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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

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

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

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

1 INTRODUCTION

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

Obviously, HW are prolonged periods of extreme temperatures but plethora of metrics exist depending on the issues at hand (see Perkins 2015 for a review). Because they involve distinct processes, night-time and daytime HW are often distinguished and their identifications rely on the use of daily minimum and maximum near-surface (2-m) dry-bulb temperature, Tn and Tx respectively (e.g., Robinson et al. 2001). In the perspective of having more representative ways of quantifying the human body sensation and stress to extreme heat, other variables, such as the wet-bulb temperature or the apparent temperature, may also be considered (e.g., Steadmann 1984, Willett and Sherwood 2012, Zhao et al. 2015, Raymond et al. 2021).

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

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

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

In the Sahel in spring, soils are climatologically already dry. Therefore, soil moisture-temperature feedback is very unlikely. In contrast, atmospheric water vapour is found to play an important role on Sahelian HW (Oueslati et al. 2017, Largeron et al. 2020, Bouniol et al. 2021), possibly leading to extreme humid heat events. Different changes in the atmospheric circulation can lead to a water vapour increase: intensification of the Saharan heat low (Knippertz and Fink 2008, Barbier 2017), northward shift of the intertropical front (Guichard et al. 2009, Couvreux et al. 2010, Largeron et al. 2020), occurrence of tropical plumes (Fröhlich et al. 2013), or presence of Rossby waves fostering south-westerly wind anomalies above the western regions of West Africa (Fontaine et al. 2013). Locally, water vapour, clouds, desert dusts, through their interactions with the radiative and turbulent boundary layer processes also control the surface air temperatures (Guichard et al. 2009, Bouniol et al. 2012, Gounou et al. 2012, Bain et al. 2010, Fontaine et al. 2013, Largeron et al. 2020). However, a detailed knowledge and understanding of their respective effects and interactions during Sahelian heatwave episodes is still lacking.

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

Over West Africa, Marsham et al. (2013), Birch et al. (2014) and Vellinga et al. (2016) showed that the representation of the west african monsoon is improved with convection-permitting simulations. More recent studies confirm this finding (e.g., Stratton et al. 2018, Vizy and Cook 2019). However, few studies focussed on HW with convection-permitting simulations (Zhang et al. 2020, Ramamurthy and Bou-Zeid 2017) and, to our knowledge, none over the Sahel. Kendon et al. (2019) indicate that the projections performed with a convection-permitting model over some regions of Africa, like the Sahel, show an increased length of dry spells whereas the use of coarser resolution models with parameterized deep convection indicates the reverse effect due to a less realistic triggering and propagation of convective systems. In the context of HW, it might be counter-intuitive to think that the deep convection parameterization can be detrimental. We shall see however that its ability to not trigger convection and associated rainfall is in fact critical for capturing the HW properties and processes at play.

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

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

2 DATA, METHOD AND SIMULATIONS

2.1 Reference datasets

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

First, numerous high-frequency ground-based measurements implemented during the African Monsoon Multi-disciplinary Analysis (AMMA) project (Redelsperger et al. 2006) are examined. This notably includes observations collected by the AMMA-CATCH network at three sites (Galle et al. 2018). Hereafter, we mainly use measurements from the Agoufou site in Mali (15°34'N, 1°48'W, local time: UTC+0h), located in the Central Sahel (Mougin et al. 2009). At this site, an automatic weather station provides near-surface air temperature, relative humidity, rainfall, wind speed and direction as well as Downwelling and Upwelling ShortWave (*SWD* and *SWU*) and LongWave (*LWD* and *LWU*) surface radiative fluxes with a 15-min time step (Guichard et al. 2009). An eddy-correlation station also provides estimates of the surface sensible (*H*) and latent (*LE*) heat fluxes at a 30-min time step (Timouk et al. 2009). Note that we reached very similar conclusions with the data of the Niger Wankama AMMA-CATCH site located in the southern Sahel (Leauthaud et al. 2017). Vertically Integrated Water Vapour amount (IWV) is derived from GPS stations deployed at the Niamey (13.48°N, 2.17°E, local time: UTC+1h) and Ouagadougou (12.35°N, 1.52°W, local time: UTC+0h) sites, at an hourly frequency (Bock et al. 2008). The Aerosol Optical Depth (AOD) provided by several sunphotometers, from the AERONET network (Holben et al. 1998), is also used. Finally, radiosondes launched twice a day at Niamey (at 0000 and 1200 UTC) allow us to characterise the vertical structure of the lower atmosphere. We only had access to their low resolution version.

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

- the Berkeley Earth System Temperature dataset (BEST, Rohde et al. 2013) which incorporates a large ensemble of weather ground stations and provides the daily minimum and maximum 2-m temperatures,
- the Cloud and Earth's Radiant Energy System (CERES) SYN1deg dataset, which combines measurements made by several spatial instruments, in particular the Moderate Resolution Imaging Spectroradiometrer (MODIS) to provide radiative fluxes, cloud cover, total AOD at 0.55 μm and IWV (see Doelling et al. 2013 for details).

Data from the European Centre for Medium-range Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-I, Dee et al. 2011) are also used to document the heatwave synoptic situation. The data is provided on a 0.75°x0.75° horizontal grid every 6 hours.

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

2.2 Radiative fluxes and surface energy balance

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

$$Rnet = SWN + LWN = H + LE + G$$

where
$$SWN = SWD - SWU$$

and
$$LWN = LWD - LWU$$

G is the ground heat flux. *SWD, SWU, LWD* and *LWU* are defined as positive. The net radiative fluxes (*Rnet, SWN* and *LWN*), as well as *G*, are counted positive downward. Surface turbulent fluxes are counted positive upward.

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

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

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

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

2.3 Model set-up

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

2.3.1 Configuration

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

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

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

The runs are initialised on April 10, 2010 at 0000 UTC using the ECMWF operational analysis (0.25° horizontal resolution). The soil water indexes of the three bucket layers are also initialised using the ECMWF operational analysis. Some sensitivity tests to the soil moisture initialisation are discussed in the Supporting Information. The wind components, potential temperature and water vapour mixing ratio are nudged at the domain lateral boundaries towards the 6-hourly ECMWF operational analyses. The nudging timescale is chosen small enough (25 s) to well constrain the simulations.

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

2.3.2 Parameterizations

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

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

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

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

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

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

3 THE APRIL 2010 HEATWAVE

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

3.1 Large-scale circulation and heatwave sequence

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

Figure 2 illustrates the HW sequence from April 10 to 19, 2010 as a function of latitude. Tn progressively increases over the period from about 27°C to 31°C. These values correspond to anomalies greater than 4°C, especially in southern Mali (north of 14°N). North of 11°N, Tx remain high all over the 10-day period, above 40 to 43°C (anomalies of about 2 to 3°C). There are however some intermittences of a few degrees between April 14 and 17. On April 19, Tx reaches almost 45°C between 14°N and 16°N. Combined with the high Tn, these conditions are particularly tough for local populations.

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

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

Figure 4 also shows the time-evolution of the latitudinal distribution of the TRMM-3B42 rainfall during the 10-day period. The rain falls south of the ITD and follows its northward shift. The precipitation estimates show several precipitating events from April 12 to 15, south of 13°N (with a 3-hourly rainfall maximum of 41 mm/day). On April 16, 2010 another event, quite north for the season, extends from 13°N to 17.5°N. The amount remains weak though (3-hourly rainfall maximum of 9 mm/day) and is questionable as in such an arid region, the occurrence of rainfall evaporation may significantly bias the TRMM 3B42 estimates of surface precipitation (Dinku et al. 2011).

3.2 Heatwave sequence at the local scale

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

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

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

From April 10 to 13, the 2-m temperature exhibits a strong diurnal amplitude (around 20°C, Figure 6a). Tx reaches 42.5 to 44°C, while the surface air layer remains very dry, the mixing ratio being below 3 g/kg (Figure 6c). Weak south-easterly winds prevail (Figure 6b,d) until the monsoon surge reaches Agoufou on April 13.

Within a couple of hours on April 13, the monsoon surge arrival in Agoufou induces a wind reversal from easterly to westerly (Figure 6b). The water vapour mixing ratio dramatically increases by about 10 g/kg (Figure 6c). This induces a substantial jump in the LWD by more than 50 W/m² (dashed line in Figure 6f) and a decrease of the net energy loss by longwave radiation, mostly due to its clean-sky contribution (not shown). This mitigates the night cooling (Guichard et al. 2009, Largeron et al. 2020) and thereby reduces the diurnal temperature range. Tn increases by 6°C from April 12 to 13, then again by 5°C from April 13 to 14 and finally reaches 34°C on April 16. Tx is less affected by the monsoon surge, and continues to vary between 42°C and 44°C. A computation of the wet-bulb temperature combining temperature and humidity following Zhao et al. (2015) indicates a significant to extreme heat stress for the population during the monsoon surge, the hardest time being April 15, 1400 UTC with 44°C as dry-bulb temperature and 8.5 g/kg of water vapour mixing ratio, that is to say 34°C as wet-bulb temperature.

Finally, on April 17, the monsoon surge retreats southward, inducing a new wind reversal from south-westerlies to north-easterlies (Figure 6b,d) and a slow decrease of near-surface water vapour (Figure 6c) until April 20. A sharp decrease in the surface SWD radiative (~150 W/m^2, thin lines in Figure 6f), net surface radiative (~70 W/m2, thick line in Figure 6f) and surface sensible heat fluxes (~40 W/m2, thin line in Figure 6e) is concomitantly observed. Based on CERES data, this can be attributed to cloud cover and AOD increases (CRE and ARE induce a surface net SW decrease of 30 and 40 W/m2, respectively, not shown). In contrast, the surface LWD radiative flux remains strong (dashed line in Figure 6f). Consistently, Tx slightly decreases from April 15 (except on April 19), while Tn remains high (except on April 17).

4 EVALUATION OF THE SIMULATIONS

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

4.1 The simulated heatwave sequence at large scale

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

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

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

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

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

EXPL exhibits a similar temporal variability and latitudinal extent of precipitation compared to the TRMM 3B42 reference (Figure 9), although not always with the right timing or intensity. EXPL captures the latitudinal precipitation distribution rather well over the 10-day period, with a slight overestimate south of 12.5°N and a slight underestimate north of 13.5°N (Figure 9c). North of 13.5°N, the two observed events of April 16 and the April 17-18 night occur in EXPL but the rainfall evaporates before reaching the ground (not shown). This explains the negative bias but, as already mentioned, the occurrence of TRMM-3B42 surface rainfall may also be questionable. Finally, on average, EXPL triggers rain approximately at the right time in the day but the rainfall peak is slightly too early, between 1700 and 2100 UTC against near midnight for TRMM 3B42 estimates (Figure 9d).

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

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

4.2 The simulated heatwave sequence at the local scale

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

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

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

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

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

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

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

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

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

The radiosoundings at Niamey indicate a significant diurnal cycle of the boundary layer with a strong nocturnal temperature inversion (Figure 10c) and a 1.5-km-deep well-mixed convective boundary layer not yet completely developed at 1200 UTC (Figure 10b,d) typical of this pre-monsoon period (Lothon et al. 2008, Guichard et al. 2009, Gounou et al. 2012). The “nocturnal” Low Level Jet (LLJ) is also typical of the end of the dry season in the Sahel, as deep convection does not disturb its nighttime development (Parker et al. 2005a, Lothon et al. 2008). At 0000 UTC, the observed LLJ is not yet fully developed (only its southerly component is visible, stars in Figure 10e,g). At 1200 UTC, it has already begun to retreat and only its remaining bell-shape maximum near 500 m above ground level is noticeable (Figure 10f,h). The LLJ advects water vapour from the southern regions, which is vertically redistributed during daytime by the Boundary Layer (BL) turbulent mixing. The BL is topped by a weak African easterly jet around 4 km height, which results from the thermal wind balance (Parker et al. 2005b). The wind shear between the LLJ and the African easterly jet amplifies the entrainment at the top of the BL (Canut et al. 2010, Gounou et al. 2012).

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

In the entire BL in Niamey and at the surface in Agoufou, PARAM spurious precipitation leads to strong negative temperature biases and strong positive water vapour content and wind biases. In the next section, we investigate the underlying processes and mechanisms behind the differences of these two simulations through the use of thermodynamic and wind budgets.

5 PHYSICAL MECHANISMS

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

5.1 Early stage of the simulations

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

Over the [13°N-15°N] latitude band, none of the simulations produce precipitation (see also Figure 9). Consistently, the deep convection scheme does not trigger ($\overline{DeepCV_\theta}$ is zero) and the microphysical scheme does not cool the BL ($\overline{Micro\Phi_\theta}$ is zero). The EXPL total potential temperature tendency (Figure 11a) is positive during daytime as a result of heating by solar radiation and turbulence. It is negative during nighttime and early morning due to longwave radiation and advective cooling (associated with the Low Level Jet LLJ). Note that for this thin 500m-high layer, the turbulence source (Turb) is driven by the surface temperature evolution. The subgrid turbulence scheme contribution encompasses the shallow convection scheme contribution. The subgrid turbulence warms the layer during daytime due to the large positive sensible heat flux, while it slightly cools the layer at night following the weakly negative sensible heat flux. The shallow convection scheme only acts during daytime: it first warms the layer between 0800 and 1100 UTC and then cools it until sunset due to the enhanced mixing between the lowest 500 m of the BL and the upper part of the BL which depth largely overpasses 500 m in the afternoon (not shown).

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

The stronger advective cooling $\overline{Adv_\theta}$ in PARAM during nighttime is thus key to understanding the temperature differences between PARAM and EXPL at latitudes without precipitation since the beginning of the simulation. Figure 12 indicates that, on average over the first two days, PARAM is significantly colder than EXPL between the surface and about 2.5 km above the ground, associated with higher pressure, at PARAM rainy latitudes (up to 12.5°N, cf. Figure 9) but also beyond between 12.5°N and 15°N. Indeed, the stronger PARAM pressure gradient drives stronger meridional winds (purple contours in Figure 12), especially when the LLJ forms during nighttime, contributing to the increased advective cooling and moistening (green contours in Figure 12) further north. Above 3.5 km, and south of 15°N, PARAM is warmer, consistently with the occurrence of latent heat release by deep convection. EXPL simulates weak convection during the first two days, which is confined to the southernmost latitudes.

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

Figure 13 further analyses the meridional wind (v) budget at latitudes intermediate between the two previous latitude bands. The meridional component results from a complex equilibrium between different terms (Equation 3 in Appendix A2, Figure 13a). The pressure term is the primary force that drives wind acceleration. It is the major driver of the monsoon surge, during this pre-monsoon season (consistently with Couvreux et al. 2010). Its diurnal cycle is mainly linked to the daytime heating. The Coriolis force accelerates *v* as long as the easterly winds are established. The advection term also accelerates the meridional component. Only the daytime subgrid turbulence ($\overline{Turb_{sb}}$) slows it. Indeed, the thermals accelerate the wind over the 500 lowest meters (the wind is maximum around 500 m above the ground so that $\overline{Turb_{th}}$ is positive from the surface to 500 m and negative above, not shown) but $\overline{Turb_{sb}}$ dominates the turbulent term which finally decelerates *v*. The sum of those contributions leads mainly to a southerly meridional wind (dash black line).

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

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

Figure 14 (first column) shows the budget of both simulations as a function of latitude. Budgets are averaged over the first two days and still between the surface and 500 m. As already discussed, PARAM is colder than EXPL south of the ITD up to 1K (black lines). The bias originates from rainy PARAM latitudes (11°N to 13°N) with a cooling $\overline{Micro\Phi_{\theta}}$+$\overline{DeepCV_{\theta}}$ term (Figure 14a), associated with a stronger surface latent heat flux (cyan line in Figure 14c). The radiation (yellow lines) also plays an important role in the θ tendency difference due to PARAM colder surface, associated with a smaller surface sensible heat flux (orange lines in Figure 14a) and subgrid turbulence. The advection propagates this bias northward up to 15.5°N, due to a slightly stronger PARAM south wind accelerated by a stronger pressure force and temperature meridional gradient (Figure 14e).

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

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

5.2 Following evolution

For the next two days (April 12 0000 UTC to 14 0000 UTC), the main balances of the first two days hold but shifted 1.5° northward due to the synoptic monsoon surge imposed by the nudging at the boundaries (Figure 14b,d,f). Indeed the southerly component of the wind increases (up to +2 m/s at 13°N on average over two days, black lines in Figure 14e,f) mainly driven by the intensification of the meridional pressure force (blue lines). Due to the Coriolis force and to a lesser extent to the advection (not shown), the westerly component of the wind also increases boosting the low level water vapour for all latitudes (up to +4 g/kg at 13°N, black lines in Figure 14c,d). As already explained, the radiation budget, by the intermediate of the long wave greenhouse effect, increases the temperature tendency at latitudes with no rainfall for each simulation (compare the yellow lines in Figure 14a,b). So that despite the advective and microphysical cooling, the heatwave strengthens up to 1°C (black lines, stronger effect for EXPL).

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

Figure 15 presents the differences between the integrated terms of the potential temperature budgets of EXPL and PARAM for the entire period. The scenario described in the previous section (notably Figure 11) remains valid over most of the 10-day period, providing that it follows the northward migration of the monsoon surge. In the [13°N-15°N] band, the precipitation has a strong impact in PARAM with a larger sink in potential temperature due to the $\overline{Micro\Phi_\theta}$+$\overline{DeepCV_\theta}$ and the radiative terms from April 13 onward. The cooling by advection is larger in EXPL starting on April 14 as observed south of 13°N for April 12-13 (not shown). Further north, in the [15°N-17°N] band, the cooling by advection is larger in PARAM except in the EXPL rainy nights of April 14-15 and April 16-17 (Figure 15a).

6 CONCLUSION

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

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

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

The analysis of thermodynamic and dynamic budgets in the low atmospheric levels emphasises a balance between the daytime heating/drying by turbulence and the night-time cooling/moistening by advection, largely operated by the nocturnal Low Level Jet (LLJ). The dynamic budget further highlights the pressure gradient as the major driver of the monsoon pulsation, during this pre-monsoon season (consistently with Couvreux et al. 2010) and also shown by Birch et al. (2014) during the core monsoon season. Figures 16a,b synthesise the behaviour of the EXPL simulations. The lower surface pressures are located near the ITD in the middle of the domain and are more pronounced in the afternoon and early night mainly due to the daytime heating (consistently with Parker et al. 2005a). Consequently, the meridional pressure gradient accelerates the meridional wind component as soon as the daytime turbulent mixing weakens around sunset (blue arrows Figure 16b). This cool and moist nocturnal LLJ supplies the low layers with water vapour, shifting the ITD northward. The atmosphere is then destabilised south of the ITD that eventually leads to convection triggering and precipitation over the southern part of the domain. This schematic view is relevant for the whole heatwave sequence simulated by EXPL, except that the processes at play move northward as the heatwave settles down over the Central Sahel, following the location of the ITD. The water vapour increase over the Central Sahel by nocturnal meridional advection then enhances the downwelling longwave radiation at the surface, thereby leading to high surface temperatures and extreme humid heat.

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

Based also on a convection-permitting model, Marsham et al. (2013) and Birch et al. (2014, see their Figure 1) previously identified similar behaviour of the LLJ for the summer mean state, but with distinct balance of processes in summer. Together with our findings, this emphasises that the low-level meridional pressure gradient can be influenced by three main processes whose balance varies across the annual cycle of the Sahelian climate:

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

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

In this study, we have shown, for the pre-monsoon period, that the premature convection triggering, which occurs in the simulation where the deep convection parameterization is turned on damps the daytime pressure decrease because of reduced heating by radiation fluxes and enhanced cooling by rainfall evaporation. Then the southerly wind between the spurious precipitation band and the ITD is reinforced before the evening and efficiently shifts the ITD northward. Overall, the budget analysis shows that the impact of parameterized deep convection is not restricted to changes in the thermodynamics but also involves profound modifications of the dynamics. This conclusion is in line with Marsham et al. (2013) and Birch et al. (2014). However, we also find that it involves distinct balances among processes which are likely due to differences in the large-scale Sahelian environment between the pre-monsoon and full monsoon seasons.

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

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IN MEMORIAM

During the final stage of the writing of this paper, Françoise Guichard suddenly and unexpectedly passed away. She was the inspiration behind this work. The final form of this paper had her full approval. Her departure is a terrible loss for our community and we suffer already without her guidance and contributions. Her spontaneity, goodwill, generosity and warmth along with her intellect, curiosity, open-mindedness and integrity will be greatly missed. Her shining will not extinguish as a true Breton lighthouse.

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.

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

APPENDIX A2: Evolution equation of the thermodynamic variables

The budget equation of a given variable α reads:

$$\overline{\frac{\partial \alpha}{\partial t}} = \underbrace{-\overline{\vec{u}.\vec{\nabla}\alpha}}_{\overline{Adv_\alpha}} + \sum_p \overline{S\alpha_p} \tag{1}$$

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

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

Following Equation (1), the budget of the potential temperature θ reads:

$$\overline{\frac{\partial \theta}{\partial t}} = \overline{Adv_\theta} + \overline{Turb_\theta} + \overline{Micro\Phi_\theta} + \overline{DeepCV_\theta} + \overline{Rad} \tag{2}$$

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

For the ith component of the wind $\vec{u}$, the budget equation (1) reads:

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

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

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

$$\overline{\frac{\partial r_v}{\partial t}} = \overline{Adv_{r_v}} + \overline{Micro\Phi_{r_v}} + \overline{Turb_{r_v}} + \overline{DeepCV_{r_v}} \tag{4}$$

In some figures, each budget p of the variable α is time integrated from the beginning of the simulation until the time t:

$$\int_0^t \overline{S\alpha_p}.dt \tag{5}$$

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

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

REFERENCES

Argüeso D., Di Luca A., Perkins-Kirkpatrick S. et al. (2016) Seasonal mean temperature changes control future heat waves, *Geophys. Res. Lett.*, n°43, p7653–7660, https://doi.org/10.1002/2016GL069408

Azongo D., Awine T., Wak G. et al. (2012) A time series analysis of weather variability and all-cause mortality in the Kasena-Nankana districts of northern Ghana, 1995–2010, *Global Health Action*, n°5, p14-23, https://doi.org/10.3402/gha.v5i0.19073

Bador M., Terray L., Boé J., et al. (2017) Future summer mega-heatwave and record-breaking temperatures in a warmer France climate, *Environmental Research Letters,* vol.12, n°7, https://doi.org/10.1088/1748-9326/aa751c

Bain CL., Parker CM., Taylor L. et al. (2010) Observation of the nocturnal boundary layer associated with the West African Monsoon, *Mon. Weather Rev.*, n°138, p3142-3156, https://doi.org/10.1175/2010 MWR 3287.1

Barbier J. (2017) Extrêmes climatiques - les vagues de chaleur au printemps sahélien, Thèse de doctorat en Océan, atmosphère, climat, Toulouse INPT, http://www.theses.fr/2017INPT0122/document

Barbier J., Guichard F., Bouniol D. et al. (2018) Detection of intraseasonal large-scale heat waves: Characteristics and historical trends during the Sahelian Spring, *J. of Clim.,* vol.31, issue 1, p61-80, https://doi.org/10.1175/JCLI-D-17-0244.1

Bechtold P., Bazile E., Guichard F. et al. (2001) A mass-flux convection scheme for regional and global models, *QJRMS*, n°127, p869-886 , https://doi.org/10.1256/smsqj.57308

Berthou S., Kendon E., Chan S. et al. (2020) Pan-European climate at convection-permitting scale: a model intercomparison study. *Clim Dyn,* n°55, p35-59, https://doi.org/10.1007/s00382-018-4114-6

Beucher F., Lafore JP., Karbou F. et al. (2014) High-resolution prediction of a major convective period over West Africa, *QJRMS*, vol.140, issue 682 part A, p1409-1425, https://doi.org/10.1002/qj.2225

Birch C., Parker D., Marsham J. et al. (2014) A seamless assessment of the role of convection in the water cycle of the West African Monsoon, *JGR*, vol.119, n°6, p2890-2912, https://doi.org/10.1002/2013JD020887

Black E., Blackburn M., Harrison G. et al. (2004) Factors contributing to the summer 2003 European heatwave, *Weather*, n°59(8), p217-223, https://doi.org/10.1256/wea.74.0

Bock O., Bouin M., Doerflinger E. et al. (2008) West African Monsoon observed with ground-based GPS receivers during African Monsoon Multidisciplinary Analysis (AMMA), *JGR: Atm.*, vol.113, issue D21, https://doi.org/10.1029/2008JD010327

Bougeault P. (1982) Cloud-ensemble relations based on the gamma probability distribution for the higher-order models of the planetary boundary layer, *J. Atmos. Sci*., n°39, p2691-2700

Bougeault P. and Lacarrère P. (1989) Parameterization of orography-induced turbulence in a meso-beta scale model, *Mon. Weather Rev.*, n°117, p1870–1888

Bouniol D., Couvreux F., Kamsu-Tamo PH. et al. (2012) Diurnal and seasonal cycle of cloud occurrences, types and radiative impact over West Africa, *J. of Applied Met. And Clim.*, n°51, p534-553, https://doi.org/10.1175/JAMC-D-11-051.1

Bouniol D., Guichard F., Barbier J., Couvreux F. and Roehrig R. (2021) Sahelian heat wave characterization from observational data sets, *Journal of Geophysical Research: Atmospheres*, n°126, e2020JD034465, https://doi.org/10.1029/2020JD034465

Caniaux G., Redelsperger JL. and Lafore JP. (1994) A Numerical Study of the Stratiform Region of a Fast-Moving Squall Line. Part I: General Description and Water and Heat Budgets, *J. Atmos. Sci.*, vol51, n°14, p2046-2074, https://doi.org/10.1175/1520-0469(1994)051<2046:ANSOTS>2.0.CO;2

Canut G., Lothon M., Saïd F. et al. (2010) Observation of entrainment at the interface between monsoonal flow and the saharan air layer, *QJRMS*, n°136(S1), p34-46, https://doi.org/10.1002/qj.471

Chaboureau JP. and Bechtold P. (2005) Statistical representation of clouds in a regional model and the impact on the diurnal cycle of convection during Tropical Convection, Cirrus and Nitrogen Oxides (TROCCINOX), *J. Geophys. Res.*, n°110(D17), https://doi.org/10.1029/2004JD005645

Christidis N., McCarthy M. and Stott P. (2020) The increasing likelihood of temperatures above 30 to 40 °C in the United Kingdom, *Nat Commun,* n°11, p3093-3102, https://doi.org/10.1038/s41467-020-16834-0

Couvreux F., Guichard F., Bock O. et al. (2010) Synoptic variability of the monsoon flux over West Africa prior to the onset, *QJRMS*, n°135, p159-173, https://doi.org/10.1002/qj.473

Cowan T., Purich A., Perkins S. et al. (2014) More frequent, longer, and hotter heat waves for Australia in the twenty-first century, *J. Climate*, n°27, p5851–5871, https://doi.org/10.1175/JCLI-D-14-00092.1

Cuxart J., Bougeault P. and Redelsperger JL. (2000) A turbulence scheme allowing for mesoscale and large eddy simulations, *QJRMS*, n°126, p1-30, https://doi.org/10.1002/qj.49712656202

Dee D., Uppala S., Simmons A. et al. (2011) The ERA-Interim reanalysis: Configuration and performance of the data assimilation system, *QJRMS*, n°137, p553-597, https://doi.org/10.1002/qj.828

Della-Marta P.M., Luterbacher J., von Weissenfluh H. et al. (2007) Summer heat waves over western Europe 1880–2003, their relationship to large-scale forcings and predictability, *Clim. Dyn.,* n°29 (2–3), p251–275, https://doi.org/10.1007/s00382-007-0233-1

Déqué M., Calmanti S., Christensen O. B. et al. (2017) A multi-model climate response over tropical Africa at +2°C, *Climate Serv.*, n°7, p87–95, https://doi.org/10.1016/j.cliser.2016.06.002

Diallo F., Hourdin F., Rio C. et al. (2017) The surface energy budget computed at the grid-scale of a climate model challenged by station data in West Africa, *J. of Advances in modeling earth systems,* n°9, p2710–2738, https://doi.org/10.1002/2017MS001081

Diboulo E., Sié A. and Rocklöv J. (2012) Weather and mortality: A 10 year retrospective analysis of the Nouna Health and Demographic Surveillance System, Burkina Faso, *Global Health Action*, n°5, p6–13, https://doi.org/10.3402/gha.v5i0.19078

Ding T., Qian W. and Yan Z. (2010) Changes in hot days and heat waves in China during 1961–2007, *Int. J. Climatol.*, n°30(10), p1452–1462, https://doi.org/10.1002/joc.1989

Dinku T., Ceccato P. and Connor S. J. (2011) Challenges of satellite rainfall estimation over 626 mountainous and arid parts of east Africa, *Int. J. Remote Sens.*, n°30, p5965–5979, https://doi.org/10.1080/01431161.2010.499381

Diongue A., Lafore JP., Redelsperger, JL. et al. (2002) Numerical study of a Sahelian synoptic weather system: Initiation and mature stages of convection and its interactions with the large-scale dynamics, *QJRMS*, n°128, p1899-1927, https://doi.org/10.1256/003590002320603467

Doelling D., Loeb N., Keyes D. et al. (2013) Geostationary enhanced temporal interpolation for CERES flux products. *J. Atmos Ocean Technol.*, n°30(6), p1072–1090, https://doi.org/10.1175/JTECH-D-12-00136.1

Easterling D., Horton B., Jones P. et al. (1997) Maximum and minimum temperature trends for the globe, *Science*, n°277, p364–367, https://doi.org/10.1126/science.277.5324.364

Fink A., Brücher T., Krüger A. et al. (2004) The 2003 European summer heatwaves and drought: Synoptic diagnosis and impacts, *Weather,* n°59, p209–216, https://doi.org/10.1256/wea.73.04

Fischer E., Seneviratne S., Lüthi D. et al. (2007a) Contribution of land-atmosphere coupling to recent European heatwaves, *Geophys. Res. Lett.,* n°34, L06707, https://doi.org/10.1029/2006GL029068

Fischer E., Seneviratne S., Vidale P. et al. (2007b) Soil moisture-atmosphere interaction during the 2003 European heatwave, *J. of Clim*, vol.20, p5081-5099, https://doi.org/10.1175/JCLI4288.1

Fontaine B., Janicot S. and Monerie PA. (2013) Recent changes in air temperatures, heat waves occurrences and atmospheric circulation in Northern Africa, *JGR*,vol.118(15), p8536-8552, https://doi.org/10.1002/jgrd.50667

Fouquart Y. and Bonnel B. (1980) Computations of solar heating of the earth's atmosphere – A new parameterization, *Beitraege zur Physik der Atmosphaere*, n°53, p35-62

Fröhlich L., Knippertz P., Fink A. et al. (2013) An Objective Climatology of Tropical Plumes, *J. Climate*, n°26, p5044–5060, https://doi.org/10.1175/JCLI-D-12-00351.1

Galle S., Grippa M., Peugeot C. et al. (2018) AMMA-CATCH, a critical zone observatory in West Africa monitoring a region in transitio, V*adose Zone J.*, 17:180062, https://doi.org/10.2136/vzj2018.03.006

García-Herrera R., Díaz J., Trigo R. M. et al. (2010) A Review of the European Summer Heat Wave of 2003, *Critical Reviews in Environmental Science and Technology,* n°40, p267-306, https://doi.org/10.1080/10643380802238137

Gounou A., Guichard F. and Couvreux F. (2012) Observations of diurnal cycles over West African meridional transect: pre-monsoon and full-monsoon seasons, *Boundary-Layer Met.*, n°144, p329–357, https://doi.org/10.1007/s10546-012-9723-8

Gruhier C., De Rosnay P., Hasenauer S. et al. (2010) Soil moisture active and passive microwave products: intercomparison and evaluation over a Sahelian site, *Hydrology and Earth system Sc.*, n°14(1), https://doi.org/10.5194/hessd-6-5303-2009

Guichard F., Kergoat L., Mougin E. et al. (2009) Surface thermodynamics and radiative budget in the Sahelian Gourma: Seasonal and diurnal cycles, *Journal of Hydrology*, n° 375, p161-177, https://doi.org/10.1016/j.jhydrol.2008.09.007

Guichard F., Kergoat L., Hourdin F. et al. (2015) Le réchauffement climatique observé depuis 1950 au Sahel in “Evolutions récentes et futures du climat en Afrique de l’Ouest : Evidences, incertitudes et perceptions”, “*Les sociétés rurales face aux changements climatiques et environnementaux en Afrique de l’Ouest, Sultan B. et al., IRD editions*, p23-42

Harris I., Jones P., Osborn T. et al. (2014) Updated high-resolution grids of monthly climatic observations - the CRU TS3.10 Dataset, *International journal of climatology*, n°34(3), p623-642, https://doi.org/10.1002/joc.3711

Holben B., Eck T., Slutsker I. et al. (1998) AERONET – A federated instrument network and data archive for aerosol characterization, *Remote Sens. Environ.*, n°66, p1-16, https://doi.org/10.1016/S0034-4257(98)00031-5

Huffman G., Bovin D., Nelkin E. et al. (2007) The TRMM Multisatellite Precipitation Analysis (TMPA): Quasi-global, multiyear, combined sensor precipitation estimates at fine scales, *J. Hydrometeorol.*, n°8, p38-55, https://doi.org/10.1175/JHM560.1

IPCC (2021) Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change [Core Writing Team, Pachauri R. and Meyer L. (eds.)], *IPCC, Geneva, Switzerland*, 151 pp.

Kain J. and Fritsch M. (1990) A One-Dimensional Entraining/Detraining Plume Model and Its Application in Convective Parameterization, *JAS*, n°47, p2784-2802, https://doi.org/10.1175/1520-0469(1990)047<2784:AODEPM>2.0.CO;2

Kendon E., Stratton R., Tucker S. et al. (2019) Enhanced future changes in wet and dry extremes over Africa at convection-permitting scale, *Nat Commun* n°10 (1794), https://doi.org/10.1038/s41467-019-09776-9

Knippertz P. and Fink A. (2008) Dry-season precipitation in tropical West Africa and its relation to forcing from the extratropics, *Mon.Weather Rev.*, n°136, p3579-3596, https://doi.org/10.1175/2008MWR2295.1

Knippertz P. and Todd M. (2012) Mineral dust aerosols over the Sahara: Meteorological controls on emission and transport and implications for modeling, *Review of Geophys.,* n°50(1), https://doi.org/10.1029/2011RG000362

Lac C., Chaboureau JP., Masson V. et al. (2018) Overview of the MesoNH model version 5.4 and its applications, *Geosci. Model Dev.*, n°11, p1929-1969, https://doi.org/10.5194/gmd-11-1929-2018

Lafore JP., Stein J., Asencio N. et al. (1998) The MesoNH Atmospheric Simulation System. Part I: Adiabatic formulation and control simulations, *Ann. Geophys.*, n°16, p90-109, https://doi.org/10.1007/s00585-997-0090-6

Largeron Y., Guichard F., Roehrig R. et al. (2020) The April 2010 North African heatwave: when the water vapor greenhouse effect drives nighttime temperatures, *Clim. Dyn.*, n°54 p3879-3905, https://doi.org/10.1007/s00382-020-05204-7

Leauthaud C., Cappelaere B., Demarty J. et al. (2017) A 60-year reconstructed high-resolution local meteorological data set in Central Sahel (1950–2009): evaluation, analysis and application to land surface modelling, *International Journal of Climatology*, n°37(5), p2699-2718, https://doi.org/10.1002/joc.4874

Lothon M., Said F., Lohou F. et al. (2008) Observation of the diurnal cycle in the low troposphere of West Africa, *Mon. Wea. Rev.*, n°136, p3477–3500, https://doi.org/10.1175/2008MWR2427.1

Marsham J., Dixon N., Garcia-Carreras L. et al. (2013) The role of moist convection in the West African monsoon system: Insights from continental-scale convection-permitting simulations, *Geophysical Research letters,* n°40, p1843-1849, https://doi.org/10.1002/grl.50347

Masson V., Champeaux JL., Chauvin C. et al. (2003) A global database of land surface parameters at 1 km resolution for use in meteorological and climate models, *J. Climate*, n°16, p1261-1282, https://doi.org/10.1175/1520-0442(2003)16<1261:AGDOLS>2.0.CO;2

Masson V., Le Moigne P., Martin E. et al. (2013) The SURFEXv7.2 land and ocean surface platform for coupled or offline simulation of earth surface variables and fluxes, *Geosci. Model Dev*., n°6, p929–960, https://doi.org/10.5194/gmd-6-929-2013

Meehl G. and Tebaldi C. (2004) More intense, more frequent and longer lasting heat waves in the 21st century, *Science*, n°305(5686), p.94-997, https://doi.org/10.1126/science.1098704

Miralles D., van den Berg M., Teuling A. et al. (2012) Soil moisture-temperature coupling: A multiscale observational analysis, *Geophys. Res. Lett.,* n°39, L21707, https://doi.org/10.1029/2012GL053703

Miralles D., Teuling A., van Heerwaarden C. et al. (2014) Mega-heatwave temperatures due to combined soil desiccation and atmospheric heat accumulation, *Nature Geoscience,* n°7, p345–349, https://doi.org/10.1038/ngeo2141

Mlawer E, Taubman S, Brown P. et al. (1997) Radiative Transfer for Inhomogeneous Atmospheres: RRTM, a Validated Correlated-k Model for the Longwave, *JGR*, n°102, p16663-16682, https://doi.org/10.1029/97JD00237

Moralles S., Cook E.R., Barichivich J. et al. (2020) Six hundred years of South American tree rings reveal an increase in severe hydroclimatic events since mid-20th century, *PNAS*, n°117(29), p16816-16823, https://doi.org/10.1073/pnas.2002411117

Morcrette JJ. (2002) The surface downward longwave radiation in the ECMWF forecast system, *J. Climate*, n°15, p1875–1892, https://doi.org/10.1175/1520-0442(2002)015<1875:TSDLRI>2.0.CO;2

Moron V., Oueslati B., Pohl B. et al. (2016) Trends of mean temperatures and warm extremes in norther tropical Africa (1961–2014) from observed and PPCA reconstructed time series, *J. Geophys. Res. Atmos.*, n°121, p5298–5319, https://doi.org/10.1002/2015JD024303

Mougin E., Hiernaux P., Kergoat L. et al. (2009) The AMMA-CATCH Gourma observatory site in Mali: relating climatic variations to changes in vegetation, surface hydrology, fluxes and natural resources, *J. Hydrol.*, n°375(1), p14-33, https://doi.org/10.1016/j.jhydrol.2009.06.045

Nikulin G., Jones C., Giorgi F. et al. (2012) Precipitation climatology in an ensemble of Cordex-Africa regional climate simulations. *J. Climate*, n°25(18), p6057-6078, https://doi.org/10.1175/JCLI-D-11-00375.1

Noilhan J. and Planton S. (1989) A simple parameterization of land surface processes for meteorological models, *Mon. Weather Rev.*, n°117(3), p536–549, https://doi.org/10.1175/1520-0493(1989)117<0536:ASPOLS>2.0.CO;2

Ogi M., Yamazaki K. and Tachibana Y. (2005) The summer northern annular mode and abnormal summer weather in 2003, *Geophys. Res. Lett.,* n°32, L04706, https://doi.org/10.1029/2004GL021528

Oueslati B., Pohl B., Moron V. et al. (2017) Characterization of Heat Waves in the Sahel and Associated Physical Mechanisms, *J. Climate*, n°30, p3095–3115, https://doi.org/10.1175/JCLI-D-16-0432.1

Parker D., Burton R., Diongue-Niang A. et al. (2005a) The diurnal cycle of the West African monsoon circulation, *QJRMS*, n°131, p2839-2860, https://doi.org/10.1256/qj.04.52

Parker D., Thorncroft C., Burton R. et al. (2005b) Analysis of the African Easterly Jet using aircraft observations from the JET2000 experiment, *QJRMS*, n°131(608), p1461-1482, https://doi.org/10.1256/qj.03.189

Pergaud J., Masson V., Malardel S. et al. (2009) A Parameterization of Dry Thermals and Shallow Cumuli for Mesoscale Numerical Weather Prediction, *BLM*, n°132, p83-106, https://doi.org/10.1007/s10546-009-9388-0

Perkins S. (2015) A review on the scientific understanding of heatwaves: Their measurement, driving mechanisms and changes at the global scale, *Atmos. Res*., n°164-165, p242-267, https://doi.org/10.1016/j.atmosres.2015.05.014

Perkins S., Argüeso D. and White C. (2015) Relationships between climate variability, soil moisture and on Australian heatwaves, *J. Geophys. Res.*, n°120(16), p8144-8164, https://doi.org/10.1002/2015JD023592

Peyrillé P., Lafore JP. and Boone A. (2016) The annual cycle of the West African monsoon in a two-dimensional model: mechanisms of the rain-band migration, *QJRMS*, n°142, p1473-1489, https://doi.org/10.1002/qj.2750

Pinty JP and Jabouille P. (1998) A mixed-phase cloud parameterization for use in a mesoscale non-hydrostatic model: simulations of a squall line and of orographic precipitation, *Conf. On Cloud Physics, Everett, WA, USA, Amer. Meteor. Soc.*, p217-220

Pospichal B., Bou Karam D., Crewell S. et al. (2010) Diurnal cycle of the intertropical discontinuity over West Africa analysed by remote sensing and mesoscale modelling, QJRMS, n°136, p92-106, https://doi.org/10.1002/qj.435

Quesada B., Vautard R., Yiou P. et al. (2012) Asymmetric European summer heat predictability from wet and dry southern winters and springs, *Nat. Clim. Chang.,* n°2 (10), p736–741, https://doi.org/10.1038/nclimate1536

Ramamurthy P. and Bou-Zeid E. (2017) Heatwaves and urban heat islands: A comparative analysis of multiple cities, *J. Geophys. Res. Atmos.*, n°122, p168-178, https://doi.org/10.1002/2016JD025357

Ramanathan V., Cess R., Harrison E. et al. (1989) Cloud-radiative forcing and climate: results from the earth radiation budget experiment, *Science*, n°243(4887), p57-63, https://doi.org/10.1126/science.243.4887.57

Raymond et al. (2021) On the controlling factors for globally extreme humid heat, *Geophysical Research letters,* n°48, e2021GL096082, **https://doi.org/10.1029/2021GL096082**

Redelsperger JL. and Sommeria G. (1986) Three-dimensional simulation of a convective storm: Sensitivity studies on subgrid parameterization and spatial resolution, *J. Atmos. Sci.*, n°43(22), p2619-2635, https://doi.org/10.1175/1520-0469(1986)043<2619:TDSOAC>2.0.CO;2

Redelsperger JL., Thorncroft C., Diedhiou A. et al. (2006) African Monsoon Multidisciplinary Analysis: An international research project and field campaign, *Bulletin of the Amer. Meteor. Soc.*, n°87(12), p1739-1746, https://doi.org/10.1175/BAMS-87-12-1739

Robinson P. (2001) On the Definition of a Heat Wave, *J. Appl. Meteor.*, n°40, p762–775, https://doi.org/10.1175/1520-0450(2001)040<0762:OTDOAH>2.0.CO;2

Rohde R., Muller R., Jacobsen R. et al. (2013) Berkeley earth temperature averaging process, *Geoinf for Geostat: An Overview 1:2,* https://doi.org/10.4172/2327-4581.1000103

Roehrig R., Bouniol D., Guichard F. et al. (2013) The present and future of the west African Monsoon: a process-oriented assessment of CMIP5 simulation along the AMMA transect, *J. Clim.*, n°26(17), https://doi.org/10.1175/JCLI-D-12.00505.1

Russo S., Dosio A., Graversen R. et al. (2014) Magnitude of extreme heat waves in present climate and their projection in a warming world, *J. Geophys. Res. Atmos.,* n°119(22), p12500-12512, https://doi.org/10.1002/2014JD022098

Schoetter R., Cattiaux J. and Douville H. (2015) Changes of western European heat wave characteristics projected by the CMIP5 ensemble, *Clim. Dyn.*, n°45(5-6), p1601-1616, https://doi.org/10.1007/s00382-014-2434-8

Seneviratne S., Nicholls N., Easterling D. et al. (2012) Changes in climate extremes and their impacts on the natural physical environment, *in Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation, edited by C. B. Field et al., A Special Report of Working Groups I and II of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, UK, and New York, NY, USA,* p109-230

Steadman R. (1984) A Universal Scale for Apparent Temperature, *J. of Applied Met. And Clim.*, n°23 (12), p1674-1684, https://doi.org/10.1175/1520-0450(1984)023<1674:AUSOAT>2.0.CO;2

Stéfanon M., Drobinski P., D'Andrea F. et al. (2012) Effects of interactive vegetation phenology on the 2003 summer heatwaves, *J. Geophys. Res.: Atmospheres*, n°117, D24103, p1-15, https://doi.org/10.1029/2012JD018187

Stott P., Stone D. and Allen M. (2004) Human contribution to the European heatwave of 2003, *Nature,* n°432, p610-614**,** https://doi.org/10.1038/nature03089

Stratton R., Senior C., Vosper S. et al. (2018) A Pan-African Convection-Permitting Regional Climate Simulation with the Met Office Unified Model: CP4-Africa, *J. Climate*, n°31(9), p3485–3508, https://doi.org/10.1175/JCLI-D-17-0503.1

Sutton R., Dong B. and Gregory J. (2007) Land/sea warming ratio in response to climate change: IPCC AR4 model results and comparison with observations, *Geophysical Research Letters*, n°34(2), https://doi.org/10.1029/2006GL028164

Taylor K., Stouffer R. and Meehl G. (2012a) An overview of CMIP5 and the experiment design, *Bull. Amer. Meteor. Soc*., n°93(4), p485-498, https://doi.org/10.1175/BAMS-D-11-00094.1

Taylor C., de Jeu R. and Guichard F. (2012b) Afternoon rain more likely over drier soils, *Nature,* n°489(7416), p423-426, https://doi.org/10.1038/nature11377

Tegen I., Hollrig P., Chin M. et al. (1997) Contribution of different aerosol species to the global aerosol extinction optical thickness: Estimates from model results, *JGR*, n°102, p23895-23915, https://doi.org/10.1029/97JD01864

Timouk F., Kergoat L., Mougin E. et al. (2009) Response of surface energy balance to water regime and vegetation development in a Sahelian landscape, *Journal of hydrology*, n°375(1-2), p178-189, https://doi.org/10.1016/j.jhydrol.2009.04.022

Vautard R., Gobiet A., Jacob D. et al. (2013) The simulation of European heat waves from an ensemble of regional climate models within the EURO-CORDEX project, *Clim Dyn,* n°41, p2555-2575, https://doi.org/10.1007/s00382-013-1714-z

Vellinga M., Roberts M., Luigi P. et al. (2016) Sahel decadal rainfall variability and the role of model horizontal resolution, *Geophys. Res. Lett.*, n°43, p326–333, https://doi.org/10.1002/2015GL066690

Vizy E. and Cook K. (2019) Understanding the summertime diurnal cycle of precipitation over sub-Saharan West Africa: regions with daytime rainfall peaks in the absence of significant topographic features, *Clim Dyn,* n°52, p2903-2922, https://doi.org/10.1007/s00382-018-4315-z

Weisheimer A., Doblas-Reyes F., Jung T. et al. (2011) On the predictability of the extreme summer 2003 over Europe, *Geophys Res Lett*, n°38, L05704, https://doi.org/10.1029/2010GL046455

Willett K.M. and Sherwood S. (2012) Exceedance of heat index thresholds for 15 regions under a warming climate using the wet-bulb globe temperature, *Int. J. Climatol.*, **32**, p161–177, https://doi.org/10.1002/joc.2257.

Zaitchik B., Macalady A., Bonneau L. et al. (2006) Europe's 2003 heat wave: a satellite view of impacts and land–atmosphere feedbacks, *Int. J. Climatol.*, n°26(6), p743-769, https://doi.org/10.1002/joc.1280

Zhang Z., Li Y., Chen F. et al. (2020) Evaluation of convection-permitting WRF CONUS simulation on the relationship between soil moisture and heatwaves, *Clim Dyn,* **n°55,** p235-252, https://doi.org/10.1007/s00382-018-4508-5

Zhao Y., Ducharme A., Sultan B. et al (2015) Estimating heat stress from climate-based indicators: present-day biases and future spreads in the CMIP5 global climate model ensemble**,** *Environ. Res. Lett***., vol. 10, n°8,** https://doi.org/10.1088/1748-9326/10/8/084013

Zhou L., Dickinson R., Dai A. et al. (2010) Detection and attribution of anthropogenic forcing to diurnal temperature range changes from 1950 to 1999: comparing multi-model simulations with observations, *Climate Dynamics*, n°35(7-8), p1289-1307, https://doi.org/10.1007/s00382-009-0644-2

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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^2) **(f)** day-mean of LWD (dash), SWD (thin line) and net radiation (thick line) at the surface (W/m^2) 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 Tn charts and at 1800 UTC in Tx charts. The PARAM ITD isoline 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 T2m (°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, rv, 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, $g.kg^{-1}.day^{-1}$) and **(e&f)** 10-m meridional wind ($m.s^{-1}.day^{-1}$) 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 $g.kg^{-1}.day^{-1}$ 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/m²). The surface latent heat flux (LE) is added in (b) in cyan (divided by 10 then shifted of 12 W/m²). 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 1*5* 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 *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.