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# *Spurious effects of the deep convection parameterization on the simulation of a Sahelian heatwave*

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## ***Abstract***

A severe heatwave occurred in April 2010 over West Africa. It was characterised by a particularly high daily minimum temperature reaching more than 35°C locally and a high water vapour content. In this study we analyse the ability of a mesoscale limited area model to represent such an event and investigate the advantage of using an explicit representation of deep convection for such a case associated with very limited precipitation amounts. Two high-resolution simulations (5 km x 5 km horizontal grid) have been performed from 10 to 19 April 2010; they are identical except that one uses a deep convection parameterization (simulation PARAM) and the other does not (simulation EXPL).

These simulations are evaluated with different observational datasets including gridded products as well as local meteorological measurements and radiosoundings. Overall, both simulations display a negative temperature bias in the low levels but this bias is much more pronounced in PARAM, mainly due to evaporative cooling of spurious precipitation.

Indeed, in PARAM, precipitation is too frequently triggered (around mid-day, i.e. several hours too early) and too strong; the Inter-Tropical Discontinuity (ITD) propagates too far north during this 10-day sequence. Conversely, in EXPL, the observed northward shift of the ITD is correctly simulated and precipitation displays a better timing, variability, intensity and latitudinal extent. It

24 thus appears that the representation of deep convection affects the atmospheric circulation  
25 associated with the heatwave event.

26 The mechanisms involved in this humid heatwave are further investigated with thermodynamic  
27 and dynamic budgets which also underline the main differences between the two simulations. A  
28 proper representation of deep convection on sub-diurnal time scale turns out to be necessary for the  
29 simulation of this heatwave episode, which points to the interest of convection-permitting  
30 simulations for the study of heatwaves even though they are generally characterised by very little  
31 precipitation.

32 *Key Words: Convection-permitting model, Deep convection parameterization, Heatwave, Inter-*  
33 *Tropical Discontinuity, Monsoon Surge, Sahel, Thermodynamic and dynamic budgets*

## 34 **1 INTRODUCTION**

35 Global mean surface air temperature has increased by 1.1°C since 1900 (IPCC 2021). This  
36 warming is more pronounced over land than ocean (Sutton et al. 2007) and generally stronger  
37 during night-time than daytime (e.g., Easterling et al. 1997, Zhou et al. 2010, Harris et al. 2014).  
38 The global warming is also accompanied by an increase of extreme weather events in frequency and  
39 intensity like droughts, floods, cyclones and HeatWaves (HW) (Seneviratne et al. 2012, Morales et  
40 al. 2020). In the future warmer climate, as projected by climate models, contemporary extreme  
41 temperature events will become more frequent and warmer, will last longer and will cover more  
42 extended areas worldwide (Meehl and Tebaldi 2004, Stott et al. 2004, Russo et al. 2014). They have  
43 been analysed in detail over Europe (Christidis et al. 2020, Schoetter et al. 2015, Bador et al. 2017),  
44 Australia (Cowan et al. 2014, Perkins et al. 2015) and North America (Argüeso et al. 2016).

45 Obviously, HW are prolonged periods of extreme temperatures but plethora of metrics exist  
46 depending on the issues at hand (see Perkins 2015 for a review). Because they involve distinct

47 processes, night-time and daytime HW are often distinguished and their identifications rely on the  
48 use of daily minimum and maximum near-surface (2-m) dry-bulb temperature,  $T_n$  and  $T_x$   
49 respectively (e.g., Robinson et al. 2001). In the perspective of having more representative ways of  
50 quantifying the human body sensation and stress to extreme heat, other variables, such as the wet-  
51 bulb temperature or the apparent temperature, may also be considered (e.g., Steadmann 1984,  
52 Willett and Sherwood 2012, Zhao et al. 2015, Raymond et al. 2021).

53 In this study, we focus on the subtropical Sahel, where little attention has been devoted to HW so  
54 far. In spring, prior to the monsoon season, the Sahel records particularly high temperatures during  
55 both night-time and daytime, with monthly mean  $T_n$  and  $T_x$  typically reaching 30°C and 40°C  
56 respectively (Guichard et al. 2015). In addition, the long-term temperature trend over the Sahel is  
57 particularly large during this hot season (Fontaine et al. 2013, Guichard et al. 2015): for the period  
58 1979-2011, it almost reached 2°C, significantly more than the global trend. As the HW frequency  
59 and intensity are mainly driven by the mean temperature trend (Argüeso et al. 2016, Déqué et al.  
60 2017, Barbier et al. 2018), Sahelian HW have become more frequent and more intense (Fontaine et  
61 al. 2013, Moron et al. 2016).

62 A few observed HW have been extensively studied in order to get insight on their driving  
63 mechanisms. In the mid-latitudes, the presence of a blocking high pressure system is often  
64 identified as a synoptic pattern, favouring HW, as it builds up warm air in the low layers through  
65 adiabatic heating by the associated large-scale subsidence (Black et al. 2004). It also favours clear  
66 skies and thereby positive surface net radiation anomalies, light winds and warm-air advection in its  
67 southern sector. This was the case during the Chicago 1995 (Meehl and Tebaldi 2004), European  
68 2003 (Ogi et al. 2005, Garcia-Herrera et al. 2010), Russian 2010 (Miralles et al. 2014) and several  
69 Chinese (Ding et al. 2010) heatwaves. Low-frequency modes of variability such as the North

70 Atlantic Oscillation and the Atlantic Multi-decadal Oscillation are also suggested to play a role  
71 (Della-Marta et al. 2007). Soil moisture-temperature feedback can also strengthen HW intensity  
72 (Quesada et al. 2012, Perkins et al. 2015), as it was highlighted for the Europe 2003 and Russia  
73 2010 HW (Fink et al. 2004, Stéfanon et al. 2012, Miralles et al. 2014). Reduced precipitation over  
74 Europe during the 2003 spring (Fischer et al. 2007a) associated with an early vegetation green-up  
75 enhancing evapotranspiration (Zaitchik et al. 2006) contributed to rapid soil moisture depletion  
76 (Fischer et al. 2007b, García-Herrera et al. 2010) and enhancement of surface sensible heat flux  
77 (Zaitchik et al. 2006). This positive feedback deepened and dried the boundary layer by  
78 accumulating heat, day after day (Miralles et al. 2012), reinforcing the pressure anomaly. The  
79 impact of HW may also be strongly modulated by the humidity of the air, leading sometimes to  
80 conditions at the limit of the human body tolerance (e.g., Raymond et al. 2021 and references  
81 therein). The processes driving the air humidity must therefore also be identified and understood. In  
82 the context of extreme humid heat events, Raymond et al. (2021) emphasises for instance the role of  
83 boundary-layer moisture fluxes, through e.g., sea breezes, combined with strong capping inversions  
84 inhibiting deep convection.

85 In the Sahel in spring, soils are climatologically already dry. Therefore, soil moisture-  
86 temperature feedback is very unlikely. In contrast, atmospheric water vapour is found to play an  
87 important role on Sahelian HW (Oueslati et al. 2017, Largeron et al. 2020, Bouniol et al. 2021),  
88 possibly leading to extreme humid heat events. Different changes in the atmospheric circulation can  
89 lead to a water vapour increase: intensification of the Saharan heat low (Knippertz and Fink 2008,  
90 Barbier 2017), northward shift of the intertropical front (Guichard et al. 2009, Couvreux et al. 2010,  
91 Largeron et al. 2020), occurrence of tropical plumes (Fröhlich et al. 2013), or presence of Rossby  
92 waves fostering south-westerly wind anomalies above the western regions of West Africa (Fontaine  
93 et al. 2013). Locally, water vapour, clouds, desert dusts, through their interactions with the radiative

94 and turbulent boundary layer processes also control the surface air temperatures (Guichard et al.  
95 2009, Bouniol et al. 2012, Gounou et al. 2012, Bain et al. 2010, Fontaine et al. 2013, Llargeron et al.  
96 2020). However, a detailed knowledge and understanding of their respective effects and interactions  
97 during Sahelian heatwave episodes is still lacking.

98 Numerical Global Climate Models (GCM) provide consistent projections of HW over Australia  
99 (Cowan et al. 2014) and Europe (Schoetter et al. 2015). In contrast, over Africa, and particularly  
100 over the Sahel, 2-m temperatures simulated by the GCM involved in the fifth phase of the Coupled  
101 Model Inter-comparison Project (CMIP5, Taylor et al. 2012a) show large biases (several degrees)  
102 and particularly during the dry season (Roehrig et al. 2013). These biases have been related to  
103 numerous processes, such as deep convection (Nikulin et al. 2012, Taylor et al. 2012b, Vautard et al.  
104 2013), microphysics (Vautard et al. 2013, Diallo et al. 2017), radiative cloud properties (Foster et al.  
105 2007), aerosol properties and their indirect effects on clouds (Knippertz and Todd 2012), or surface  
106 characteristics (Weisheimer et al. 2011, Diallo et al. 2017). Difficulties in representing the key HW  
107 processes with regional climate models and GCM with a resolution on the order of tens or hundreds  
108 of kilometres limit the conclusions that can be drawn from their numerical simulations of the Sahel  
109 climate.

110 Over West Africa, Marsham et al. (2013), Birch et al. (2014) and Vellinga et al. (2016) showed  
111 that the representation of the west african monsoon is improved with convection-permitting  
112 simulations. More recent studies confirm this finding (e.g., Stratton et al. 2018, Vizy and Cook  
113 2019). However, few studies focussed on HW with convection-permitting simulations (Zhang et al.  
114 2020, Ramamurthy and Bou-Zeid 2017) and, to our knowledge, none over the Sahel. Kendon et al.  
115 (2019) indicate that the projections performed with a convection-permitting model over some  
116 regions of Africa, like the Sahel, show an increased length of dry spells whereas the use of coarser

117 resolution models with parameterized deep convection indicates the reverse effect due to a less  
118 realistic triggering and propagation of convective systems. In the context of HW, it might be  
119 counter-intuitive to think that the deep convection parameterization can be detrimental. We shall see  
120 however that its ability to not trigger convection and associated rainfall is in fact critical for  
121 capturing the HW properties and processes at play.

122 Consequently, the present study analyses a severe HW that happened in April 2010 over the  
123 Sahel (Barbier et al. 2018, Largeron et al. 2020) and affected local economies and population health  
124 (Azongo et al. 2012, Diboulo et al. 2012). The main objectives are twofold: i) to test the ability of a  
125 limited-area model to represent the properties of the HW and ii) to analyse the potential added-value  
126 of using an explicit representation of deep convection in this HW context, during which  
127 precipitation is rare.

128 After having described the data and numerical set-up in Section 2, the April 2010 Sahelian  
129 heatwave episode is documented in Section 3, both at synoptic and local scales, based on a wide  
130 variety of observations. Section 4 then evaluates the simulations in view of the observations. In  
131 Section 5, a budget analysis is used to further understand the mechanisms at play during the  
132 heatwave, in particular to highlight the role of deep convection.

## 133 **2 DATA, METHOD AND SIMULATIONS**

### 134 **2.1 Reference datasets**

135 Several observational datasets, including in-situ and gridded measurements as well as satellite  
136 estimates, are used to analyse the Sahelian heatwave and evaluate the different simulations.

137 First, numerous high-frequency ground-based measurements implemented during the African  
138 Monsoon Multi-disciplinary Analysis (AMMA) project (Redelsperger et al. 2006) are examined.  
139 This notably includes observations collected by the AMMA-CATCH network at three sites (Galle et

140 al. 2018). Hereafter, we mainly use measurements from the Agoufou site in Mali (15°34'N, 1°48'W,  
141 local time: UTC+0h), located in the Central Sahel (Mougin et al. 2009). At this site, an automatic  
142 weather station provides near-surface air temperature, relative humidity, rainfall, wind speed and  
143 direction as well as Downwelling and Upwelling ShortWave (*SWD* and *SWU*) and LongWave  
144 (*LWD* and *LWU*) surface radiative fluxes with a 15-min time step (Guichard et al. 2009). An eddy-  
145 correlation station also provides estimates of the surface sensible (*H*) and latent (*LE*) heat fluxes at a  
146 30-min time step (Timouk et al. 2009). Note that we reached very similar conclusions with the data  
147 of the Niger Wankama AMMA-CATCH site located in the southern Sahel (Leauthaud et al. 2017).  
148 Vertically Integrated Water Vapour amount (*IWV*) is derived from GPS stations deployed at the  
149 Niamey (13.48°N, 2.17°E, local time: UTC+1h) and Ouagadougou (12.35°N, 1.52°W, local time:  
150 UTC+0h) sites, at an hourly frequency (Bock et al. 2008). The Aerosol Optical Depth (*AOD*)  
151 provided by several sunphotometers, from the AERONET network (Holben et al. 1998), is also  
152 used. Finally, radiosondes launched twice a day at Niamey (at 0000 and 1200 UTC) allow us to  
153 characterise the vertical structure of the lower atmosphere. We only had access to their low  
154 resolution version.

155 Various daily average and global regular 1°x1° gridded datasets are used here:

- 156 • the Berkeley Earth System Temperature dataset (BEST, Rohde et al. 2013) which  
157 incorporates a large ensemble of weather ground stations and provides the daily  
158 minimum and maximum 2-m temperatures,
- 159 • the Cloud and Earth's Radiant Energy System (CERES) SYN1deg dataset, which  
160 combines measurements made by several spatial instruments, in particular the Moderate  
161 Resolution Imaging Spectroradiometer (MODIS) to provide radiative fluxes, cloud  
162 cover, total AOD at 0.55  $\mu\text{m}$  and *IWV* (see Doelling et al. 2013 for details).

163 Data from the European Centre for Medium-range Weather Forecasts (ECMWF) Interim Re-  
164 Analysis (ERA-I, Dee et al. 2011) are also used to document the heatwave synoptic situation. The  
165 data is provided on a  $0.75^\circ \times 0.75^\circ$  horizontal grid every 6 hours.

166 Finally, the Tropical Rainfall Measuring Mission 3B42 product (TRMM-3B42, Huffman et al.  
167 2007) provides 3-hourly precipitation estimates at a  $0.25^\circ$  horizontal resolution.

## 168 **2.2 Radiative fluxes and surface energy balance**

169 The radiative fluxes at the surface are linked to the surface energy balance:

$$R_{net} = SWN + LWN = H + LE + G$$

170 where  $SWN = SWD - SWU$

171 and  $LWN = LWD - LWU$

172  $G$  is the ground heat flux.  $SWD$ ,  $SWU$ ,  $LWD$  and  $LWU$  are defined as positive. The net radiative  
173 fluxes ( $R_{net}$ ,  $SWN$  and  $LWN$ ), as well as  $G$ , are counted positive downward. Surface turbulent fluxes  
174 are counted positive upward.

175 The contribution of clouds and aerosols to the surface energy balance are quantified by the Cloud  
176 Radiative Effect ( $CRE$ ) and the Aerosol Radiative Effect ( $ARE$ ), respectively, following  
177 Ramanathan et al. (1989):

$$CRE = F - F_{clear-sky}$$

$$ARE = F_{clear-sky} - F_{clean-sky}$$

178 where, for any radiative flux  $F$ ,  $F_{clear-sky}$  is the cloud free  $F$  and  $F_{clean-sky}$  the cloud and aerosol free  $F$ .

## 179 **2.3 Model set-up**

180 The simulations are performed with the atmospheric limited-area model MésoNH version 5.2  
181 (Lafore et al. 1998, Lac et al. 2018). It includes SURFEX version 7.4 for the representation of  
182 surface processes (Masson et al. 2013).

### 183 **2.3.1 Configuration**

184 The model domain is centred on Central Sahel 10°N-18°N/5°W-3°E (see square in Figure 1) and  
185 the model horizontal resolution is 5 km. Two simulations are run, one in which the deep convection  
186 scheme is switched off (EXPL), and one in which it is turned on (PARAM), following a similar  
187 approach to that of Marsham et al. (2013) and Birch et al. (2014).

188 A stretched vertical grid of 87 levels is used with a finer resolution near the surface (first level at  
189 2 m) increasing with the altitude and reaching 1300 m at the 20-km top of the domain. The  
190 refinement under 4 km (65 levels) is expected to lead to a better representation of boundary-layer  
191 processes and surface-atmosphere interactions. In particular a first level at 2 metres facilitates the  
192 comparison to observations. A 3-km deep damping layer is added at the top of the domain to limit  
193 the reflection of gravity waves.

194 In terms of numerics, a fourth-order centred scheme coupled to an explicit fourth-order centred  
195 Runge-Kutta time-splitting is used for the wind advection while the forward-in-time piecewise  
196 parabolic method scheme is applied to scalar variables (Lac et al. 2018).

197 The runs are initialised on April 10, 2010 at 0000 UTC using the ECMWF operational analysis  
198 (0.25° horizontal resolution). The soil water indexes of the three bucket layers are also initialised  
199 using the ECMWF operational analysis. Some sensitivity tests to the soil moisture initialisation are  
200 discussed in the Supporting Information. The wind components, potential temperature and water  
201 vapour mixing ratio are nudged at the domain lateral boundaries towards the 6-hourly ECMWF

202 operational analyses. The nudging timescale is chosen small enough (25 s) to well constrain the  
203 simulations.

204 The model is integrated over 10 days, until April 20, 2010 0000 UTC.

### 205 **2.3.2 Parameterizations**

206 The ISBA (Interactions between Soil, Biosphere and Atmosphere, Noilhan and Planton 1989)  
207 surface scheme included in the SURFEX platform computes the soil energy and water budgets, and  
208 provides surface fluxes to the atmosphere. The surface physiographic information (soil occupation,  
209 vegetal cover, topography) is provided by the ECOCLIMAP 2 data base (Masson et al. 2003).

210 The deep convection scheme is based on the work of Kain and Fritsch (1990) and Bechtold et al.  
211 (2001). It is a mass-flux scheme with a CAPE closure using a 1-hour time-scale. It is activated only  
212 in the PARAM run. For both EXPL and PARAM runs, boundary-layer convection is represented  
213 with an eddy-diffusivity mass-flux formulation (Pergaud et al. 2009). The turbulent scheme of  
214 Cuxart et al. (2000) is used in its 1-D version and is based on a prognostic equation of the subgrid  
215 turbulent kinetic energy (Redelsperger and Sommeria 1986) closed with the turbulence mixing  
216 length of Bougeault and Lacarrère (1989).

217 The microphysics one-moment scheme predicts the mixing ratio of five hydrometeors: cloud  
218 droplets, raindrops, pristine ice crystals, snow aggregates and graupels. It uses a Kessler scheme for  
219 warm processes (Caniaux et al. 1994, Pinty and Jabouille 1998). A subgrid cloud scheme is also  
220 activated, which relies on the subgrid distribution of the saturation deficit (Bougeault 1982,  
221 Chaboureau and Bechtold 2005).

222 The ECMWF version of the Rapid Radiation Transfer Model (RRTM) is used for long-wave  
223 radiation (Mlawer et al. 1997, Morcrette 2002). The short-wave radiation scheme is based on

224 Fouquart and Bonnel (1980). Those two schemes are called every 15 min in clear-sky columns and  
225 every 5 min in cloudy columns.

226 Aerosols are present over the area especially during spring. The simulations use the six-class  
227 dataset of Tegen et al. (1997) which provides monthly-mean Aerosol Optical Depth (AOD) for each  
228 class. The 2D maps of AOD are then converted into 3D AOD using given aerosol concentration  
229 vertical profiles, fixed for each class. April 2010 is characterised by a large positive AOD anomaly  
230 (Largeron et al. 2020 and Figure 3c) which is thus not captured by the setup. This probably induces  
231 systematic errors. A sensitivity test to the aerosol content is documented in the Supporting  
232 Information.

### 233 **3 THE APRIL 2010 HEATWAVE**

234 In this section, the HeatWave (HW) episode is described using first satellite and gridded products  
235 over the whole area, then in-situ observations at the local scale.

#### 236 **3.1 Large-scale circulation and heatwave sequence**

237 In April 2010, a HW occurred over a large part of West Africa. It is captured both by daily  
238 minimum (Tn) and maximum (Tx) 2-m temperatures. Following the HW detection method of  
239 Barbier et al. (2018), the HW impacted the Central Sahel and southern Sahara (Mali, Mauritania,  
240 Burkina Faso, Niger) mostly from April 10 to 25, 2010 (see Figure 3 of Largeron et al. 2020). In the  
241 present work, we focus on the April 10 to 19, 2010 period. On average over this 10-day period, Tn  
242 and Tx anomalies over the Central Sahel range between 1 to 3°C, with respect to the April 10-19,  
243 1980-2010 BEST climatology (Figure 1). This corresponds to Tn and Tx above 30°C and 44°C,  
244 respectively (not shown).

245 Figure 2 illustrates the HW sequence from April 10 to 19, 2010 as a function of latitude. Tn  
246 progressively increases over the period from about 27°C to 31°C. These values correspond to

247 anomalies greater than 4°C, especially in southern Mali (north of 14°N). North of 11°N, Tx remain  
248 high all over the 10-day period, above 40 to 43°C (anomalies of about 2 to 3°C). There are however  
249 some intermittences of a few degrees between April 14 and 17. On April 19, Tx reaches almost  
250 45°C between 14°N and 16°N. Combined with the high Tn, these conditions are particularly tough  
251 for local populations.

252 The HW episode is associated with a large positive Integrated Water Vapour (IWV) anomaly,  
253 especially over southern Mali and along a wide band covering the Sahara from the south-west to the  
254 north-east (Figure 3a). This moist band is presumably the footprint of a tropical plume (e.g.,  
255 Fröhlich et al. 2013), generated by a quasi-stationary low which remained blocked for most of the  
256 period to the west coast of Morocco (not shown). The tropical plume is also visible in the cloud  
257 field (Figure 3b), and even associated with rainfall over northern Mauritania, northern Mali and  
258 Algeria (not shown). Over southern Mali, the moist anomaly is positioned across the trade winds  
259 convergence zone, named the Inter-Tropical Discontinuity (ITD, defined here as the 8 g/kg 2-m  
260 water vapour mixing ratio isoline, grey line). To its north, the quasi-stationary low enhances the  
261 zonal advection of moist air from the Atlantic Ocean (see 10-m wind anomalies in Figures 1 and 3)  
262 and reduces the meridional advection of cool maritime Mediterranean air over the Sahara, thereby  
263 allowing a strengthening of the Saharan heat low, especially around April 12 (pressure anomalies,  
264 shading in Figure 4). The latter, in turn, increases the meridional and zonal (not shown) pressure  
265 gradient, which is favourable to the intensification of the southwesterly moist flow. As a result, the  
266 ITD shifts northward. This pattern will be referred to as a monsoon surge in the following. The ITD  
267 retreats on April 18 when the meridional pressure gradient decreases. On average over the ten days,  
268 the ITD is 2° further north than the climatology. The moisture anomaly over the Central Sahel thus  
269 results from both southwesterly advection by the enhanced monsoon flow and northwesterly

270 advection by the enhanced Atlantic inflow. A quantitative assessment of their respective  
271 contribution is, however, beyond the scope of the present work.

272 LARGERON et al. (2020) showed that these moist anomalies are critical to understanding the  
273 increased  $T_n$ , through their “greenhouse” effect on the surface Downwelling LongWave radiative  
274 flux (LWD, see also Figure 3g) but did not conclude on the factors increasing  $T_x$ . Clouds weakly  
275 impact the surface LWD (Figure 3h), while significantly reducing the Downwelling ShortWave  
276 radiation at the surface (SWD), especially over the north-west part of the region, impacted by the  
277 tropical plume (Figure 3e). The aerosol optical depth, which is anomalously high over the region of  
278 interest (Mali, Niger, Burkina Faso, Figure 3c), significantly reduces the downwelling SW and LW  
279 fluxes at the surface (Figures 3f,i,l). The sum of the cloud and aerosol radiative effect anomalies is  
280 negative (Figure 3j), and thus should cool surface temperatures. As a result, the positive  $T_x$   
281 anomalies do not link to the shortwave surface cloud radiative budget, in contrast to the classical  
282 scheme of European HW. As for positive night-time temperature anomalies, the positive daytime  
283 temperature anomalies are likely the footprint of the higher water loading in the atmosphere (Figure  
284 3a) and the associated higher LWD (Figure 3g). Thus the greenhouse effect impacts both night-time  
285 and daytime temperatures. but other processes may be at play, such as an increased entrainment of  
286 free tropospheric air within the boundary layer, or an increased warm air horizontal advection from  
287 the Sahara.

288 Figure 4 also shows the time-evolution of the latitudinal distribution of the TRMM-3B42 rainfall  
289 during the 10-day period. The rain falls south of the ITD and follows its northward shift. The  
290 precipitation estimates show several precipitating events from April 12 to 15, south of  $13^\circ\text{N}$  (with a  
291 3-hourly rainfall maximum of 41 mm/day). On April 16, 2010 another event, quite north for the  
292 season, extends from  $13^\circ\text{N}$  to  $17.5^\circ\text{N}$ . The amount remains weak though (3-hourly rainfall

293 maximum of 9 mm/day) and is questionable as in such an arid region, the occurrence of rainfall  
294 evaporation may significantly bias the TRMM 3B42 estimates of surface precipitation (Dinku et al.  
295 2011).

### 296 **3.2 Heatwave sequence at the local scale**

297 In this section, we use in-situ observations to further document the event at the local scale. These  
298 observations are independent of the datasets used in the previous section, and, as described below,  
299 they provide results consistent with the previous section findings, therefore emphasising their  
300 robustness.

301 Figure 5 presents the time evolution of the IWV observed by GPS (thick black lines) at the  
302 Niamey and Ouagadougou sites (see locations in Figure 1). At both stations, the northward shift of  
303 the ITD yields a moisture increase at a rate of about 10 mm/day from April 10 to 14, 2010. IWV  
304 thus quadruples, reaching high values for the season (consistent with the CERES IWV anomalies  
305 indicated in Figure 3a). Note that Niamey GPS data are lacking on April 10 to 12 but the  
306 AERONET data at Banizoumbou, close to Niamey indicate a similar rise. The monsoon surge  
307 withdrawal begins on April 17 but is clearer at Niamey than at Ouagadougou (located further  
308 South). No rainfall is observed at Niamey although precipitation is observed at Ouagadougou  
309 during the April 14 and 16 nights (thin black line in Figure 5b).

310 The Agoufou site (Mali, the northernmost point in Figure 1) provides measurements of most of  
311 the surface energy budget components, together with the surface meteorology (black lines in Figure  
312 6). It thus enables a detailed investigation of the surface processes at play during the heatwave, at  
313 least at the local scale.

314 From April 10 to 13, the 2-m temperature exhibits a strong diurnal amplitude (around 20°C,  
315 Figure 6a). Tx reaches 42.5 to 44°C, while the surface air layer remains very dry, the mixing ratio

316 being below 3 g/kg (Figure 6c). Weak south-easterly winds prevail (Figure 6b,d) until the monsoon  
317 surge reaches Agoufou on April 13.

318 Within a couple of hours on April 13, the monsoon surge arrival in Agoufou induces a wind  
319 reversal from easterly to westerly (Figure 6b). The water vapour mixing ratio dramatically increases  
320 by about 10 g/kg (Figure 6c). This induces a substantial jump in the LWD by more than 50 W/m<sup>2</sup>  
321 (dashed line in Figure 6f) and a decrease of the net energy loss by longwave radiation, mostly due to  
322 its clean-sky contribution (not shown). This mitigates the night cooling (Guichard et al. 2009,  
323 Largeron et al. 2020) and thereby reduces the diurnal temperature range. T<sub>n</sub> increases by 6°C from  
324 April 12 to 13, then again by 5°C from April 13 to 14 and finally reaches 34°C on April 16. T<sub>x</sub> is  
325 less affected by the monsoon surge, and continues to vary between 42°C and 44°C. A computation  
326 of the wet-bulb temperature combining temperature and humidity following Zhao et al. (2015)  
327 indicates a significant to extreme heat stress for the population during the monsoon surge, the  
328 hardest time being April 15, 1400 UTC with 44°C as dry-bulb temperature and 8.5 g/kg of water  
329 vapour mixing ratio, that is to say 34°C as wet-bulb temperature.

330 Finally, on April 17, the monsoon surge retreats southward, inducing a new wind reversal from  
331 south-westerlies to north-easterlies (Figure 6b,d) and a slow decrease of near-surface water vapour  
332 (Figure 6c) until April 20. A sharp decrease in the surface SWD radiative (~150 W/m<sup>2</sup>, thin lines in  
333 Figure 6f), net surface radiative (~70 W/m<sup>2</sup>, thick line in Figure 6f) and surface sensible heat fluxes  
334 (~40 W/m<sup>2</sup>, thin line in Figure 6e) is concomitantly observed. Based on CERES data, this can be  
335 attributed to cloud cover and AOD increases (CRE and ARE induce a surface net SW decrease of 30  
336 and 40 W/m<sup>2</sup>, respectively, not shown). In contrast, the surface LWD radiative flux remains strong  
337 (dashed line in Figure 6f). Consistently, T<sub>x</sub> slightly decreases from April 15 (except on April 19),  
338 while T<sub>n</sub> remains high (except on April 17).

## 339 **4 EVALUATION OF THE SIMULATIONS**

340 This section evaluates the ability of the two simulations, with the parameterization of the deep  
341 convection turned off (EXPL) and on (PARAM), to reproduce the scenario previously described.  
342 We first focus on the HeatWave (HW) sequence, at the scale of the whole simulation domain, then  
343 at the scale of several sites. The last sub-section uses the radiosoundings launched at Niamey, Niger  
344 to assess the representation of the entire boundary layer.

### 345 **4.1 The simulated heatwave sequence at large scale**

346 Figure 7 presents the daily 2-m minimum and maximum temperature (Tn and Tx respectively)  
347 Hovmöller time-latitude diagrams for the simulations EXPL and PARAM, similarly to Figure 2.  
348 The simulation EXPL captures the strong increase of Tn following the monsoon surge from April 12  
349 to 17, 2010. In particular, the very high Tn (often above 29.5°C) between 14°N and 17°N, around  
350 the Inter-Tropical Discontinuity (ITD) are well captured. PARAM agrees less, with Tn rarely  
351 reaching 29°C.

352 Both Tn and Tx display negative biases compared to BEST, with quite a bit of variability in both  
353 space and time though. During the first four days (April 10-13), the simulation cold bias, which is  
354 similar in both simulations, is stronger north of the ITD. It reaches there -4° for Tn and -3° for Tx.  
355 This bias is slightly larger than that of the ECMWF analyses used to initialise the model and to  
356 provide lateral boundary conditions (see Appendix A1). The initial cold bias is likely related to an  
357 overestimate of the soil moisture as provided by the ECMWF analysis. Given the dryness of the air  
358 north of the ITD, the excess of soil moisture rapidly evaporates and cools the low levels of the  
359 model for a few days. However, drying the ground at the start of the simulation leads to even colder  
360 minima north of the ITD (see sensitivity test in the Supporting Information). The negative biases are  
361 weak south of the ITD, except for PARAM.

362 Then, from April 14 to 19, Tn and Tx remain underestimated by a few degrees south of the ITD,  
363 in the core of the monsoon surge, where precipitation occurs (Tn is slightly worse than in the  
364 ECMWF analysis). Only the EXPL simulation captures similar or slightly higher temperatures than  
365 those observed (April 14 for Tn and Tx, 16 for Tx, April 18 for Tn).

366 Tn biases are expected to be strongly linked to surface Downwelling LongWave (LWD) biases.  
367 Indeed, a comparison with the CERES data indicates a strong lack of cloud cover and aerosol  
368 content in both simulations, thereby leading to underestimated LWD over the whole domain (not  
369 shown). It could partly explain the colder temperature compared to observations.

370 Figure 8 presents the near-surface temperature (T2m) differences between the PARAM and  
371 EXPL simulations. PARAM is always colder than EXPL (up to almost  $-9^{\circ}\text{C}$ , shading) south of the  
372 ITD (colored isolines) except when precipitation occurs in EXPL (April 14-17 nights, see next  
373 paragraph). Interestingly, this bias presents a strong diurnal cycle and, as shown later, is associated  
374 to spurious precipitation in PARAM. The ITD diurnal cycle is 4 hours ahead in PARAM and  
375 reaches latitudes slightly further north than EXPL does ( $0.8^{\circ}$  on average). North of the ITD, where  
376 no precipitation occurs (Figures 4 and 9), the two simulations agree well.

377 EXPL exhibits a similar temporal variability and latitudinal extent of precipitation compared to  
378 the TRMM 3B42 reference (Figure 9), although not always with the right timing or intensity. EXPL  
379 captures the latitudinal precipitation distribution rather well over the 10-day period, with a slight  
380 overestimate south of  $12.5^{\circ}\text{N}$  and a slight underestimate north of  $13.5^{\circ}\text{N}$  (Figure 9c). North of  
381  $13.5^{\circ}\text{N}$ , the two observed events of April 16 and the April 17-18 night occur in EXPL but the  
382 rainfall evaporates before reaching the ground (not shown). This explains the negative bias but, as  
383 already mentioned, the occurrence of TRMM-3B42 surface rainfall may also be questionable.  
384 Finally, on average, EXPL triggers rain approximately at the right time in the day but the rainfall

385 peak is slightly too early, between 1700 and 2100 UTC against near midnight for TRMM 3B42  
386 estimates (Figure 9d).

387 In contrast, the PARAM simulation triggers deep convection every day around noon, which then  
388 lasts until midnight (Figure 9b). This systematic triggering of deep convection is consistent with the  
389 diurnal cycle of the temperature difference between PARAM and EXPL (Figure 8). PARAM rainfall  
390 is significant up to the ITD. On average over the 10 days, precipitation is severely overestimated at  
391 all latitudes (Figure 9c). The composite diurnal maximum of rainfall also occurs too early, between  
392 1400 and 1500 UTC (Figure 9d). The numerical model with parameterized convection of Marsham  
393 et al. (2013) exhibits similar behaviour during the Sahel wet season (see their Figure 1a). The  
394 spurious precipitation have a detrimental impact on surface temperatures particularly on the  
395 temperature minima (cooling associated to the evaporation of precipitation) as seen in Figure 8.

396 Finally, note that the ITD has strong and regular diurnal fluctuations in both simulations as  
397 usually observed in this area (Pospichal et al. 2010). During a day, it can move northward up to  $2^\circ$   
398 (Figure 9a,b), coupled with the occurrence of a strong low-level jet at the end of the night (see also  
399 next sections). In contrast, the daily ITD latitudinal displacement is weaker in ECMWF analyses  
400 (up to  $0.5^\circ$  during a few days, dashed in Figure 9b) suggesting that high resolution is needed to  
401 capture those diurnal fluctuations.

## 402 **4.2 The simulated heatwave sequence at the local scale**

403 As shown in Figure 5, the IWV increase is correctly reproduced by both simulations at the two  
404 GPS stations located in Niamey and Ouagadougou. The IWV at Niamey is however systematically  
405 overestimated in the PARAM simulation (Figure 5a). At Ouagadougou, where the atmosphere is  
406 closer to the saturation, it is better captured (Figure 5b). The EXPL IWV is generally closer to  
407 observations, with two exceptions. The observed event on April 14 0000 UTC is only captured by

408 EXPL at Ouagadougou neighbouring grid points (not shown) and the strong IWV peak on April 15  
409 0000 UTC associated with the occurrence of a convective event (thin green lines in Figure 5b) is not  
410 observed. Except for this last event, PARAM produces more rainfall than EXPL, almost every  
411 afternoon from April 12 at Ouagadougou and from April 14 at Niamey. In contrast, no rain is  
412 observed at Niamey during the period, and only a few events occur at Ouagadougou.

413 On April 17 1200 UTC, the GPS IWV starts to decrease following the monsoon surge retreat.  
414 This decrease is delayed in the simulations, inducing moist biases during the last three days of the  
415 period. This departure is consistent with the maintenance of the south-westerly wind in the  
416 simulations (not shown). This delay is also present in the lateral nudging model (see the ECMWF  
417 ITD indicated by the black dashed line in Figure 9).

418 The Agoufou measurements (Section 3.2) are now used to evaluate the locally-simulated surface  
419 energy budget (Figure 6). From April 10 to 13, the simulations are too cold by 2 to 3°C, both in  
420 terms of  $T_x$  and  $T_n$ . These biases are consistent with those observed at larger scale. They are also  
421 consistent with an underestimated net radiative flux at the surface ( $R_{net}$ ) by 20 W/m<sup>2</sup> (thick line of  
422 Figure 6f). The probably overestimated initial soil moisture leads to an overestimated latent heat  
423 flux during the first three days, which then largely reduces, at least in EXPL (thick lines in Figure  
424 6e). This extra evaporative cooling then likely contributes to the cold biases of both simulations (see  
425 also the Supporting Information).

426 On April 13, the moisture increase, associated with the increased  $T_n$ , and the change in wind  
427 direction from southeasterlies to southwesterlies, are qualitatively well reproduced by the EXPL and  
428 PARAM simulations. However, the PARAM spurious precipitation events (Figure 6c) increase the  
429 water recycling through surface evaporation (e.g., see spike of day-mean LE up to 44 W/m<sup>2</sup> in

430 Figure 6e) and likely yield the strong PARAM wet biases (more than 4 g/kg, Figure 6c) and the cold  
431 biases in Tx (1°C) and Tn (more than 5°C).

432 The monsoon surge retreat and wind direction shift are significantly delayed in the two  
433 simulations (Figure 6c,d), which behave similarly to the ECMWF analyses (black stars in Figure 6a  
434 to d). The sharp decrease in the downwelling shortwave, net surface radiative and sensible heat  
435 fluxes associated with a large cloud cover is also not simulated during this period. Indeed, the  
436 CERES cloud fraction increases from 5% to 35% from April 12 to 16 and stays at this level  
437 afterwards while it evolves only from 2 to 4% in the simulations (not shown). This lack of cloud  
438 cover mitigates the Tx cold bias mainly during the second period of the simulation. The simulated  
439 LWD are then underestimated (dashed in Figure 6f) but as Rnet is larger in the simulations  
440 compared to the observations, radiation can not explain alone the strongest negative Tn biases of the  
441 second half of the period. Other mechanisms are involved, possibly the cooling by south-westerly  
442 winds which reverse only the last day in the simulations.

443 To summarise, the monsoon surge from April 13 to 16 is well simulated and better than its  
444 withdrawal. A negative air near-surface temperatures bias persists nevertheless relatively insensitive  
445 to several aspects of the representation of physical processes and to the initial and boundary  
446 conditions (see sensitivity tests in the Supporting Information). The spurious precipitation by the  
447 deep convection scheme enhances this bias.

448 In Figure 10, we compare the composite simulated atmospheric profiles of some meteorological  
449 variables at the closest grid point (colour lines) with radiosoundings (stars) launched at Niamey in  
450 Niger (middle point in Figure 1).

451 The radiosoundings at Niamey indicate a significant diurnal cycle of the boundary layer with a  
452 strong nocturnal temperature inversion (Figure 10c) and a 1.5-km-deep well-mixed convective

453 boundary layer not yet completely developed at 1200 UTC (Figure 10b,d) typical of this pre-  
454 monsoon period (Lothon et al. 2008, Guichard et al. 2009, Gounou et al. 2012). The “nocturnal”  
455 Low Level Jet (LLJ) is also typical of the end of the dry season in the Sahel, as deep convection  
456 does not disturb its nighttime development (Parker et al. 2005a, Lothon et al. 2008). At 0000 UTC,  
457 the observed LLJ is not yet fully developed (only its southerly component is visible, stars in Figure  
458 10e,g). At 1200 UTC, it has already begun to retreat and only its remaining bell-shape maximum  
459 near 500 m above ground level is noticeable (Figure 10f,h). The LLJ advects water vapour from the  
460 southern regions, which is vertically redistributed during daytime by the Boundary Layer (BL)  
461 turbulent mixing. The BL is topped by a weak African easterly jet around 4 km height, which results  
462 from the thermal wind balance (Parker et al. 2005b). The wind shear between the LLJ and the  
463 African easterly jet amplifies the entrainment at the top of the BL (Canut et al. 2010, Gounou et al.  
464 2012).

465 The main temperature and water vapour biases found at the surface actually impact the entire BL  
466 depth (Figures 10a-d). The PARAM humidity bias in the BL is consistent with the IWV bias seen in  
467 Figure 5a (§4.1). Indeed, most of the contribution to IWV results from the low levels (Couvreur et  
468 al. 2010). At 0000 UTC, in the stable BL, the water vapour bias reaches 5 g/kg for PARAM against  
469 less than 2 g/kg for EXPL (Figure 10a), consistently with the LLJ being too strong in PARAM  
470 (Figure 10e,g). The remarkable nocturnal inversion at the surface is more pronounced in PARAM  
471 than in EXPL. At 1200 UTC, the negative temperature and positive water vapour biases persist in  
472 PARAM (Figure 10b,d). EXPL better captures the wind component profiles (Figure 10f,h).

473 In the entire BL in Niamey and at the surface in Agoufou, PARAM spurious precipitation leads  
474 to strong negative temperature biases and strong positive water vapour content and wind biases. In

475 the next section, we investigate the underlying processes and mechanisms behind the differences of  
476 these two simulations through the use of thermodynamic and wind budgets.

## 477 **5 PHYSICAL MECHANISMS**

478 The two simulations EXPL and PARAM strongly differ in terms of rainfall amount and extent,  
479 low-layer temperature and humidity. Here, we use the budget of the different prognostic variables to  
480 understand these differences. They develop early, during the first two days, as shown in Section 5.1.  
481 Section 5.2 then shows how these differences, which emanate from the rainier lower latitudes  
482 extend over the highest latitudes with the monsoon surge. All budgets are averaged over the 500  
483 lowest metres and the longitudes [4.5°W-2.5°E].

### 484 **5.1 Early stage of the simulations**

485 Figure 11 focuses on the first two days with the evolution of the different terms of the hourly  
486 potential temperature ( $\theta$ ) budget (see Equation 2 in Appendix A2 for details) for the EXPL  
487 simulation (Figure 11a,d) and the difference between PARAM and EXPL (Figure 11b,e). The solid  
488 coloured lines represent the individual source terms and the brown dashed lines, their sum. The  
489 budget term differences between PARAM and EXPL are further integrated in time from the  
490 beginning of the simulation in Figure 11c,f (see equation 5), to assess the contribution of individual  
491 processes to the temperature biases exhibited in the previous sections (see equation 6). The sum of  
492 these contributions (black dash line) is equal to the temperature bias. Two latitude bands are  
493 analysed:

494 Over the [13°N-15°N] latitude band, none of the simulations produce precipitation (see also  
495 Figure 9). Consistently, the deep convection scheme does not trigger ( $\overline{DeepCV_\theta}$  is zero) and the  
496 microphysical scheme does not cool the BL ( $\overline{Micro\Phi_\theta}$  is zero). The EXPL total potential temperature  
497 tendency (Figure 11a) is positive during daytime as a result of heating by solar radiation and

498 turbulence. It is negative during nighttime and early morning due to longwave radiation and  
499 advective cooling (associated with the Low Level Jet LLJ). Note that for this thin 500m-high layer,  
500 the turbulence source (Turb) is driven by the surface temperature evolution. The subgrid turbulence  
501 scheme contribution encompasses the shallow convection scheme contribution. The subgrid  
502 turbulence warms the layer during daytime due to the large positive sensible heat flux, while it  
503 slightly cools the layer at night following the weakly negative sensible heat flux. The shallow  
504 convection scheme only acts during daytime: it first warms the layer between 0800 and 1100 UTC  
505 and then cools it until sunset due to the enhanced mixing between the lowest 500 m of the BL and  
506 the upper part of the BL which depth largely overpasses 500 m in the afternoon (not shown).

507 PARAM displays a budget evolution similar to EXPL, except that the advective cooling begins  
508 earlier (Figure 11b) and drives a stronger overnight cooling, which leads to a colder potential  
509 temperature of the layer (Figure 11c, up to 0.2 K and 0.9 K during the first and second nights,  
510 respectively). The enhanced overnight advective cooling is compensated during early daytime by an  
511 enhanced turbulence warming, so that the daytime temperature difference between the two  
512 simulations is negligible.

513 The stronger advective cooling  $\overline{Adv_\theta}$  in PARAM during nighttime is thus key to understanding  
514 the temperature differences between PARAM and EXPL at latitudes without precipitation since the  
515 beginning of the simulation. Figure 12 indicates that, on average over the first two days, PARAM is  
516 significantly colder than EXPL between the surface and about 2.5 km above the ground, associated  
517 with higher pressure, at PARAM rainy latitudes (up to 12.5°N, cf. Figure 9) but also beyond  
518 between 12.5°N and 15°N. Indeed, the stronger PARAM pressure gradient drives stronger  
519 meridional winds (purple contours in Figure 12), especially when the LLJ forms during nighttime,  
520 contributing to the increased advective cooling and moistening (green contours in Figure 12) further

521 north. Above 3.5 km, and south of 15°N, PARAM is warmer, consistently with the occurrence of  
522 latent heat release by deep convection. EXPL simulates weak convection during the first two days,  
523 which is confined to the southernmost latitudes.

524 Over the [11°N-13°N] latitude band, the potential temperature of EXPL evolves similarly to its  
525 evolution over the [13°N-15°N] latitude band except that the advective cooling begins in the early  
526 evening (Figure 11d). The diurnal cycle of PARAM  $\overline{Adv_\theta}$  is also shifted earlier and drives cooling of  
527 the layer only during the early night (Figure 11e). Above all, the colder temperature in PARAM is  
528 mostly explained by the activation of the deep convection scheme (see also Figure 9) and associated  
529 precipitation evaporation which cools the layer during the first hours of the simulation (possibly  
530 model spin-up), then from April 10, 1200 to 2000 UTC and finally from April 11, 1600 UTC (blue  
531 line in Figure 11e). Even though it is weak, the radiation term (yellow lines) always keeps the same  
532 sign and therefore also contributes to the difference between both simulations (see the integrated  
533 term in Figure 11f). The enhanced longwave radiative cooling in PARAM is mainly due to a lower  
534 surface upward longwave (not shown but differences between simulations in upward longwave term  
535 are larger at the surface than at the top of the 500 m-layer) which contributes to the decrease in the  
536 low-level temperature. To summarise, at those latitudes, while it does not rain in EXPL, the  
537 negative bias (PARAM being colder than EXPL, dashed black line in Figure 11f) increases mostly  
538 because of the  $\overline{DeepCV_\theta}$  and radiative terms.

539 Figure 13 further analyses the meridional wind ( $v$ ) budget at latitudes intermediate between the  
540 two previous latitude bands. The meridional component results from a complex equilibrium  
541 between different terms (Equation 3 in Appendix A2, Figure 13a). The pressure term is the primary  
542 force that drives wind acceleration. It is the major driver of the monsoon surge, during this pre-  
543 monsoon season (consistently with Couvreux et al. 2010). Its diurnal cycle is mainly linked to the

544 daytime heating. The Coriolis force accelerates  $v$  as long as the easterly winds are established. The  
 545 advection term also accelerates the meridional component. Only the daytime subgrid turbulence (  
 546  $\overline{Turb_{sb}}$ ) slows it. Indeed, the thermals accelerate the wind over the 500 lowest meters (the wind is  
 547 maximum around 500 m above the ground so that  $\overline{Turb_{th}}$  is positive from the surface to 500 m and  
 548 negative above, not shown) but  $\overline{Turb_{sb}}$  dominates the turbulent term which finally decelerates  $v$ . The  
 549 sum of those contributions leads mainly to a southerly meridional wind (dash black line).

550 The main difference between PARAM and EXPL in the meridional wind budget is thus the  
 551 pressure contribution (Figure 13b). Consistently with the PARAM increased pressure over [11°N-  
 552 13°N], compared to EXPL (black contours in Figure 12), the pressure term between 12°N and 14°N  
 553 accelerates the PARAM southern component more. It is nearly balanced by the Coriolis term. The  
 554 PARAM enhanced meridional advection of  $v$  and the turbulence terms slowing down more also  
 555 contribute to the difference to a lesser extent. This leads to an acceleration of the southerly wind in  
 556 PARAM during the night, thereby participating in the colder advection over the [13°N-15°N]  
 557 latitudes (Figure 11c). The first night, both simulations have similar LLJ. The second night PARAM  
 558 nocturnal LLJ starts earlier and is stronger, consistent with the forward and stronger  $\overline{Adv_{\theta}}$ .

559 As a consequence, right from the first days, the triggering of the convection scheme over [11°N-  
 560 13°N] in PARAM, cools the low layers, increases the pressure in the low levels and enhances the  
 561 LLJ between 12°N and 14°N. Then, the low layers of the dry northern latitudes [13°N-15°N] cool  
 562 by advection.

563 Figure 14 (first column) shows the budget of both simulations as a function of latitude. Budgets  
 564 are averaged over the first two days and still between the surface and 500 m. As already discussed,  
 565 PARAM is colder than EXPL south of the ITD up to 1K (black lines). The bias originates from  
 566 rainy PARAM latitudes (11°N to 13°N) with a cooling  $\overline{Micro\Phi_{\theta}+DeepCV_{\theta}}$  term (Figure 14a),

567 associated with a stronger surface latent heat flux (cyan line in Figure 14c). The radiation (yellow  
568 lines) also plays an important role in the  $\theta$  tendency difference due to PARAM colder surface,  
569 associated with a smaller surface sensible heat flux (orange lines in Figure 14a) and subgrid  
570 turbulence. The advection propagates this bias northward up to 15.5°N, due to a slightly stronger  
571 PARAM south wind accelerated by a stronger pressure force and temperature meridional gradient  
572 (Figure 14e).

573 This stronger PARAM nocturnal wind also provides more water vapour northward at dry  
574 latitudes (advection red line in Figure 14c, see also Equation 4 in Appendix A2). For both  
575 simulations,  $\overline{Adv_{rv}}$  almost balances  $\overline{Turb_{rv}}$ .  $\overline{Adv_{rv}}$  is the main positive contributor but rainfall  
576 evaporation (included in  $\overline{Micro\Phi_{rv}}$ ) and surface water evaporation (included in  $\overline{Turb_{sb_{rv}}}$ ) also  
577 supply low layers in water vapour to a lesser extent. Conversely,  $\overline{Turb_{th_{rv}}}$ , by mixing the wettest  
578 lowest 500 m with the drier layers above, depletes the layer of its water vapour. The combination of  
579 these two turbulent effects results in a drying by the total turbulence  $\overline{Turb_{rv}}$ . Finally, the total water  
580 vapour tendency is positive but that of PARAM is stronger.

581 As EXPL does not trigger any precipitation the first two days, its microphysical term is zero. For  
582 PARAM, the  $\overline{Micro\Phi_{rv}+DeepCV_{rv}}$  is close to zero because the microphysical part (source) cancels  
583 the convection part (sink). North of PARAM rainy latitudes (13°N on average over the first two  
584 days), the water vapour source is only the advection, stronger for PARAM than for EXPL, leading  
585 to at least 1.5 g/kg more.

## 586 **5.2 Following evolution**

587 For the next two days (April 12 0000 UTC to 14 0000 UTC), the main balances of the first two  
588 days hold but shifted 1.5° northward due to the synoptic monsoon surge imposed by the nudging at  
589 the boundaries (Figure 14b,d,f). Indeed the southerly component of the wind increases (up to +2 m/s

590 at 13°N on average over two days, black lines in Figure 14e,f) mainly driven by the intensification  
 591 of the meridional pressure force (blue lines). Due to the Coriolis force and to a lesser extent to the  
 592 advection (not shown), the westerly component of the wind also increases boosting the low level  
 593 water vapour for all latitudes (up to +4 g/kg at 13°N, black lines in Figure 14c,d). As already  
 594 explained, the radiation budget, by the intermediate of the long wave greenhouse effect, increases  
 595 the temperature tendency at latitudes with no rainfall for each simulation (compare the yellow lines  
 596 in Figure 14a,b). So that despite the advective and microphysical cooling, the heatwave strengthens  
 597 up to 1°C (black lines, stronger effect for EXPL).

598 The temperature and water vapour differences between PARAM and EXPL increase and spread  
 599 more northward than the synoptic monsoon shift because of the growing difference in rainfall  
 600 amount (which is noticeable on the greater difference on the surface sensible heat flux  $H$  in Figure  
 601 14b and latent heat flux  $LE$  in Figure 14d but also in Figure 9). Yet, precipitation begins for EXPL  
 602 but stays south of 13°N while south of 15°N for PARAM. The EXPL microphysical terms are no  
 603 longer zero. EXPL  $\overline{Adv}_\theta$  becomes strong and stronger than PARAM  $\overline{Adv}_\theta$ . South of 13°N,  $\overline{Adv}_\theta$  and  
 604  $\overline{Micro}\Phi_\theta$  sinks, not completely balanced by the  $\overline{Turb}_\theta$  and  $\overline{Rad}$  sources, lead to an EXPL potential  
 605 temperature total tendency a little more negative than the PARAM one.

606 Figure 15 presents the differences between the integrated terms of the potential temperature  
 607 budgets of EXPL and PARAM for the entire period. The scenario described in the previous section  
 608 (notably Figure 11) remains valid over most of the 10-day period, providing that it follows the  
 609 northward migration of the monsoon surge. In the [13°N-15°N] band, the precipitation has a strong  
 610 impact in PARAM with a larger sink in potential temperature due to the  $\overline{Micro}\Phi_\theta + \overline{DeepCV}_\theta$  and the  
 611 radiative terms from April 13 onward. The cooling by advection is larger in EXPL starting on April  
 612 14 as observed south of 13°N for April 12-13 (not shown). Further north, in the [15°N-17°N] band,

613 the cooling by advection is larger in PARAM except in the EXPL rainy nights of April 14-15 and  
614 April 16-17 (Figure 15a).

## 615 **6 CONCLUSION**

616 An observed Sahelian heatwave episode has been simulated with a high-resolution limited-area  
617 model focusing on the area (10°N-18°N, 5°W-3°E) over the April 10-20, 2010 period. This case  
618 study contrasts with the European heatwave cases investigated by Miralles et al. (2014) which  
619 strongly involved the soil desiccation whereas for the Sahelian zone, soils are already very dry at  
620 the end of the dry season (Guichard et al. 2009, Gruhier et al. 2010, Llargeron et al. 2020). The  
621 studied period begins with high low-level temperature maxima (above 42°C, +1.5° above  
622 climatology) north of the Inter Tropical Discontinuity (ITD) over the Sahel, with a low level  
623 easterly wind. From the first days, a monsoon surge, coming from the south-west, extends  
624 progressively to the north of the area. This cool and moist monsoon incursion is linked to the  
625 northern shift of the ITD which is located further North in April 2010 than in the climatology. We  
626 emphasise the major role of the integrated water vapour, reaching twice its climatological value on  
627 average over the period, and which induces a significant warming associated with its longwave  
628 greenhouse effect. The latter outweighs the cloud and aerosol cooling effects. This radiative  
629 warming impacts low-level minimum temperatures (above 30°C, +3° above the climatology) and is  
630 responsible for an intense humid stress on local populations, even though it is slightly mitigated by  
631 the monsoon surge cool advection. Weak and intermittent precipitation, as well as the occurrence of  
632 clouds, also slightly temper temperature rises, mostly south of 14°N.

633 The numerical simulation of this heatwave episode uses a 5 km horizontal resolution model with  
634 the deep convection parameterization either turned off (EXPL) or on (PARAM). Both simulations  
635 present a negative temperature bias compared to BEST observations partly related to the one of the

636 ECMWF analysis used for the initial and boundary conditions of the simulations. However, there is  
637 no temperature cold drift during the simulations. Such a cold bias is found in numerous models  
638 during spring (e.g., Barbier 2017) and likely involves errors in the physical processes such as those  
639 related to clouds and aerosols as well as issues with the parameterizations of the land surface and  
640 turbulence (e.g., Diallo et al. 2017). The evaporative cooling of the soil moisture excess, issued also  
641 from the ECMWF-based initialisation, participates in this cold bias, mostly north of the ITD (the  
642 driest latitudes) and daytime (then impacting more the daily maximum temperatures). Yet, EXPL  
643 simulates qualitatively well the monsoon surge with the associated temperature evolution and the  
644 spatial variation of the precipitation despite a premature diurnal cycle compared to observations.  
645 PARAM exhibits an even earlier diurnal cycle than EXPL with an excess of precipitation which  
646 enhances the cold bias when it evaporates.

647 The analysis of thermodynamic and dynamic budgets in the low atmospheric levels emphasises a  
648 balance between the daytime heating/drying by turbulence and the night-time cooling/moistening by  
649 advection, largely operated by the nocturnal Low Level Jet (LLJ). The dynamic budget further  
650 highlights the pressure gradient as the major driver of the monsoon pulsation, during this pre-  
651 monsoon season (consistently with Couvreux et al. 2010) and also shown by Birch et al. (2014)  
652 during the core monsoon season. Figures 16a,b synthesise the behaviour of the EXPL simulations.  
653 The lower surface pressures are located near the ITD in the middle of the domain and are more  
654 pronounced in the afternoon and early night mainly due to the daytime heating (consistently with  
655 Parker et al. 2005a). Consequently, the meridional pressure gradient accelerates the meridional wind  
656 component as soon as the daytime turbulent mixing weakens around sunset (blue arrows Figure  
657 16b). This cool and moist nocturnal LLJ supplies the low layers with water vapour, shifting the ITD  
658 northward. The atmosphere is then destabilised south of the ITD that eventually leads to convection  
659 triggering and precipitation over the southern part of the domain. This schematic view is relevant

660 for the whole heatwave sequence simulated by EXPL, except that the processes at play move  
661 northward as the heatwave settles down over the Central Sahel, following the location of the ITD.  
662 The water vapour increase over the Central Sahel by nocturnal meridional advection then enhances  
663 the downwelling longwave radiation at the surface, thereby leading to high surface temperatures  
664 and extreme humid heat.

665 Taking EXPL as a reference, Figures 16c,d then exhibit how the previous scenario is modified  
666 when the deep convection scheme is switched on (PARAM). First note that a short precipitating  
667 event in the southern part of the domain during the first hours of the PARAM simulation, due  
668 probably to the model spin-up, induces near-surface temperature colder than in EXPL. Then, the  
669 deep convection scheme triggers in the early afternoon, thereby enhancing the low-level  
670 atmospheric cooling (consider the yellow colour in Figure 16a versus the orange colour in Figure  
671 16c), mostly due to precipitation evaporation within the atmosphere and enhanced surface  
672 evaporation. Cloud radiative effects do not contribute much to the simulation differences. The  
673 colder temperatures in the southern part of the domain both weaken the vertical mixing within the  
674 boundary layer and increase the meridional pressure gradient. As a result, the LLJ initiates earlier in  
675 the day (see the blue arrow in Figure 16c) and becomes stronger than in EXPL up to the ITD  
676 (Figure 16d versus 16b). The increased advection of cooler and wetter air to the northern latitudes  
677 supports an ITD at higher latitudes, and reduces the heatwave intensity. This scenario slowly  
678 propagates northward. Then, the increased low-level moisture in the northern latitudes can help  
679 trigger new convective events there (Figure 16d), which will further contribute to reduce the  
680 heatwave intensity compared to that simulated in EXPL.

681 Based also on a convection-permitting model, Marsham et al. (2013) and Birch et al. (2014, see  
682 their Figure 1) previously identified similar behaviour of the LLJ for the summer mean state, but

683 with distinct balance of processes in summer. Together with our findings, this emphasises that the  
684 low-level meridional pressure gradient can be influenced by three main processes whose balance  
685 varies across the annual cycle of the Sahelian climate:

- 686 1. the amount of solar radiation reaching the surface varies significantly between spring and  
687 summer. On April 15, the sun reaches the zenith at  $10^{\circ}\text{N}$  near the south of our domain  
688 against  $18^{\circ}\text{N}$  on August 1 at the north of the domain used in Marsham et al. (2013) and  
689 Birch et al. (2014). The pattern of the radiative heating therefore impacts the shape and  
690 intensity of the pressure gradient,
- 691 2. the latent heat release due to condensation within deep convective clouds is stronger during  
692 the core monsoon season. This heating leads to a surface pressure decrease (because the  
693 depth between two pressure layers is proportional to its virtual temperature),
- 694 3. the rainfall evaporation cooling in the sub-cloud layer is expected to be stronger during the  
695 pre-monsoon season, as the air is drier. This process contributes to increase the low level  
696 pressure.

697 During the pre-monsoon period, the cooling by precipitation evaporation below cloud base  
698 dominates the warming by the latent heat release aloft so that finally deep convection mainly leads  
699 to pressure increase at the surface. This feature is less pronounced in EXPL as deep convection is  
700 weaker and located more to the south.

701 In this study, we have shown, for the pre-monsoon period, that the premature convection  
702 triggering, which occurs in the simulation where the deep convection parameterization is turned on  
703 damps the daytime pressure decrease because of reduced heating by radiation fluxes and enhanced  
704 cooling by rainfall evaporation. Then the southerly wind between the spurious precipitation band  
705 and the ITD is reinforced before the evening and efficiently shifts the ITD northward. Overall, the

706 budget analysis shows that the impact of parameterized deep convection is not restricted to changes  
707 in the thermodynamics but also involves profound modifications of the dynamics. This conclusion  
708 is in line with Marsham et al. (2013) and Birch et al. (2014). However, we also find that it involves  
709 distinct balances among processes which are likely due to differences in the large-scale Sahelian  
710 environment between the pre-monsoon and full monsoon seasons.

711 The interest of convection-permitting simulations for the study of the West African monsoon has  
712 been demonstrated by numerous studies (e.g., Diongue et al. 2002, Marsham et al. 2013, Beucher et  
713 al. 2014, Berthou et al. 2020). Our results further underline the non-intuitive added value of a  
714 convection-permitting resolution for the study of West African heatwaves, i.e. for meteorological  
715 events typically characterised by relatively low precipitation. More studies are now needed to  
716 evaluate the modelling of these humid heatwaves by regional and global models in more detail and  
717 to assess the role of the convection parameterization in their performances.

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### 723 ***IN MEMORIAM***

724 During the final stage of the writing of this paper, Françoise Guichard suddenly and  
725 unexpectedly passed away. She was the inspiration behind this work. The final form of this paper  
726 had her full approval. Her departure is a terrible loss for our community and we suffer already  
727 without her guidance and contributions. Her spontaneity, goodwill, generosity and warmth along

728 with her intellect, curiosity, open-mindedness and integrity will be greatly missed. Her shining will  
 729 not extinguish as a true Breton lighthouse.

730 ***APPENDIX A1: Figure***

**Appendix A1:** 2-m air temperatures as a function of the latitude and time from the April 10 to 19, 2010 for the anomalies of the ECMWF analyses compared to BEST (ECMWF-BEST). The ITD location is indicated with the black line (2-m water vapour mixing ratio contour of 8 g/kg). Values are at 0600 UTC typical of the hour of the daily minimum. All data are averaged over the longitudes [4.5°W-2.5°E]. Numbers on the bottom and top of the colour bars are respectively the minimum and the maximum level of the chart.

731 This figure shows the 10-day sequence of the 2-m temperature anomalies at 0600 UTC of the  
 732 ECMWF model versus the latitude. The initialisation and nudging model appears colder than the  
 733 BEST product north of 13°N.

734 ***APPENDIX A2: Evolution equation of the thermodynamic variables***

735 The budget equation of a given variable  $\alpha$  reads:

736 
$$\frac{\overline{\partial\alpha}}{\partial t} = -\underbrace{\overline{\vec{u} \cdot \nabla \alpha}}_{Adv_\alpha} + \sum_p \overline{S\alpha_p} \quad (1)$$

737 where  $\vec{u} = (u, v, w)$  is the wind vector,  $Adv_\alpha$  the total advection of  $\alpha$  (horizontal and vertical) and  
 738  $\overline{S\alpha_p}$  is the  $p^{\text{th}}$  source term of  $\alpha$ . We discard the relaxation and small-scale dissipation source terms as  
 739 they are an order of magnitude smaller than other sources (except at the borders for the relaxation).  
 740 The over-bar denotes an average over the lowest 500 m of the Boundary Layer (BL) and the  
 741 longitudes [4.5W-2.5E]. Note that the results do not change significantly when analysing the surface  
 742 layer only or the lowest 1000 m of the BL. In practice, these budgets are calculated for each grid  
 743 point and at each time step before averaging.

744 The turbulence source term  $\overline{Turb_\alpha}$  gathers the contributions of the shallow convection scheme  
745  $\overline{Turb\_th_\alpha}$  (“th” standing for BL “thermals”) and that of the sub-grid (eddy-viscosity) turbulence  
746 scheme  $\overline{Turb\_sb_\alpha}$ . In PARAM, the source due to the deep convection scheme  $\overline{DeepCV_\alpha}$  is added to  
747 that of the microphysics scheme  $\overline{Micro\Phi_\alpha}$  in order to be more comparable to the EXPL  
748 microphysical source. Indeed, the deep convection scheme represents a part of the microphysical  
749 processes. This comparison remains qualitative as the deep convection scheme also includes vertical  
750 transport which is explicitly represented by the advection term in EXPL.

751 Following Equation (1), the budget of the potential temperature  $\theta$  reads:

$$752 \quad \frac{\partial \overline{\theta}}{\partial t} = \overline{Adv_\theta} + \overline{Turb_\theta} + \overline{Micro\Phi_\theta} + \overline{DeepCV_\theta} + \overline{Rad} \quad (2)$$

753 where  $\overline{Rad}$  is the radiation source term.

754 For the  $i^{\text{th}}$  component of the wind  $\vec{u}$ , the budget equation (1) reads:

$$755 \quad \frac{\partial \overline{u_i}}{\partial t} = \overline{Adv_{u_i}} + \overline{Turb_{u_i}} - \underbrace{\frac{1}{\rho_{ref}} \frac{\partial P}{\partial x_i}}_{\overline{Pres_{u_i}}} - \underbrace{2 \varepsilon_{i,j,k} \Omega_j u_k}_{\overline{Cor_{u_i}}} + \overline{Curv_{u_i}} \quad (3)$$

756 where  $\overline{Pres_{u_i}}$  is the pressure force with  $P$  the pressure and  $\rho_{ref}$  the reference density,  $\overline{Cor_{u_i}}$  is the  
757 Coriolis force with  $\Omega_j$  the  $j^{\text{th}}$  component of the earth angular velocity and with the Einstein sum  
758 convention and  $\overline{Curv_{u_i}}$  is the curvature force (not shown because of an order of magnitude smaller  
759 than the other forces).

760 For  $\alpha=r_v$ , the water vapour mixing ratio, equation (1) reads:

$$761 \quad \frac{\partial \overline{r_v}}{\partial t} = \overline{Adv_{r_v}} + \overline{Micro\Phi_{r_v}} + \overline{Turb_{r_v}} + \overline{DeepCV_{r_v}} \quad (4)$$

762 In some figures, each budget  $p$  of the variable  $\alpha$  is time integrated from the beginning of the  
763 simulation until the time  $t$ :

764 
$$\int_0^t \overline{S \alpha_p} \cdot dt \quad (5)$$

765 Their sum is equal to the variable, by considering also  $\overline{Adv_\alpha}$  as a source term:

766 
$$\overline{\alpha_{(t)}} = \int_0^t \frac{\partial \overline{\alpha}}{\partial t} \cdot dt = \sum_p \int_0^t \overline{S \alpha_p} \cdot dt \quad (6)$$

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1087 ***List of Figures legend***

**FIGURE 1** Ten-day average from April 10 to 19, 2010 of **(a)** the BEST daily maximum 2-m temperature anomaly and ERA-I 10-m wind (0600 UTC) **(b)** the BEST daily minimum 2-m temperature and ERA-I 10-m wind (0600 UTC) anomalies. On panel (b), the climatological and 10-day average Inter-Tropical Discontinuity (ITD) at 0600 UTC are indicated with the thin and bold grey lines, respectively. The ITD is defined as the 8 g/kg ERA-I 2-m water vapour mixing ratio isoline. BEST and ERA-I climatologies are computed over the period 1980-2010. The black array emphasises the simulated domain (5°W-3°E, 10°N-18°N). The black dots locate the in-situ measurement sites: Agoufou (15.34°N, 1.48°W), Niamey (13.48°N, 2.17°E) and Ouagadougou (12.35°N, 1.52°W).

**FIGURE 2** Hovmöller diagram from April 10 to 19, 2010 of the BEST daily **(a)** minimum and **(b)** maximum air near-surface (2-m) temperatures and **(c)** and **(d)** their respective anomalies computed with respect to the BEST 1980-2010 climatology. The ITD location is indicated with the black line (2-m water vapour mixing ratio contour of 8 g/kg), at 0600 UTC in Tn charts and at 1800 UTC in Tx charts, using ECMWF operational analyses. Numbers on the bottom and top of the colour bars are respectively the minimum and the maximum level of the chart. All data are averaged over [4.5°W-2.5°E].

**FIGURE 3** Ten-day average from April 10 to 19, 2010 of the CERES **(a)** Integrated Water Vapour **(b)** cloud fraction and **(c)** 55 µm Aerosol Optical Depth anomalies. Subsequent panels indicate various contributions to the surface Downwelling radiative flux anomalies, based on CERES data: **(d)** total ShortWave (SWD) **(e)** SWD cloud radiative effects (CRE) and **(f)** SWD aerosol radiative effects (ARE); **(g)** total LongWave (LWD) **(h)** LWD CRE and **(i)** LWD ARE; **(j)** sum of the CRE and ARE, **(k)** CRE and **(l)** ARE. On panel **(a)**, the climatological and 10-day average Inter-Tropical Discontinuity (ITD) at 0600 UTC are indicated with the thin and bold grey lines, respectively. On panel **(b)** ERA-I 10-m wind and on panel **(c)** wind anomalies (0600 UTC) are computed. The CERES anomalies are computed over the period 2000-2015 and the ITD and wind over the period 1980-2010. See Figure 1 for more details.

**FIGURE 4** Hovmöller diagram from April 10 to 19, 2010 of the 2-m pressure anomalies with respect to the initial state on a sliding average over 24 hours to avoid the barometric tide (shade, hPa) for the ECMWF operational analyses. The surface precipitation from the 3-hourly TRMM 3B42 estimates has been superposed (thick black isolines from 1 to 41

every 10 mm/day). The ITD location is indicated with the dashed black line (2-m water vapour mixing ratio contour of 8 g/kg), using the ECMWF operational analyses. All data are averaged over [4.5°W-2.5°E].

**FIGURE 5** Time (UTC) evolution of the 3h-averaged Integrated Water Vapour (IWV, mm, thick lines) and surface precipitation (mm/day, thin lines) at the **(a)** Niamey and **(b)** Ouagadougou sites for the observations (black), and the EXPL (green) and PARAM (blue) simulations. Note that GPS data at Niamey are missing until April 12, 2010 early afternoon.

**FIGURE 6** Time series in Agoufou for the **(a)** the 2-m temperature every 15 mn (°C) **(b)** 10-m zonal wind component every hour (m/s) **(c)** the 2-m water vapour mixing ratio every 15 mn (g/kg) and the precipitation at the surface averaged over 1h (mm/day) **(d)** 10-m meridional wind component every hour (m/s) **(e)** day-mean of the sensible heat flux (thin line) and of the latent heat flux (thick line) at the surface (W/m<sup>2</sup>) **(f)** day-mean of LWD (dash), SWD (thin line) and net radiation (thick line) at the surface (W/m<sup>2</sup>) for the observation Automatic Weather Station (AWS) in black, the explicit run in green and the run with the parameterized convection scheme in blue. The MésoNH and ECMWF data are averaged over the four points surrounding the Agoufou station. ECMWF analyses are every 6 hours (black stars) and ECMWF forecasts every 3 hours (red stars). Note in Figure 6c that no rainfall is observed and only the PARAM simulation makes rain.

**FIGURE 7** Minimum (left) and maximum (right) of the daily 2-m temperatures as a function of the latitude and time from April 10 to 19, 2010 for the explicit run (EXPL, first row), the parameterized deep convection run (PARAM, third row) and for the associated anomalies with respect to BEST from April 10 to 19, 2010 (second and fourth rows). The ITD location (2-m water vapour mixing ratio contour of 8 g/kg) is indicated by the green lines for EXPL and blue lines for PARAM, at 0600 UTC in T<sub>n</sub> charts and at 1800 UTC in T<sub>x</sub> charts. The PARAM ITD isline breaks on April 17 0600 UTC because it is located north of 18°N. All figures are averaged over the longitudes [4.5°W-2.5°E]. Numbers on the bottom and top of the colour bars are respectively the minimum and the maximum level of the chart.

**FIGURE 8** Hovmöller diagram of T<sub>2m</sub> (°C) difference between PARAM and EXPL simulations (PARAM-EXPL) averaged over the longitudes [4.5°W-2.5°E]. The ITD seen as the 8 g/kg water vapour mixing ratio at 2-m above the surface is superimposed (green line for EXPL, blue line for PARAM). Numbers on the bottom and top of the colour bars are respectively the minimum and the maximum level of the chart.

**FIGURE 9** Surface precipitation (isolines from 1 to 41 every 10 mm/day) as a function of the latitude and time for April 10 to 19, 2010 **(a)** for the EXPL simulation (green) and **(b)** for the PARAM simulation (blue), averaged over 3 hours. Their respective ITD location is also indicated with the dashed line, averaged over 3 hours. To ease the comparison, the TRMM B42 precipitation observation product (also drawn in Figure 4) and the ITD using ECMWF operational analyses every 6 hours are superimposed in black on the panels (a) and (b). The ITD location is computed from the 2m water vapour mixing ratio contour of 8 g/kg. **(c)** presents the 10-day surface precipitation as a function of the latitude and **(d)** the surface precipitation diurnal composite averaged over the 10 days and the whole domain [4.5°W-2.5°E, 10.5°N-17.5°N]. All data are averaged over [4.5°W-2.5°E].

**FIGURE 10** Vertical profile at Niamey (Local Time=UTC+1) between the April 10, 2010 1200 UTC and April 20, 0000 UTC except the April 13&16, 1200 UTC. Composites at 0000 UTC for the left column and at 1200 UTC for the right column of the water vapour mixing ratio, the potential temperature, the zonal and meridional wind. Stars refer to the radiosoundings, black lines to the ECMWF reanalyses, green lines to the explicit MésoNH run and blue lines to the run with the deep parameterized convection scheme. For the simulation, the closest grid point to observations has been used.

**FIGURE 11** Time-evolution during the first two days of the potential temperature budget for the averaged latitudes [13°N-15°N] **(a)** for EXPL tendencies (K/day), **(b)** for PARAM-EXPL tendencies (K/day), **(c)** for PARAM-EXPL time integrated tendencies (K). **(d to f)** the same as (a to c) but for [11°N-13°N]. On each panel, the dashed lines are the sum of the solid lines. All terms are averaged over the 500 lowest metres and longitudes [4.5°W-2.5°E]. See Equation (1), (2),(5)&(6) in Appendix A2 for more details on the different terms.

**FIGURE 12** First two days' average difference PARAM-EXPL for the potential temperature (shade, K), pressure (black contours, hPa), water vapour mixing ratio (green contours,  $r_v$ , g/kg) and horizontal wind speed (purple contours, m/s) as a function of the latitude and altitude. The PARAM 8 g/kg of  $r_v$  symbolising its ITD is also drawn in thick dark blue line. All variables are averaged over longitudes [4.5°W-2.5°E]. Numbers on the bottom and top of the colour bars are respectively the minimum and the maximum level of the chart.

**FIGURE 13** Meridional wind budgets first two days' evolution of the main time integrated tendencies (m/s) for the averaged latitudes [12°N-14°N] for (a) EXPL and (b) PARAM-EXPL. On each panel, the dashed lines are the sum of the solid lines. All figures are averaged over the 500 lowest metres and longitudes [4.5°W-2.5°E]. See Equation (1),(3), (5)&(6) in Appendix A2 for more details on the different terms.

**FIGURE 14** From top to bottom, budgets of the (a&b) 2-m potential temperature (K/day), (c&d) 2-m water vapour mixing ratio ( $r_v$ , g.kg<sup>-1</sup>.day<sup>-1</sup>) and (e&f) 10-m meridional wind (m.s<sup>-1</sup>.day<sup>-1</sup>) for the PARAM (dashed line) and EXPL (solid line) simulations as a function of latitude averaged over (left) the first 2 days (April 10 0000 UTC to 12 0000 UTC) and (right) the next 2 days (April 12 0000 UTC to 14 0000 UTC). In brown the tendency, in red the total advection budget, in green the turbulence term, in blue the addition of the microphysical terms and the deep convection scheme, in yellow the radiation, in dark blue the pressure force, in pink the Coriolis force. Each variable is also plotted in black (where 296 K has been subtracted from the temperature, 14 g.kg<sup>-1</sup>.day<sup>-1</sup> has been added to  $r_v$  and 35 m/s has been added to the wind). The surface heat flux (H) is added in (a) in orange (divided by 10 then shifted of 5 W/m<sup>2</sup>). The surface latent heat flux (LE) is added in (b) in cyan (divided by 10 then shifted of 12 W/m<sup>2</sup>). To symbolise the mean ITD, an arrow at the latitude where  $r_v=8$  g/kg is added on the left hand side of each graph in green for EXPL and blue for PARAM. All variables are averaged over the 500 lowest metres and longitudes [4.5°W-2.5°E] and smoothed on 10 points of latitude (i.e. around 0.5°) to ease the visualisation.

**FIGURE 15** Ten-day evolution of the potential temperature hourly tendencies time integrated for the difference PARAM-EXPL and the averaged latitudes (a) [15°N-17°N] (b) [13°N-15°N]. On each panel, the dashed lines are the sum of the solid lines. Figures are averaged over the 500 lowest metres and longitudes [4.5°W-2.5°E]. See Equation (1), (2),(5)&(6) in Appendix A2 for more details on the different terms.

**FIGURE 16** Schema illustrating the main differences between EXPL (upper line) and PARAM (lower line) on average over the period, differentiating afternoon/night situation. Here, EXPL is taken as the reference. The pressures at the south and north borders are fixed by the ECMWF coupling and then are the same for both simulations. The schematic pressure strength varies from the lowest to the highest following the rank: L L / h H H, identical for both simulations, making it possible to compare latitude for a given simulation or values between both simulations. The low level temperature varies from the cooler to the warmer following the colours from the dark violet, dark blue, blue, yellow, orange to red. The horizontal arrows width and length are proportional to the low-level advection strength cooling and moistening in blue and heating and drying in orange. The vertical arrows represent the turbulent mixing in the boundary layer. The latent heat release is identified by a red ellipse and the evaporative cooling by a blue ellipse. These latent heats are larger when the ellipses are bigger.