the downward loop of the atmosphere vertical diffusion model and during the upward loop of the soil model, C_D is the drag coefficient and (u,v) is the wind speed at first level. These six equations determine the six variables : the fluxes Φ_{sens}, Φ_{lat} and Φ_{ground}, the surface temperature T_s and the first atmosphere level temperature T_1 and humidity q_1.

Since this case is quite new and in a very preliminary stage, we propose a limited set of simulations. Simulations should be long enough that the precipitable water reaches a nearly constant value : generally this takes a few months. The soil temperature should be nudged towards 300 K at the third level of the soil grid with a rexation time of 3600 s. The aridity coeficient β may take the following values :

$$1.,\ 0.75,\ 0.5,\ 0.25,\ 0.1,\ 0.05,\ 0.01$$

In order to characterize the workings of parametrizations, some plots are especially relevant :

– Vertical profiles of tendencies averaged over ten days.
– Time series of diabatic heating and drying and of precipitation during two or three days.
– time series of precipitation during a few days.

1.3 Some results

So far, this case has been simulated only with the LMDZ SCM. It still needs some reference results from LES or CRM simulations. Simulations from several SCMs would make it possible to study the range of continental climates and climate sensivities yielded by the various models. However, lacking such a variety, we shall content ourselves with results from the LMDZ SCM, using the LMDZ5B physics for the two first items and comparing three physical packages in the last item..

1.3.1 A modelling experiment on the link between soil moisture and ice cloud frequency of occurence

Using satellite observations, Prigent et al. (2011) analysed the impact of the inundation occurrence on the deep convection at continental scale. They focused especially on regions where the inundation is not generated by local precipitation, i.e. on regions where the inundation appears as an external forcing for the local climate. They used ice concentrations in clouds as a measure of deep convection intensity. Their main finding was that stronger convection happens in these regions during the minimum of the inundation, with a marked diurnal cycle of the deep convective activity.

In order to further this analysis, they performed a sensitivity analysis of deep convection to soil wetness in SCM simulations of Radiative Convective Equilibrium over land. Using the beta model they compared two situations associated with two values of the aridity coefficient $\beta : \beta = 0.7$ for the wet case and $\beta = 0.25$ for the dry case. The SCM simulated a convection stronger in terms of ice concentration and with a larger diurnal cycle amplitude in the dry than in the wet case, in agreement with observations.

1.3.2 Behaviour of the physical package LMDZ5B for various values of the aridity coefficient β

Various simulations of the land-RCE case were performed with the LMDZ SCM using LMDZ5B physics (ref??). We present some results of three simulations corresponding to $\beta = 1., 0.1, 0.01$.

Fig. (2) displays the time series of the precipitable water from the three simulations. It takes four to six months for the system to reach a quasi-steady regime. The values obtained for $\beta = 1.$ and 0.1 (about 75 kg m^{-2}) are reasonable for a wet maritime case, while those obtained for $\beta = 0.01$ (about 45 kg m^{-2}) are reasonable for rainy conditions of semi-arid regions. The diurnal cycle of surface fluxes and their variability is presented in Fig. (3). The day to day variability is due to the variability of the surface SW flux, i.e. to the variability

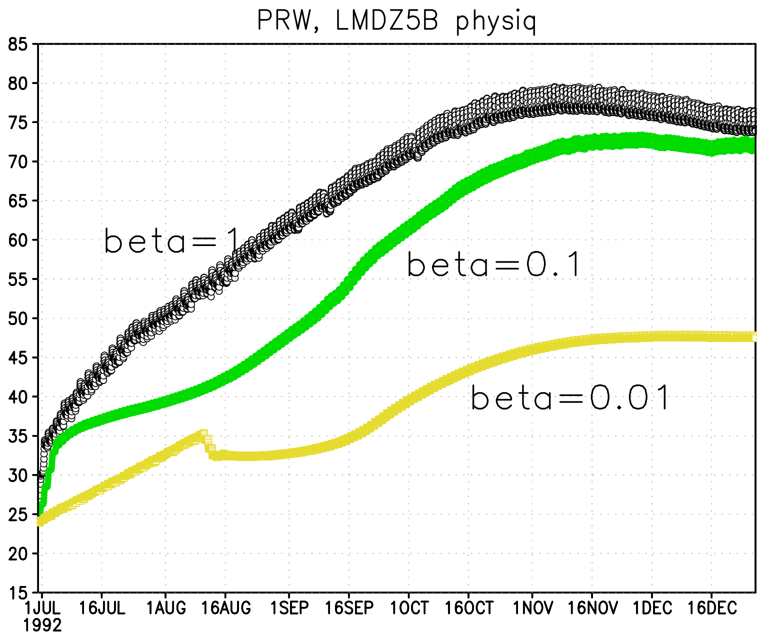

Fig. 2 – Time series of Precipitable water (kg m^{-2}) during the six months of the simulations (LMDZ5B physics).

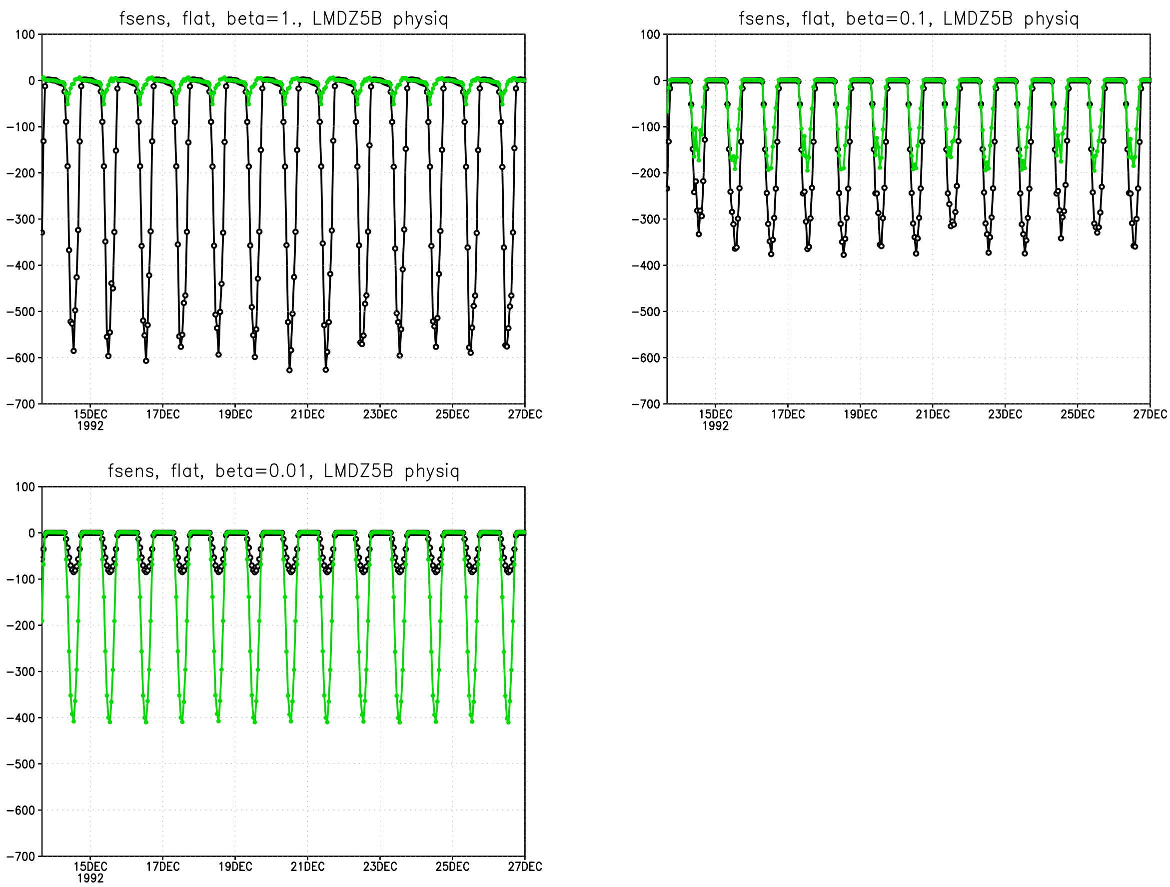

Fig. 3 – Latent (black) and sensible (green) surface fluxes for various values of coefficient β duting two weeks in the last month of simulation (LMDZ5B physics).

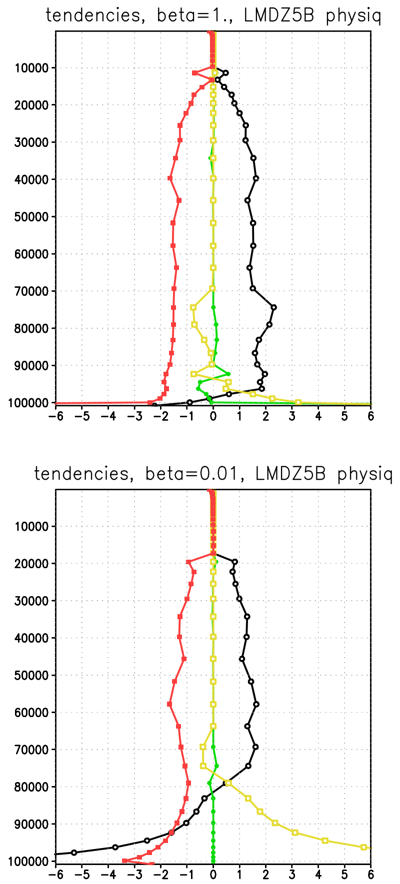

FIG. 4 – Vertical profiles of temperature tendencies for beta=1. (uppet left panel), beta=0.1 (upper right panel) and beta=0.01 (lower panel). Black : deep convection, red : radiation, green : large scale condensation, yellow : boundary layer. (LMDZ5B physics).

of cloud cover. In the $\beta = 0.01$ case, clouds disappear completely at night so that days are almost identical. The evaporative fraction ranges from 8% in the driest case to 30% in the $\beta = 0.1$ case and to 80% in the wettest case.

The time-averaged temperature tendencies associated with the various physical parametrizations are displayed in Fig. (4). Over the whole free troposphere the radiative cooling is mostly compensated by the deep convection warming. Shallow convection does play a significant role up to 650 hPa ; especially above the boundary layer the evaporation of cumulus clouds induces a cooling of the order of 1K/d, but weaker in the driest case. The boundary layer thickens with the dryness of the soil from less than 1 km in the weetest case to more than 2 km in the driest one.

The apparent heat source due to deep convection is displayed for two days in Fig. (5). The maximum heating occurs between 600 hPa and 300 hPa for the two wet cases and between 750hPa and 500 hPa in the driest case. These features need some confirmation. The below cloud cooling is thin and weak in the wet case and grows stronger and thicker when dryness increases.

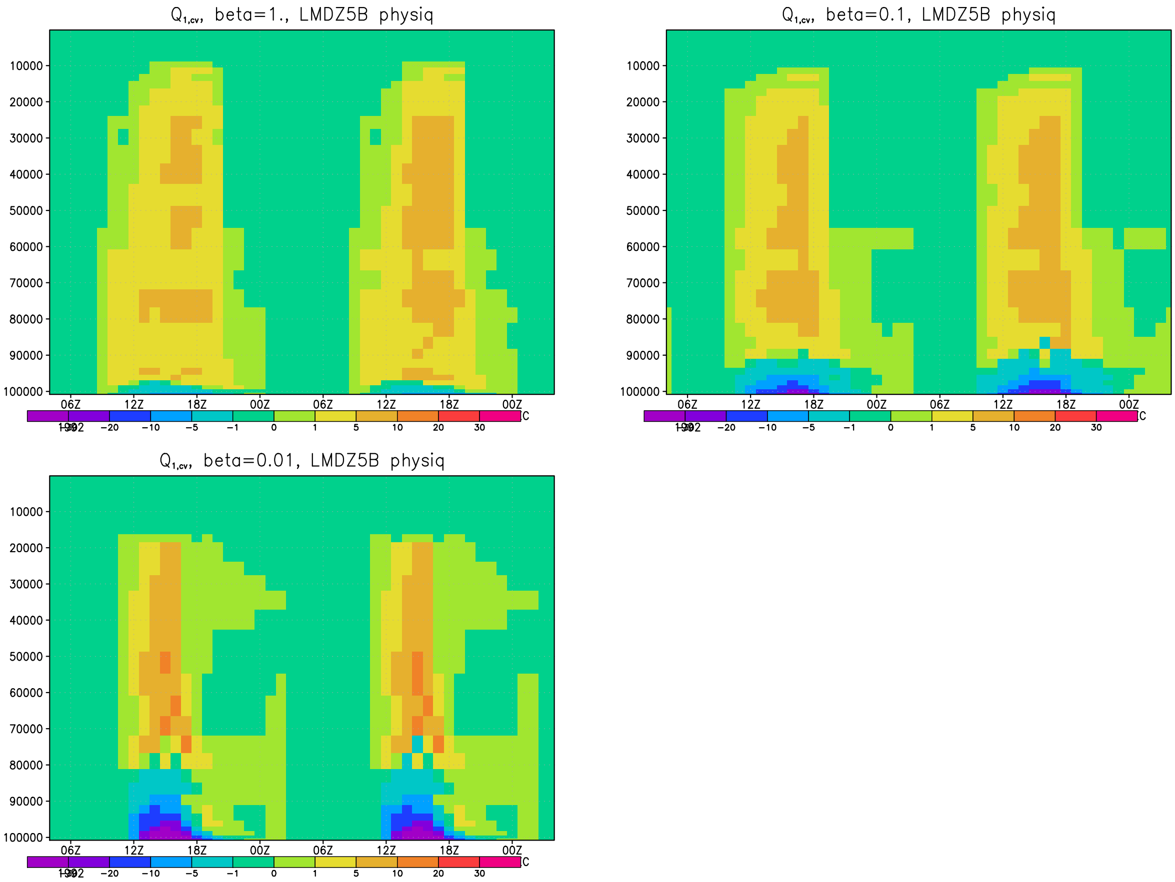

FIG. 5 – Time-altitude plot of convective heating tendency for beta=1. (uppet left panel), beta=0.1 (upper right panel) and beta=0.01 (lower panel) (LMDZ5B physics).

1.3.3 Comparison of three physical packages of LMDZ5 : LMDZ5A, LMDZ5B, LMDZ5 with Tiedtke convective scheme

Some simulation results obtained with the LMDZ SCM using three physical packages are displayed in Fig. (6).

As concerns the vertical profiles, the LMDZ5A physics looks peculiar. This is due to the homogeneization of convective tendencies below cloud base. However, one should note that the cloud base is at similar levels for LMDZ5A and LMDZ5B, showing that the boundary layer humidity is very close in the two radiative-convective equilibrium reached by the two models. In the $\beta = 0.05$ case the LMDZ5-Tiedtke version yields a weaker convection than the two other versions, a feature wich will need further analysis.

The diurnal cycles of precipitations appear well shifted relative to one another. The LMDZ5-Tiedtke yields both the ealiest initiation and maximum of precipitation. The LMDZ5A version comes an hour later, but with strong precipitation that stops earlier than the LMDZ5-Tiedtke one. The LMDZ5B version yields an initiation of precipitation an hour later and a maximum of precipitation delayed by five hours relative LMDZ5A.

References

- Prigent, C., N. Rochetin, F. Aires, E. Defer, J.Y. Grandpeix, C. Jimenez et F. Papa. Impact of the inundation occurrence on the deep convection at continental scale from satellite observations and modeling experiments. Journal of Geophysical Research, vol. 116, no. D24, page D24118, 2011.

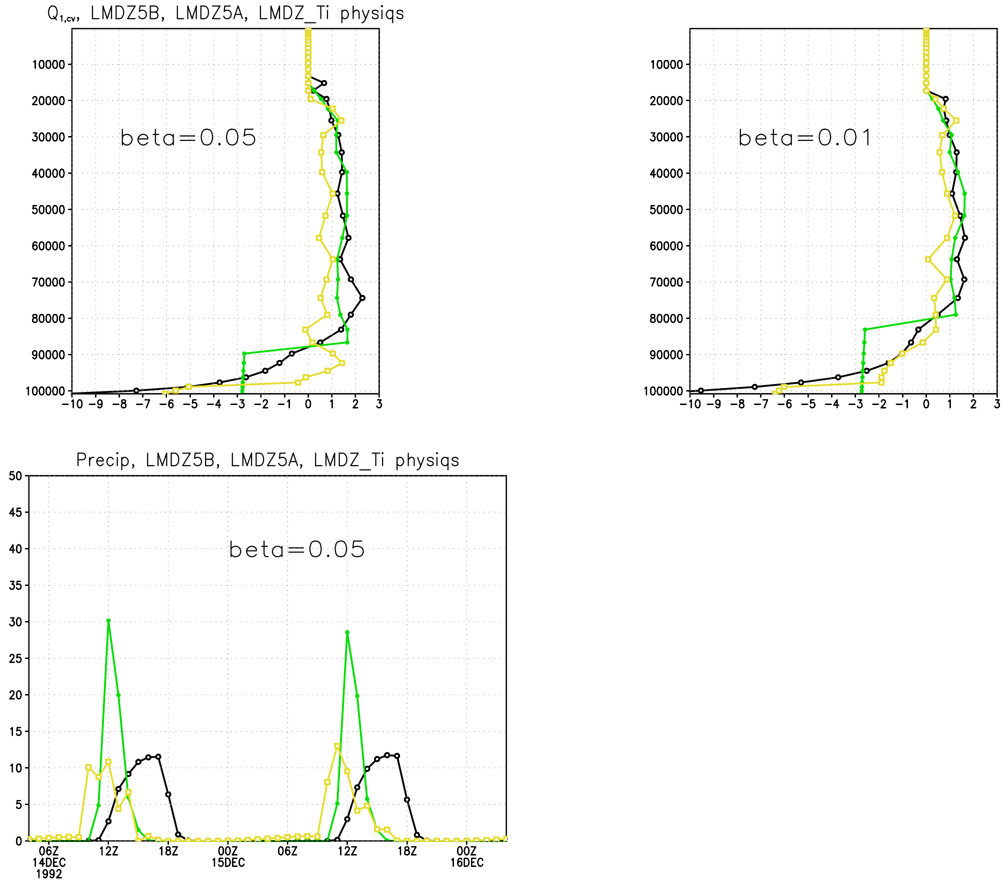

FIG. 6 – Comparison of LMDZ5A, LMDZ5B and LMDZ5 with Tiedtke convection scheme. Upper panels : vertical profiles of convective heating, averaged over two weeks, for the three physical packages and for $\beta = 0.05$ (left panel) and $\beta = 0.01$ (right panel). Lower panel : diurnal cycle of precipitation. Black : LMDZ5B, green : LMDZ5A, yellow : LMDZ5 with TIedtke convective scheme.