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The impacts of contrasting atmospheric thermodynamics on surface-convection interactions across West-Africa

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The impacts of contrasting atmospheric thermodynamics on surface-convection interactions across West-Africa

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ABSTRACT

This study aims to characterize the thermodynamics of the lower atmosphere across West Africa (WA), on the basis of observations from the AMMA project. A dataset made of several thousand high-resolution soundings is used for this purpose. Moist-convective related indexes, boundary layer (BL) parameters, and a combination of two indexes proposed by Findell and Eltahir (2003) (FE03) are computed for each sounding. These latter indexes are well suited for distinguishing between aspects of surface-atmosphere interactions involving atmospheric vertical structures.

The variability of the low levels is found to be stronger in the Sahelian zone than in the Soudanian zone, from daily scale up to the monsoon season (JJAS). The FE03 indexes suggest that surface-atmosphere feedbacks are actually significant over West-Africa, but that the feedback loops change sign with latitude and season. Thus, the thermodynamic environment at low levels is consistent with the idea of daytime convection being either suppressed or favoured over wet surface versus dry ones, depending on latitude and seasonal variations.

Furthermore, variations of the lifting condensation level (LCL) and the level of free convection (LFC) are consistent with the previous analysis. For instance, in the Sahel, a strong diurnal cycle of the LCL is found, especially early in the season, while the LFC does not fluctuate much during the day, but shifts downwards from June to August. The LCL is however not such a good indicator of BL height when the low levels are the driest. This is particularly true prior to the monsoon onset, on days when none of the daytime BL convective plumes reach their LCL. In that case, moist convective development is limited by high convective inhibition (CIN) rather than convective available potential energy (CAPE).

This finding is consistent with observations that daytime convection is not particularly favoured over wet land surfaces. Instead, it may be initiated more easily over dry soils when the atmospheric low levels are warm and unstable, as can be found over the Sahel prior to the rainfall onset.

1) Introduction

Moist convection over West Africa is mostly known for its manifestation in the form of large, mesoscale-organized convective systems, such as squall lines, which have been extensively studied (Zipser 1977, Chong 1987, Lafore and Moncrieff 1989). This spatial organization has been related to wind shear and tropospheric dryness (Barnes and Sieckman 1984). Both factors contribute to explain the propagation speed and self-sustained character of these phenomena. At larger scale, strong relationships have been identified between synoptic African easterly waves and convection (Reeds et al. 1977), although the nature of the interactions at play is still unclear, and appears to vary along the season (Lavaysse et al 2006) and latitude (Duvel 1990). This points to the importance of other factors.

Moist convection is also characterized by a well defined diurnal timing over West Africa. Widespread cumulus clouds typically develops during daytime hours, while mesoscale convective systems (MCS) often emerge from there afterwards, starting in afternoon or evening hours. A number of studies have shown that convection and rainfall -largely of convective nature- display well defined diurnal modulations with maxima taking place during late evening and nighttime hours (Mohr 2004). Thus, the diurnal cycle of radiation imposes another prominent mode of temporal organization to moist convection over this wide tropical continental region, which further connects convective processes to surface features at sub-diurnal scales.

On another side, the emphasis put on interactions between surface processes and atmospheric convection has significantly evolved and enriched during the last decades, in particular in terms of scales of concern and mechanisms. Charney (1975) put forwards that a regional-scale increase of albedo would affect rainfall amount via the radiative energy loss it generates. A number of recent studies have also highlighted effective manifestations of interactions at smaller and shorter scales, as reviewed in Nicholson (2000). They point to their significance over West African (Koster et al. 2004). Observed interactions between surface and moist-convective processes do not appear to comply to a single explanation scheme though. For instance, in the Sahel, Taylor and Lebel (1998) identify positive feedback between surface evapotranpiration and rainfall at small spatial scales (a few tens of km) while Taylor and Ellis (2006) found that afternoon convective initiation was favoured over dry soils in the Sahel at scale on the order of hundred km.

In recent years, the awareness of the low-levels significance in the monsoon systems has been increasing (Sperber & Yasunari 2007). Indeed, the monsoon flow takes place within the low levels. However, the mechanisms actually operating in the development, dissipation, and more broadly life cycle of moist convective phenomena are not well identified. Therefore, not much is known about how precisely and at which scales the low levels participate to these mechanisms.

At sub-diurnal local scales, basic considerations of convective boundary layer growth suggest a significance of the couplings between the surface state and the thermal stability in the low levels on the initiation of daytime moist convection (Ek and Mahrt 1991). Namely, the daytime convective BL growth will be less when the lapse rate (Γ=-dT/tz) is smaller (more stable atmosphere), so that the BLH may never reach the height of the morning LFC during the following daytime hours. Therefore, the height of the LFC has to decrease for moist convection to develop. This can be achieved via an increase of BL thetae, which is more likely when surface latent heat flux dominates over sensible heat flux and reduces BL top entrainment of drier air (Fig. 1a,b). Conversely, BL growth will be larger within a more unstable atmosphere (dtheta/dt smaller), so that the LFC will not decrease much during daytime. Therefore, the BL must grow significantly, which is more likey when surface sensible heat flux is high, such as over dry soil (Fig. 1a). Findell and Eltahir (2003a), hereafter referred to FE03a, extended this framework by adding considerations of low level humidity, an further showed how early morning vertical structures of atmospheric low levels can induce varied controls on feedbacks between soil moisture and daytime convection triggering, in terms of likeliness and sign.

During the monsoon, surface net radiation is high and surface states and low levels quite varied over West Africa. In the present study, we address the question of how the low-level thermodynamics can play a role in the interactions between surface and convective processes, via the arguments presented above.

The vertical structure of the low levels is typically marginally to not-well-resolved by conventional datasets, which only retain a few levels from the original sounding measurements (typically less than four in the lowest 500m AGL). Furthermore, such data had been until recently very few in this region (Parker et al. 2008), and dataset from observational campaigns not numerous and limited in time (JET2000, Thorncroft et al 2003) and, or spatial extend (HAPEX-Sahel, Goutorbe et al. 1994). They provide valuable but patchy insight. ECMWF analyses and reanalyses are, to our knowledge, the only NWP available source of high-vertical resolution datasets. However, the low levels constitute a location where these products are likely to be the less accurate, especially when and where no data are assimilated. They will, in that case, heavily rely on the model parametrizations. This is illustrated in Fig. 2, with a 1200 UTC field of CAPE as derived from two NWP analyses the 28 August 2005, during the core of the monsoon season. The magnitude of CAPE differ significantly among the two estimates. Furthermore, they depict quite distinct synoptic patterns. In this study, we take advantage of the unprecedent dataset made of several thousands of high-resolution soundings collected across West Africa within AMMA (Redelsperger et al. 2006) in 2006 (Parker et al. 2008), to perform our calculations. Namely, we make use of the indexes proposed by FE03a, together with more conventional indexes documenting basic features of the low levels, including boundary layer and moist-convection related properties.

2) Data and methodology

2.1) sounding data

High-resolution sounding data were collected at the sixteen radiosounding (RS) sites shown in Fig. 3 (full disks on map, see Parker et al. 2008 for a summary of sounding stations operated in 2006). Among them, six (red disks in Fig. 2) performed four soundings per day, at 0000, 0600, 1200 and 1800 UTC, during either early June to end of September or part of this 4-month period. A summary of numbers of soundings available per site is provided in table 1; see Nuret et al. (2008) for additional information on sonde types. In addition, vertical profiles obtained from more than 300 dropsondes, 100 driftsondes as well as soundings launched from research vessels

The sampling of the original data ranges from one to a few seconds; this corresponds to a vertical resolution of about ten metres. Data from each individual sounding have been interpolated on a grid, common to all sounding data, that uses the height above ground level as the vertical coordinate. This grid extends from the surface up to 19 km AGL, comprises 58 levels, and its resolution, stretched, ranges from a few tens of metres below 500m AGL to about 400 m above 4 km AGL. This grid is close to the resolution grid of the ECMWF model.

2.1) indexes

Boundary layer and convection-related indexes have been computed from the interpolated profiles. Several diagnostics of the daytime convective BL height (zi) were tested. Hereafter, zi is defined from 12Z to 18Z as the height of the first encountered level k where the virtual potential tempetature $\theta v(k)$ is greater than the [0,z(k)] average θv +0.25 K. Mixed-layer values of any parameter alpha, noted alphaML is then simply computed as the average of alpha over the layer [0,z(k)]. In addition, we also make use of lower -levels values, defined for any alpha, as the average of alpha over the layer [0,850m], and noted alphaLOW.

CAPE, CIN, LCL, LFC and LNB have been computed for parcels located approximately 250m AGL. We also tested alternative formulations, such as retaining the level kc within the first km for which CAPE is maximum, and then defining CAPE and CIN as CAPE(kc) and CIN(kc). Small differences are found between the two estimates. They do not however modify our conclusions.

FE03a defined two indexes: a convective triggering index (CTP) and a low-level humidity index. We chose to keep the same definitions as FE03a, even though other or simpler indexes could perhaps be defined for our purpose, so as to allow for a direct comparison of our results, over West Africa. The CTP corresponds to the pseudo-adiabatic CAPE of a parcel lifted at saturation 100 mb AGL and stopped 300 mb AGL. The CTP is correlates well with the lapse-rate Γ of the atmosphere over this low layer: the smaller the lapse-rate, i.e. the more convectively unstable the layer, the larger the CTP, and also the easier and faster the daytime BL growth. The low-level humidity index is simply an average of dewpoint depression at 50 mb and 150 mb AGL, it is therefore an indicator of the dryness of the lower levels, and is closely linked to the LCL (Betts 1997). The combination of these two indexes provides a framework allowing to distinguish between early morning soundings that are more or less favourable for daytime moist-convective development over wet versus dry surfaces, or over both.

Indexes such as proposed by FE03a provide a characterization of the land-atmosphere coupled system in a way which is to some extend similar to CAPE when considered like a convection-related characterization of the atmospheric column. CAPE alone is far from accounting for all convective features. It is more an indicator of convective strength, for instance CIN also exert a strong control on convective activity, even over tropical oceans (Parsons et al. XXXX). Similarly FE03 indexes cannot by themselves account for differences in surface net radiation, nor on daytime BL cloud-surface radiative feedbacks. They do not either address the potential influence of daytime mesoscale circulations in an explicit way, although such circulations are expected to be sensitive to the low-level lapse-rate, and therefore similarly to CTP (e.g.; Wang et al. 1998).

3) Results

Figure 2 enlightens the large contrasts between continental-type temperature profiles over West Africa; it also show how they depart from the Pacific Warm Pool profiles (TOGA-COARE data, Cielsielski et al. 2003). A strong meridional gradient of lapse-rate is found below the top of the SAL, around 500mb. In early Summer, the stratification remains weak within a dry Sahelian atmosphere (in term of relative humidity). This points to the major role of dry convection in mixing the low levels over this continental dry region in Spring and early Summer (e.g.; Gamo 1996). In Niamey, compared to the Warm Pool region, the atmosphere is warmer up to about 650 mb by a few degrees and colder above. It is associated with a higher lapse-rate up to the top of the Saharan Air Layer (SAL), located around 600-550 mb. Even on a 4-month average, the top of the SAL is fairly well marked, it is the most pronounced in June while it becomes more diffuse and somewhat lower in August. The atmosphere is also drier by a few g.kg-1 in Niamey, the difference is the more pronounced below 900 mb and concern the whole column except the [900mb,800mb] layer where both profile-types become quite close. Relative humidity is however much lower except at top of the SAL (see Appendix B). The atmosphere above Parakou in August and above the warm pool are relatively close, it is moister in Parakou, the vertical gradients of θe are also less pronounced. This suggests a higher intensity of precipitating convection, which is consistent with differences in rainfall rates reported for these two distinct areas.

On the other hand, prior to the monsoon onset, surface heat fluxes over the Sahel are typically high, because the surface is mostly dry and characterized by small latent heat fluxes (Fig. 6). The situation reverses during the core monsoon: sensible heat fluxes decrease significantly with values typically closer to those measured in the Soudanian zone throughout the Summer (Timouk et al. 2009, Ramier et al. 2009). Therefore, from a climatological perspective, over West Africa, drier (wetter) surfaces are associated with higher (lower) atmospheric lapse-rate and drier (moister) low levels.

Despite strong meridional differences in temperature and moisture, High CAPE can be found from the Guinean to the Sahelian zones (see 'mixing of latitudes/colours' in Fig. 7). Furthermore, CAPE appears to be strongly linked to low-level θe across the whole West African region. Above approximately 345K, CAPE increases by about 200J/kg per K of θe. This finding is consistent with previous studies over different regions, e.g. Williams and Reno (1993). It shows and quantifies how the low levels represent the dominant driver framing atmospheric stability and convective intensity well beyond their diurnal fluctuations.

In contrast, meridional differences in the thermodynamics of the low levels appear clearly on the scatter diagram relating the two FE03 indexes (Fig. 8). This figure also suggest a significance of cases where surface properties matter for the development of convection (cf Fig. 4, top). In the northern Sahelian and Saharan latitudes, convection appears more often limited by the dryness of the atmosphere, while the few cases for which CTP is negative are confined to the more rainy southern locations. It is interesting to notice that the quasi-equilibrium concept developed by Arakawa and Schubert (1974) involves a strong coupling between temperature lapse-rate and boundary layer humidity, i.e. parameters closely linked to FE03 indexes (e.g. see Fig. 7 in Arakawa 2004). Namely, the quasi-equilibrium state would roughly, schematically, coincides with a regression line of the scatter of points shown in Fig. 8 (after exclusion of 'dry points' for which the concept of moist convection becomes irrelevant).

In more details, FE03 indexes suggest that surface-atmosphere feedbacks are actually significantly operating over West-Africa, but that the feedback loops change sign with latitude and season (Fig. 9). For instance in Niamey, negative feedback cases are more numerous outside of the core monsoon season (August) while most positive feedback cases are found in August (blue dots). This finding is consistent with the observation that daytime convection is not always favoured over wet surfaces in the Sahel (Taylor and Ellis 2006). The variability of the low levels is also stronger in the Sahelian zone than in the Soudanian zone, from daily up to seasonal scale. In the Sahel, the large day-to-day variability appears to be organized at synoptic scale in the form of wave-like patterns which are linked to African Easterly waves according to the ECMWF analysis (see appendix C).

FE03 indexes are computed at 6Z, i.e., they do not explicitly account for the actual diurnal cycle of the low levels, but infer some of its basic features from early morning atmospheric stability, soil moisture and simple daytime BL dynamics arguments. Variations along the season and with latitude of LCL and LFC are indeed consistent with the previous analysis of FE03 indexes over West Africa (Fig. 10). For instance, in Niamey, a strong diurnal cycle of LCL is found, especially early in the season, while the LFC does not fluctuates much with the hour in the day, but shifts downwards from June to August. In contrast, the LCL displays a much weaker diurnal cycle at Parakou, while the LFC now does, it is the lowest at night and the highest around noon. It would be valuable to get estimations of the respective contributions of surface fluxes versus radiation, turbulence and advection to this result, i.e. to conduct a budget study.

Note however that the LCL can be located *above* the mixed layer top. In particular, the LCL cannot be interpreted as an indicator of the mixed layer top when the low levels are cloud free. This is particularly true early in the season in the Sahel, on days when none of the daytime boundary layer convective plumes reach their LCL. In that case, moist convective development is limited by high convective inhibition (CIN) rather than CAPE. Indeed, no clear seasonal trend of CAPE is found in Niamey, while, interestingly, it actually decreases in Parakou during the core of the monsoon season (Fig. 11). Thus, simple one-dimensional CAPE considerations are consistent with convective strength (e.g.; convective vertical velocity) being at their lowest in August compared to June and September in the Soudanian area. Consistently with the differences found for the LFC in Niamey and Parakou, the diurnal cycle of CAPE is less pronounced in Niamey than in Parakou, with an evening to early-night maximum there.

4) Conclusion

The study presented above provides a quantification from the AMMA-SOP high-resolution radiosounding dataset of the very large geographical and seasonal variations of the atmospheric boundary layer and convection-related indexes over West Africa. (An evaluation of its counterpart in the ECMWF analysis has also been carried out; see Appendix C).

Firstly, this study demonstrates the major control of the low levels on convective instability at regional scale. This notably appears on the strong correlation linking CAPE and low-level equivalent potential temperature. Above approximately 345K, CAPE increases by about 200J/kg per K (below 345K, CAPE remains close to 0). Over the region, θe is found to vary from 320 K in the Saharan heat low to more than 360K during the monsoon (JJAS).

Secondly, by applying the indexes and framework proposed by FE03 to early morning soundings, this study help drawing a map of areas where feedbacks between the surface and convective development are likely to operate. This analysis is complemented by an analysis of measured surface heat fluxes and of the diurnal cycle of boundary layer properties and convective indexes.

These basic considerations of the thermodynamic environment and surface properties point to the likeliness of a variety of modes of land surface-convection interactions over West Africa. More precisely, they suggest that positive *and* negative feedbacks between soil moisture and deep convection may frequently operate during daytime hours, as opposed to an absence of coupling. Thus, daytime convection may be either suppressed or favoured over wet surface versus dry ones, depending on latitude and time in the season at large scale, as well as on synoptic fluctuations (african easterly waves). At large scale, the Sahel appears as an area where negative feedbacks are significant prior to the monsoon onset and during the dry down period. Positive feedbacks become more likely during the core monsoon season while they dominate in the Soudanian zone. In the Guinean zone, daytime convection appears less sensitive to surface properties.

A few observational studies have show that daytime convection can be favoured over dry surfaces (Findell & Eltahir 2003), and in particular over the Sahel (Taylor & Ellis 2006, see also Gounou et al. 2009). Our analysis suggests that a positive feedback is more likely to operate under moister, cloudier conditions. Thus, positive feedbacks may be more difficult to identify from satellite data such as land surface temperature estimates from IRT.

The present study does not answer the question of the smaller scale down to which the thermodynamic environment play a role, and how, nor how this environment combines with other factors, such as turbulence and advection, cloud-induced surface radiative impact and mesoscale circulations to explain observations.
Further investigation with data collected within AMMA should help answering these questions. Process-based analysis of well documented case studies, complemented by high-resolution modelling, are valuable approaches to address these issues.

Such diagnostics could also be valuable tools for analysing land-surface convection couplings in models. It would also be useful to map the other continental regions such as South America, China and Australia to cite a few, in order to get a better view of how distinct they are. This could help to precise their climate sensitivities.

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Table 1: information on sounding sites

name of sounding site	country	longitude	latitude	altitude (m)
Abuja (*)	Nigeria	7.16°E	9.00°N	370
Agadez(*)	Niger	7.99°E	16.97°N	501
Bamako	Mali	7.95°W	12.53°N	377
Bangui	Centrafrique	18.52°E	4.40°N	395
Cotonou(*)	Benin	2.40°E	6.52°N	5
Dakar	Senegal	17.48°W	14.74°N	28
Douala	Cameroun	9.73°E	4.00°N	5
NDjamena	Tchad	15.03°E	12.13°N	295
Niamey (*)	Niger	2.17°E	13.48°N	222
Ngaoundere	Cameroun	13.56°E	7.36°N	1104
Nouakchott	Mauritanie	15.95°W	18.10°N	3
Ouagadougou	Burkina Faso	1.00°W	12.53°N	310
Parakou (*)	Benin	2.61°E	9.36°N	392
Tamale (*)	Ghana	0.85°W	9.55°N	168
Tambacounda	Senegal	13.66°W	13.73°N	49
Tombouctou	Mali	3.00°W	16.72°N	263

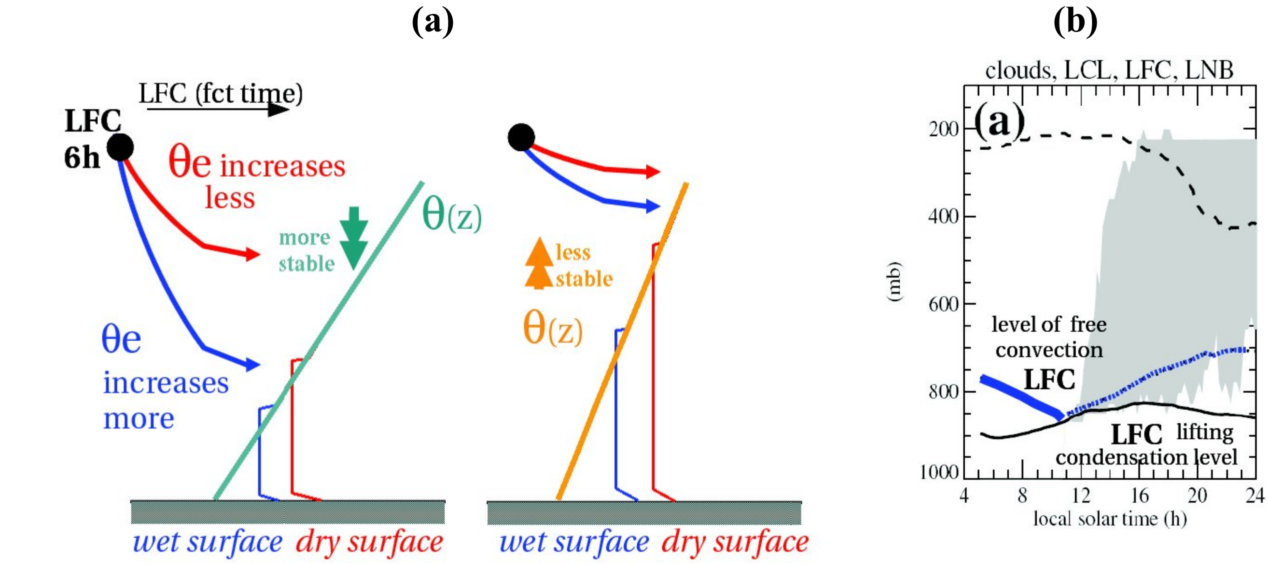


Figure 1 : (a) schematic of daytime evolution over wet (blue) versus dry (red) soil of mixed layer potential temperature and level of free convection (LFC) under more stable (left) versus less stable condition - more stable conditions correspond to a lower lapse rate, i.e. $\partial T/\partial z$ decreases less with height. (b) is adapted from Guichard et al. (2004) and illustrates a case of development of daytime convection over a wet area under relatively stable conditions : the LFC decreases in the morning by about 100 mb while the boundary layer (which here is well traced by the lifting condensation level) grows.

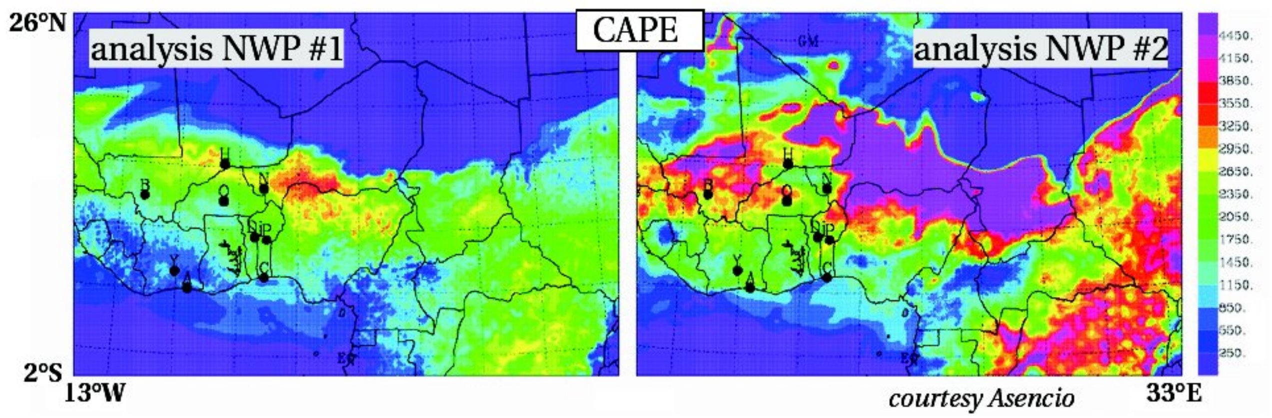


Figure 2 : map of CAPE (18 August 2005, 1200 UTC) derived from the ECMWF (left) and ARPEGE (right) analyses - CAPE was computed from fields interpolated on a common grid for parcels originating from the first vertical level, located about 30 m above the surface.

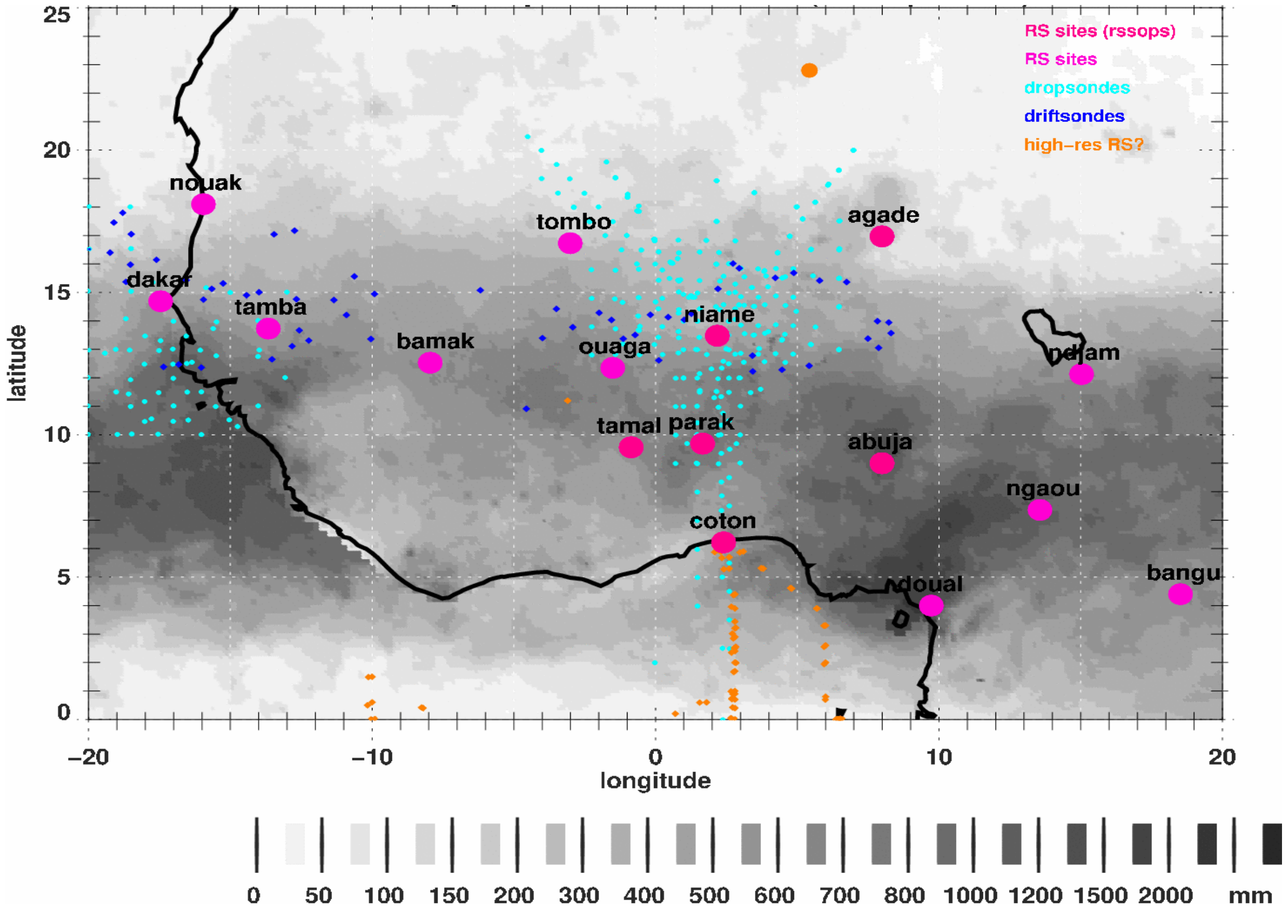


Figure 3: Map showing the location of high-resolution sounding sites (large disks), dropsondes (cyan dots), driftsondes (blue dots) and vessel soundings (orange dots except the orange dot over land which corresponds to Dano soundings); the red disks correspond to sounding sites where soundings were launched with a high temporal frequency. Figure precipitation August 2006 (CPC RFE2). These are shown on top of the August 2006 cumulative rainfall (CPC-RFE2 estimate).

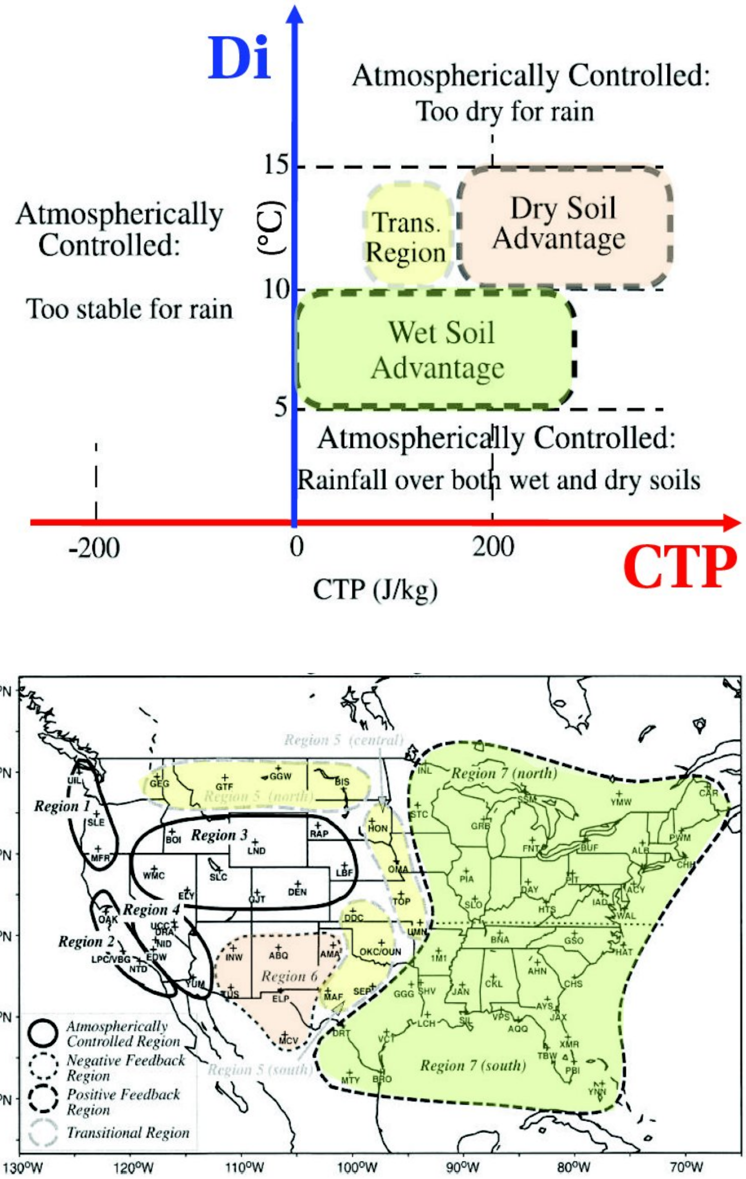


Figure 4, from Findell and Eltahir (2003), slightly adapted: (a) presents the framework proposed by FE03 to diagnose a likeliness of daytime convective initiation from morning soundings; it consists in a diagram using values of CTP and Di indexes for discriminating between favourable versus unfavourable soil and atmospheric conditions. (b) displays the mapping of the conterminious US derived from this framework from Summer time local morning soundings.

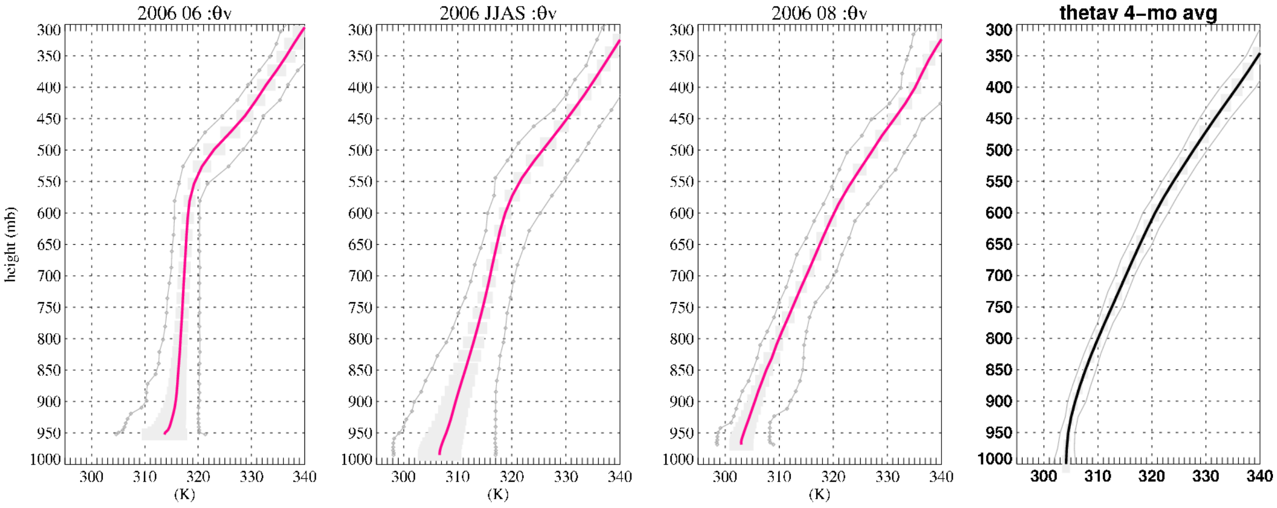


Figure 5 : Gridded vertical profiles of virtual potential temperature (θv) for (a) Agadez in June 2006, (b) Niamey in Jun-Jul-Aug-Sep 2006, (c) Parakou in August 2005, and (d) Nov-Dec 1992-Jan-Feb 1993 over the Pacific warm pool (TOGA-COARE dataset). The time-average is drawn in pink (resp. black) in (a), (b) and (c) (resp. (d)); the grey shading indicates +/- one standard deviation and the grey lines delineate the minimum and maximum values over the period. Time-average values has been computed from the four daily average of 6-hourly sampled values, in order to prevent spurious signatures from unevenly sampled diurnal cycles.

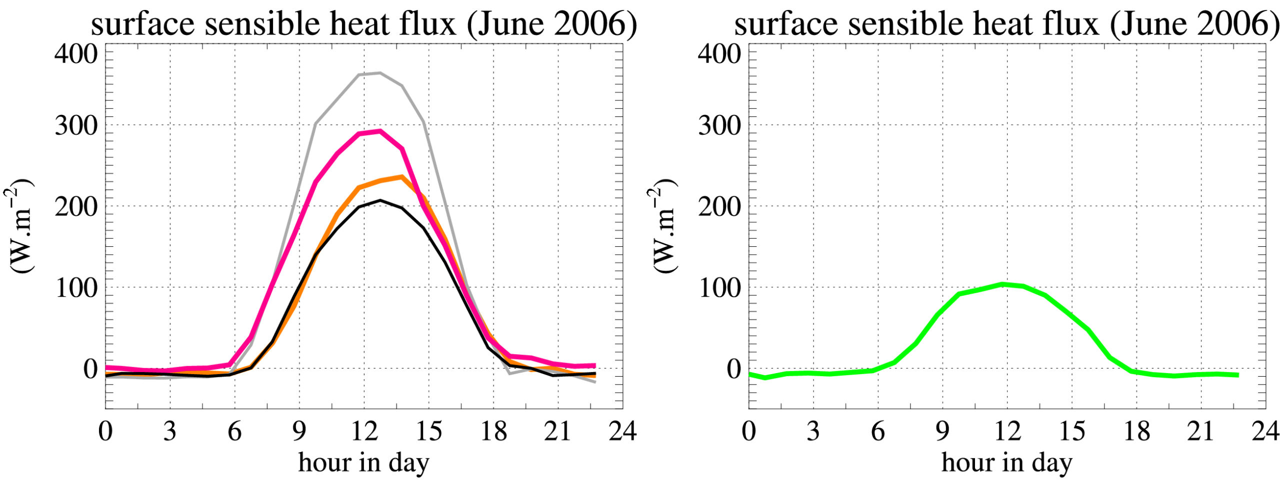


Figure 6 : Diurnal composite of surface sensible heat flux (H) for 20-30 June 2006 (a) over different Sahelian sites (black: Bamba 17°N bright sandy dune, orange: Banizombou, red: Eguerit 15°N bare soil, grey: Kelma 15°N accacia forest) and in the soudanian zone (green: Nalohou 10°N fallow).

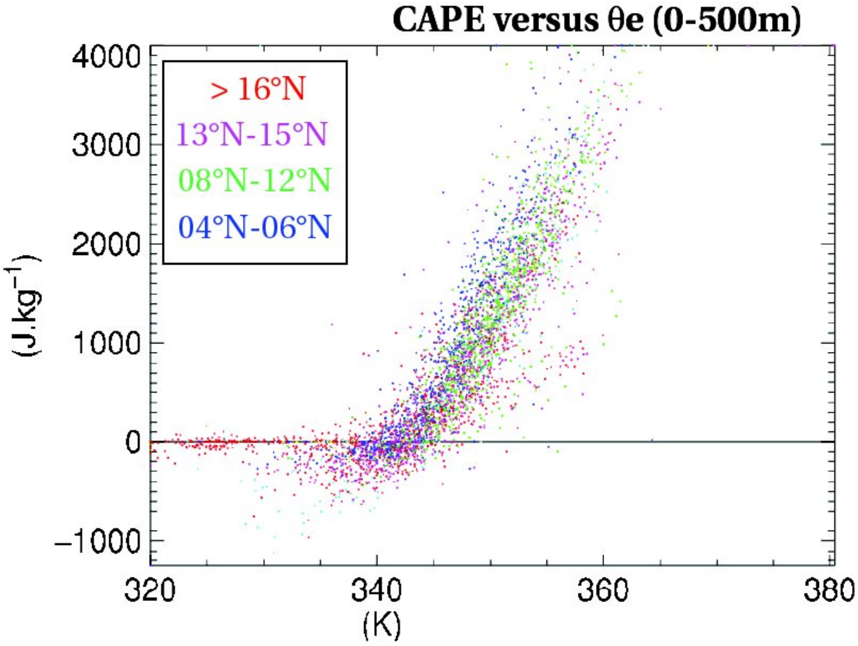


Figure 7 : Scatter diagram of CAPE versus low-level θe, colours discriminate between latitude bands (prepared with more than 4300 soundings over the period June-July-August-September 2006).

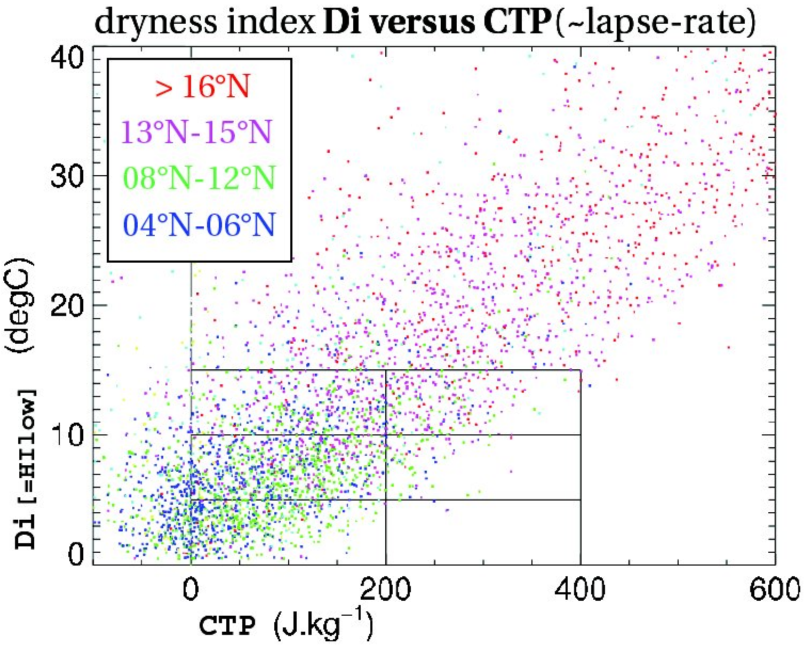


Figure 8 : Same as previous figure except for dryness index Di versus CTP.

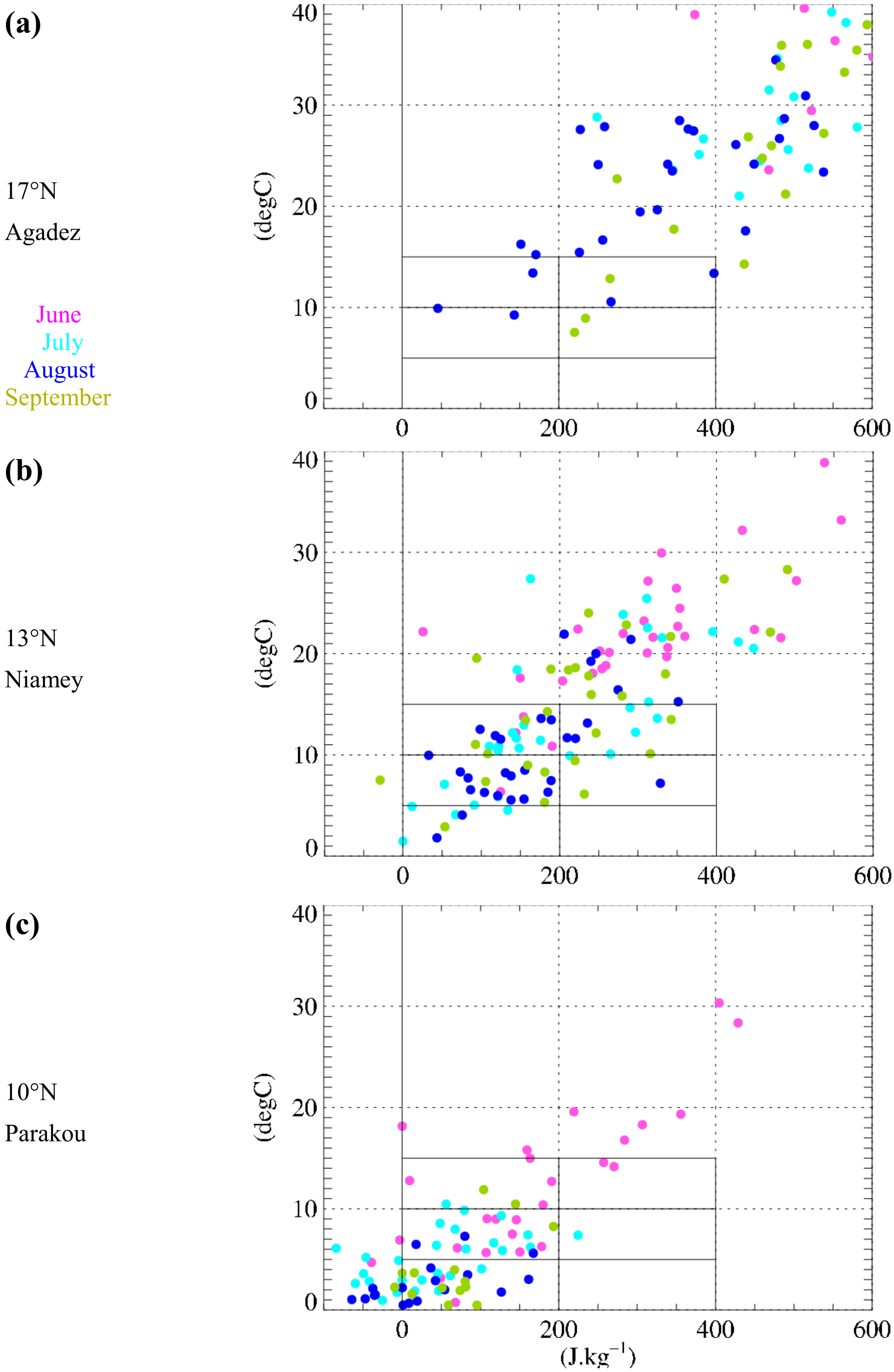
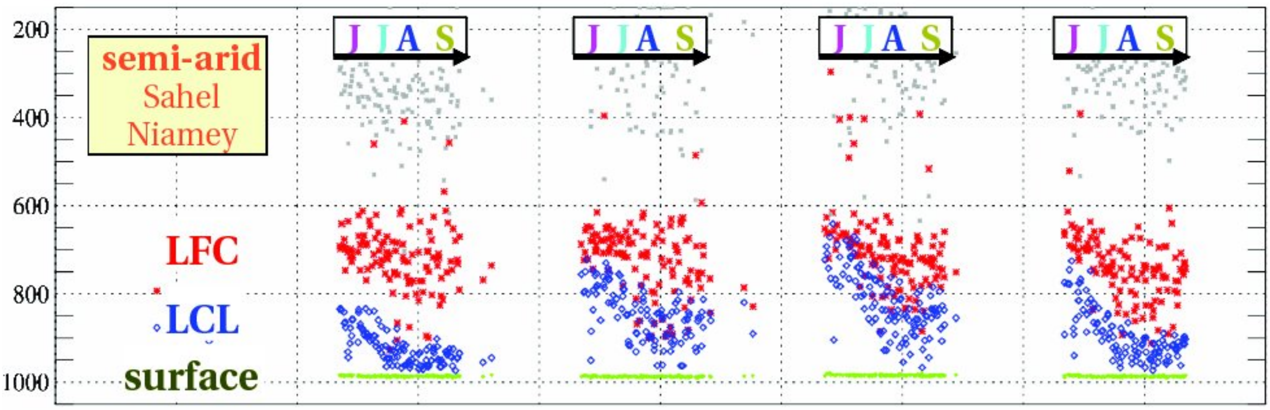


Figure 9 : Scatter diagrams of dryness index versus CTP in (a) Agadez, (b) Niamey and (c) Parakou in Jun-July-August-Sept 2006, computed from 0600 UTC soundings, colours discriminate between months.

(a)



(b)

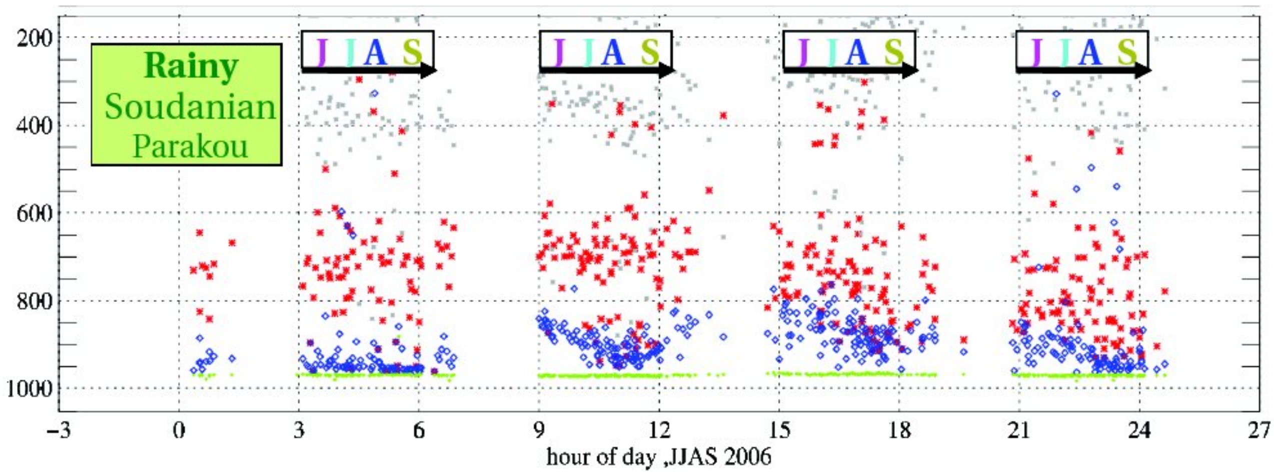
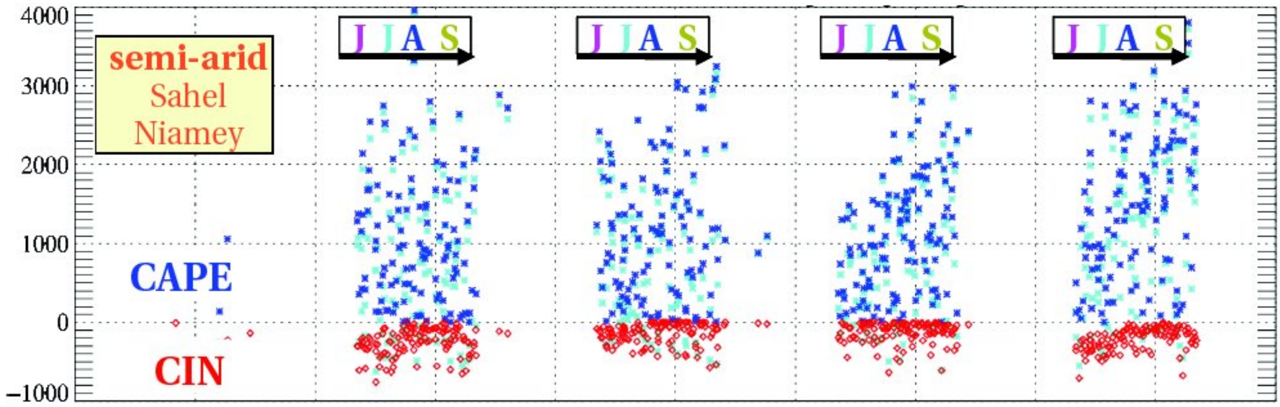


Figure 10 : diurnal variations of the heights of the lifting condensation level (blue) and level of free convection (red) thoughout JJAS in Niamey and Parakou; surface pressure is indicated by the green dots. Values were computed for soundings available around the four synoptic hours (0000, 0600, 1200 and 1800 UTC). For each synoptic hour, a monotonic increment function has been added to the hour of day in order to distinguish its seasonal evolution.

(a)



(b)

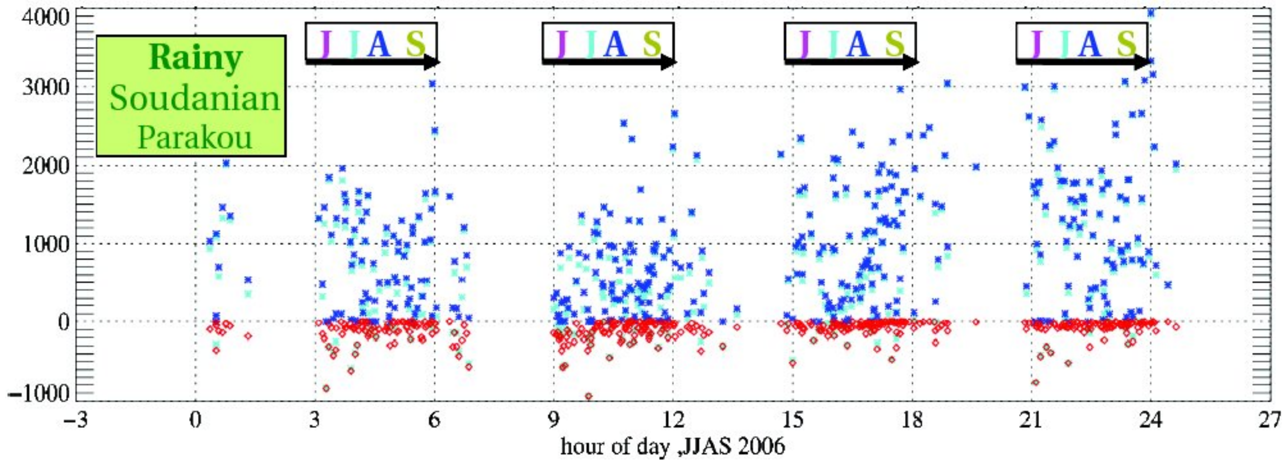


Figure 11 : Same as previous figure except for CAPE (blue) and CIN (red).

APPENDIX A : documentation of the dry bias of RS80 sondes

Figure A1 highlights the dry bias which is affecting measurement of water vapour with RS80 sondes; this issue is discussed in more details by Nuret et al. (2008), hereafter N08. RS80 soundings were used at different sites during AMMA, notably at Bamako.

The correction of humidity proposed in N08 provides a correction which enhances relative and specific humidity in the low levels. This results in higher low-level qe values which drives an increase of CAPE (Fig. A2). Version 5 of the correction is used in this figure. It only slightly differs from the version proposed in N08. Further details on the impact of the correction is presented in the form of scatter diagrams in Figs A3, A4 and A5, for specific humidity, equivalent potential temperature, lifting condensation level, level of free convection, CAPE and CIN.

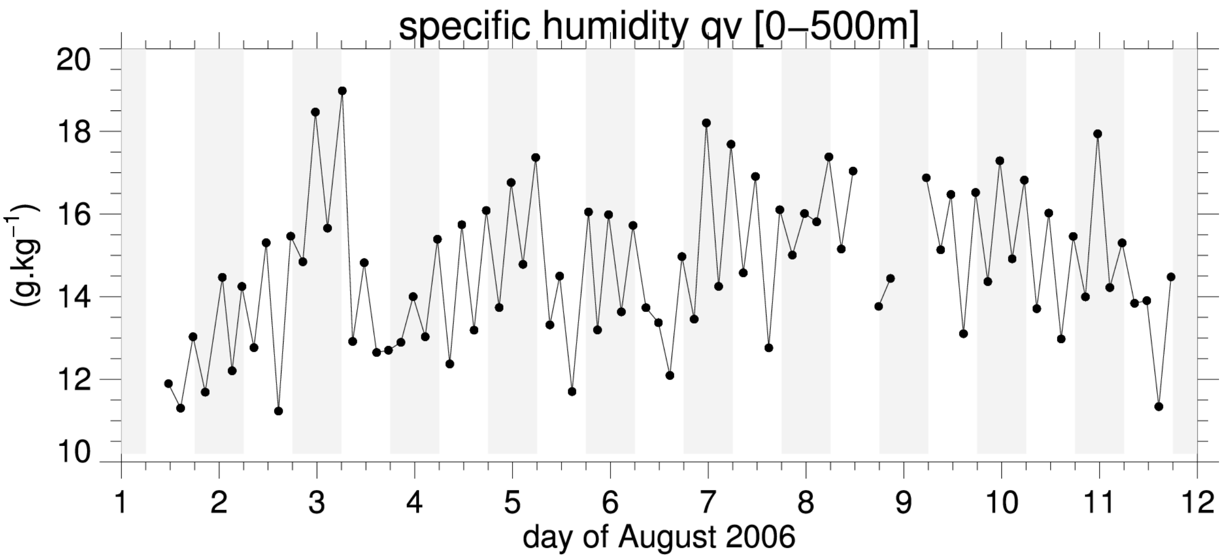


Figure A1: Time series of low-level (0-500m AGL) specific humidity in Niamey during the August sounding IOP. The seesaw which appears on this figure is non-physical and due to the alternate launch of RS80 and RS92 Vaisala soundings (correction version 3).

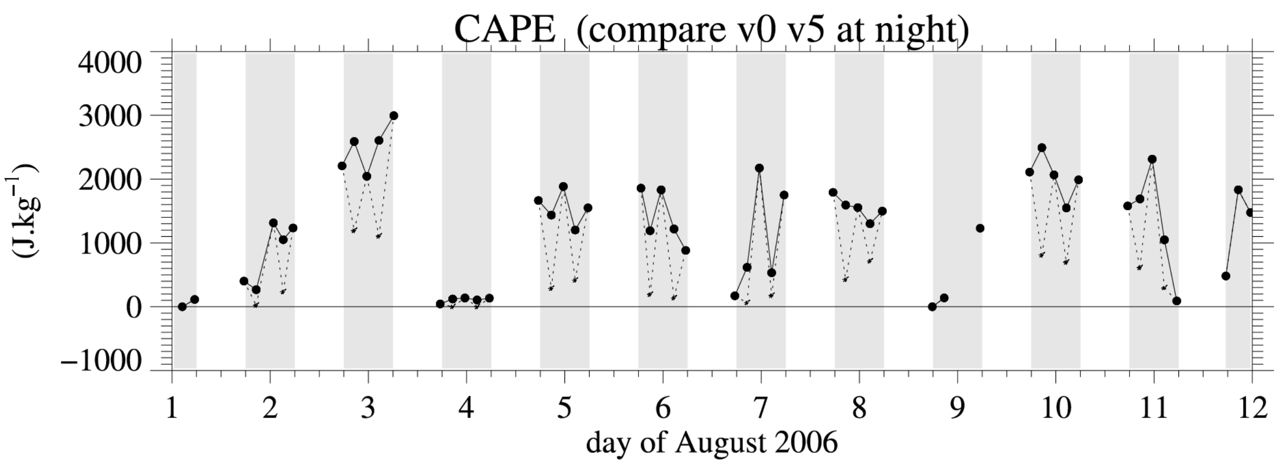


Figure A2: Time series of night-time CAPE from Niamey soundings for the same period as previous figure; dotted line for the mixed raw RS80/RS92 dataset and solid line with the corrected RS80 humidity dataset (version 5, provided by M. Nuret).

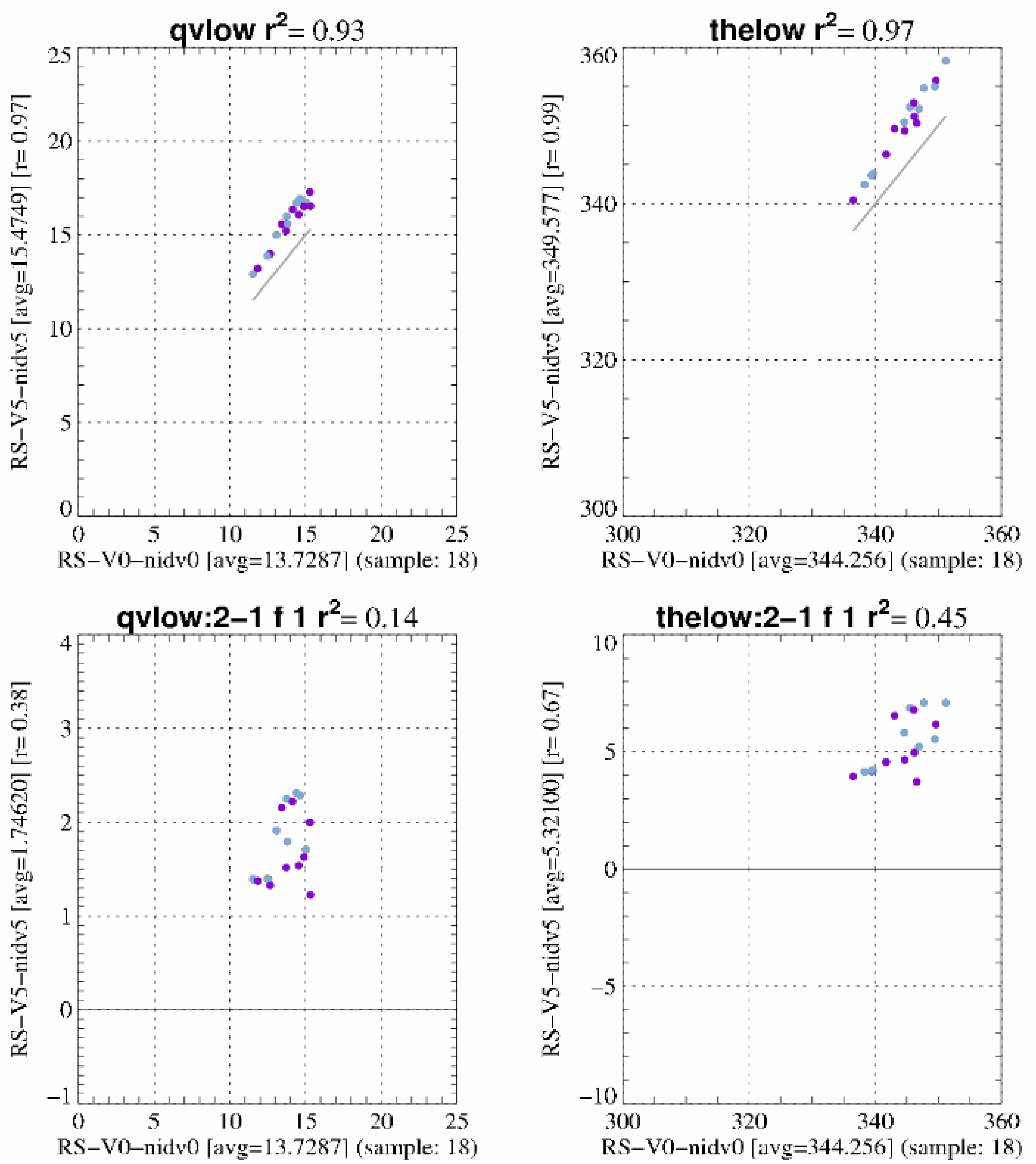


Figure A3: Scatter diagrams of low-level (0-500m) specific humidity qv (top left) and equivalent potential temperature θe (top right) computed from Niamey night-time soundings, from raw RS80 data on x-axis, from corrected data (version 5) on y-axis. Bottom: same as top except difference corrected-raw on t-axis. Colours stand for values around midnight (blue-grey symbol) and around 3h (purple symbol). The corrected RS80 humidity dataset (version 5) has been provided by M. Nuret, only nighttime soundings from the August sounding IOP in Niamey are considered here.

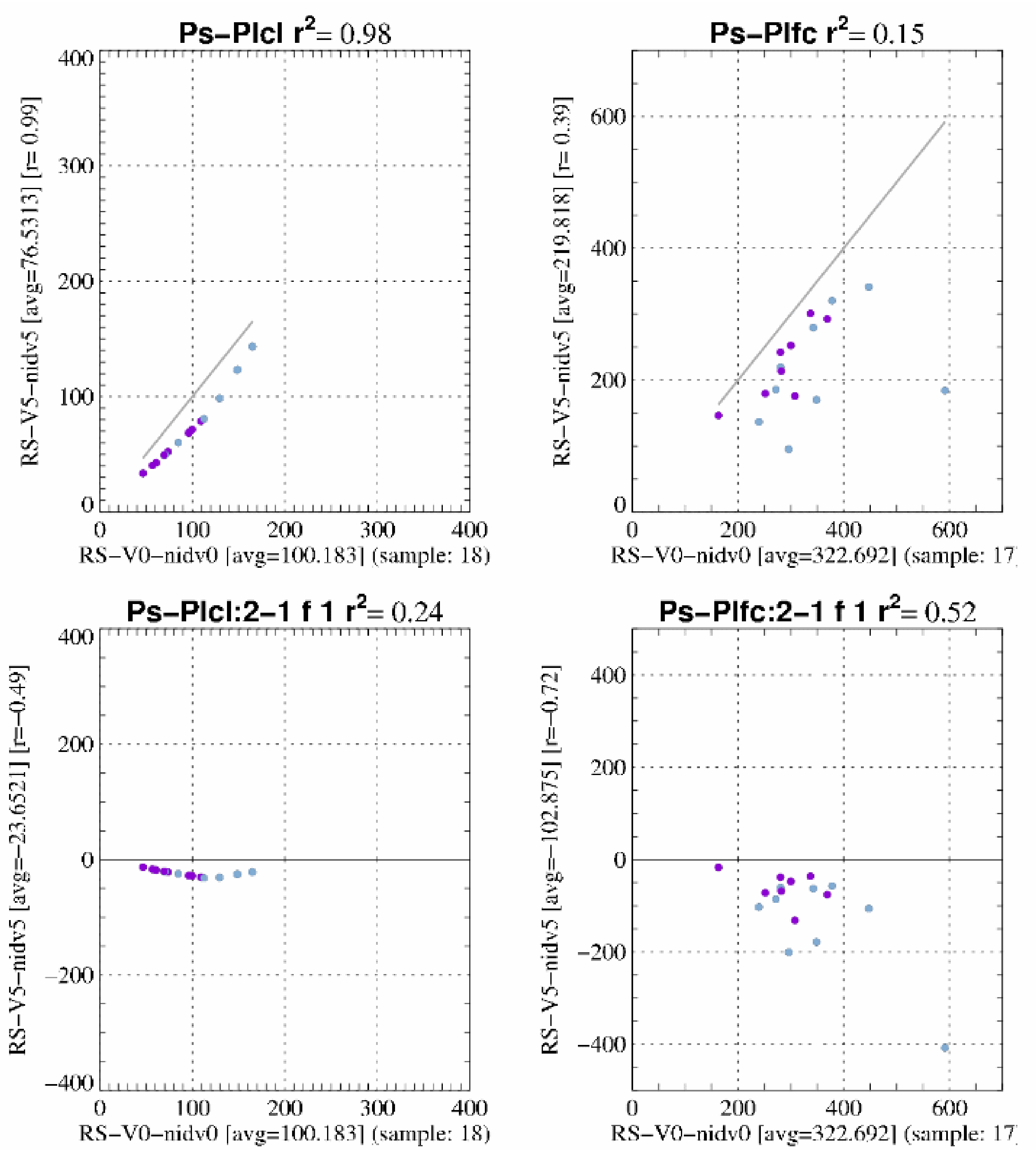


Figure A4: same as previous figure except for the lifting condensation level, Plcl (left) and level of free convection, Plfc (right), both are expressed as a difference between suface pressure and the pressure-height of the level.

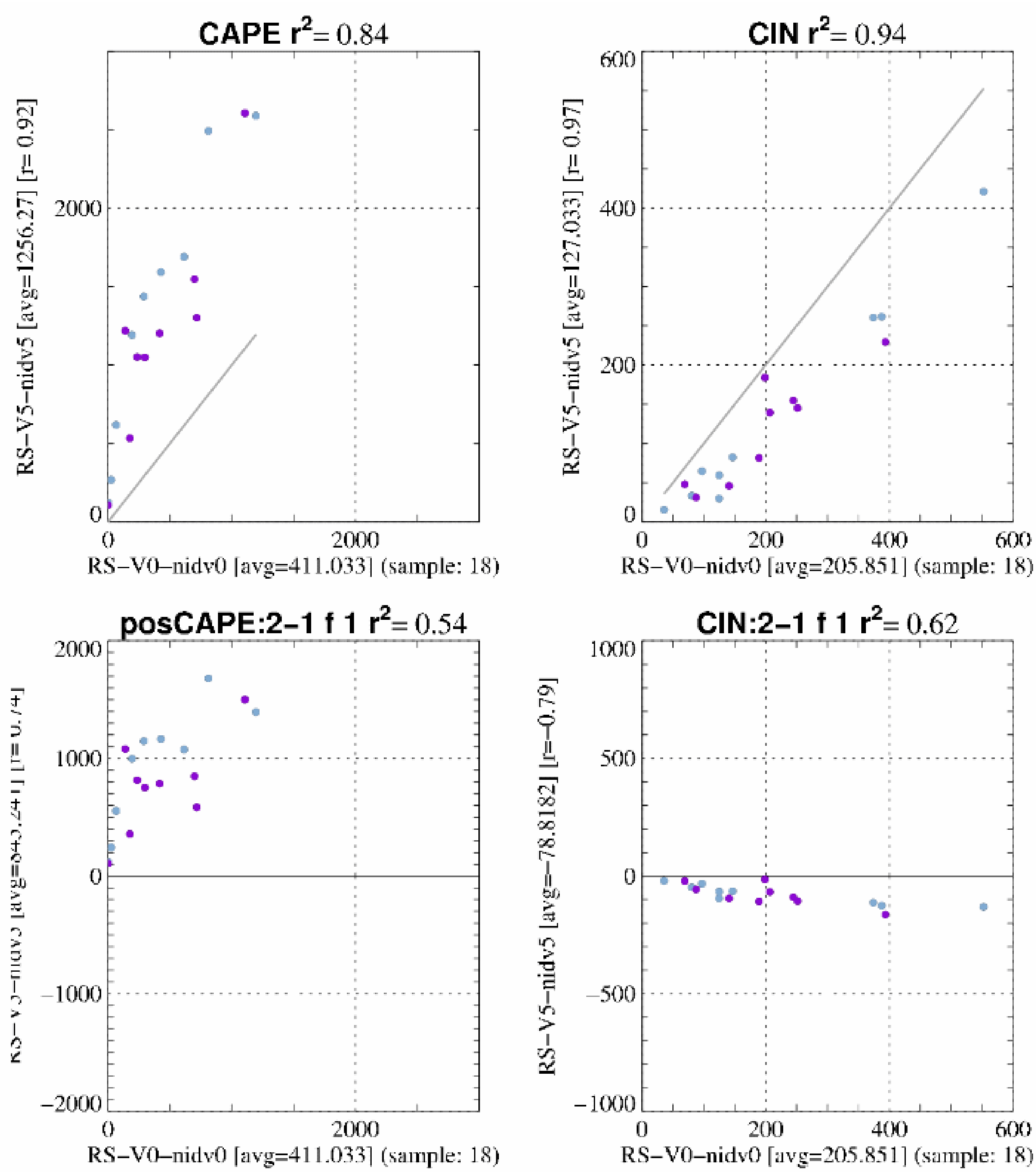


Figure A5: same as previous figure except for CAPE (left) and CIN (right).

APPENDIX B : Vertical profiles from soundings

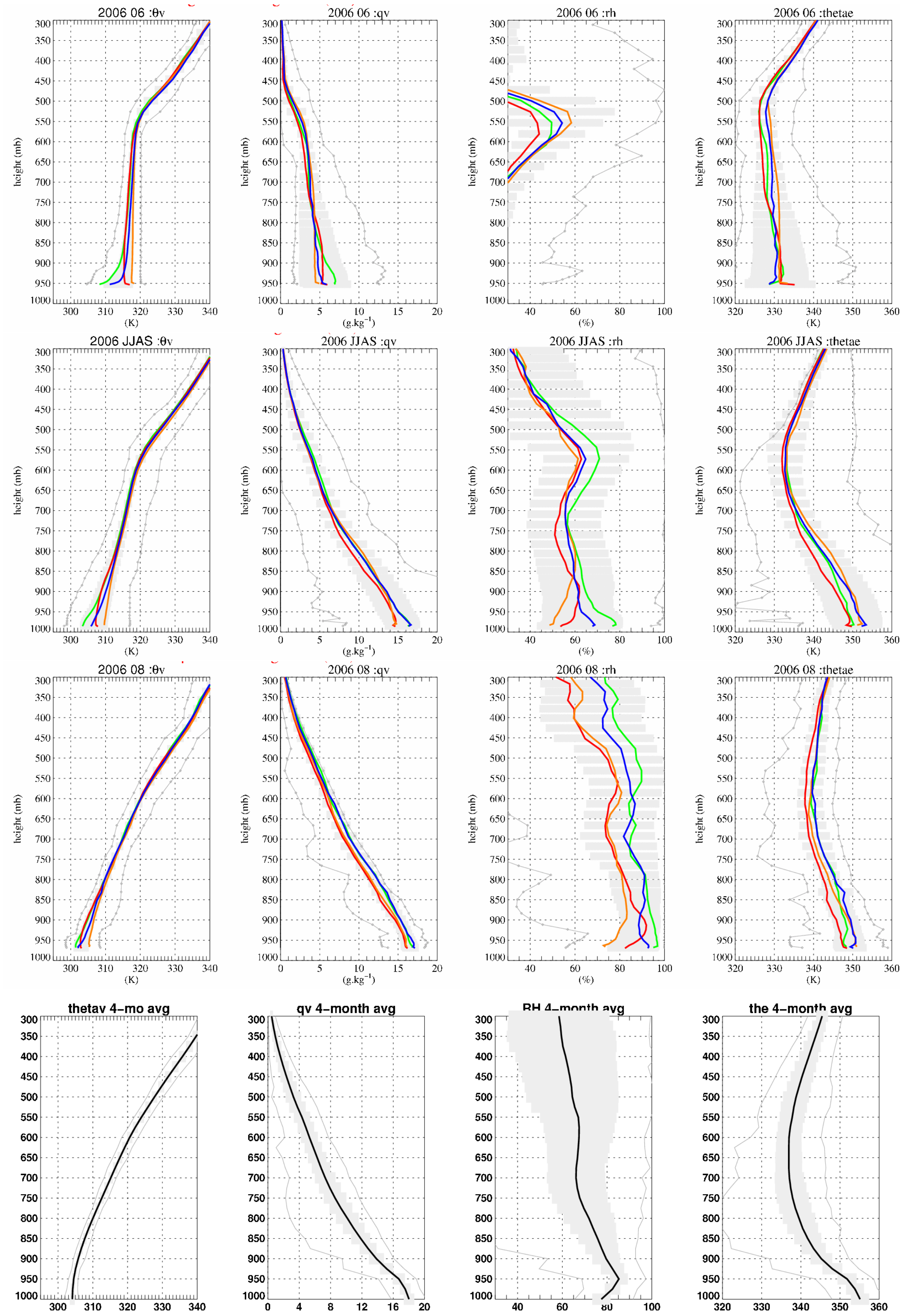


Figure B1 : Gridded vertical profiles of virtual potential temperature (θv), specific humidity, relative humidity and equivalent temperature for Agadez in June 2006 (first raw), Niamey in Jun-Jul-Aug-Sep 2006 (second raw), Parakou in August 2006 (third raw), and Nov-Dec 1992-Jan-Feb 1993 over the Pacific warm pool (TOGA-COARE dataset) (fourth raw). The colours corresponds to different hours in the day (around 0h in blue, 6h in green, 12h in red and 18h in orange); the grey shading indicates +/- one standard deviation and the grey lines delineate the minimum and maximum values over the period.

APPENDIX C : Indexes in the ECMWF-IFS analysis: evaluation and mapping

Vertical profiles from the ECMWF_IFS analysis (on the model grid) have been used for the computations presented below.

The same FE03 indexes computed from the ECWMF analysis indicate similar seasonal and latitudinal trends (Fig. C1). Some quantitative differences persist however despite the assimilation of most of these observations, and the analysis-derived indexes exhibit less variability than given by sounding data. Intraseasonal and day-to-day fluctuations are also reasonably reproduced (Fig. C2).

At larger scale, beyond a strong meridional gradient (Fig. C3 top), the day to day variability of both CTP and D is due to a significant synoptic variability, associated with African easterly wave patterns, as shown by daily maps of these indexes computed from the ECMWF analysis (e.g. 11 June 2006, Fig. C3 bottom).
It would be valuable to document the balance of processes leading to this signature.

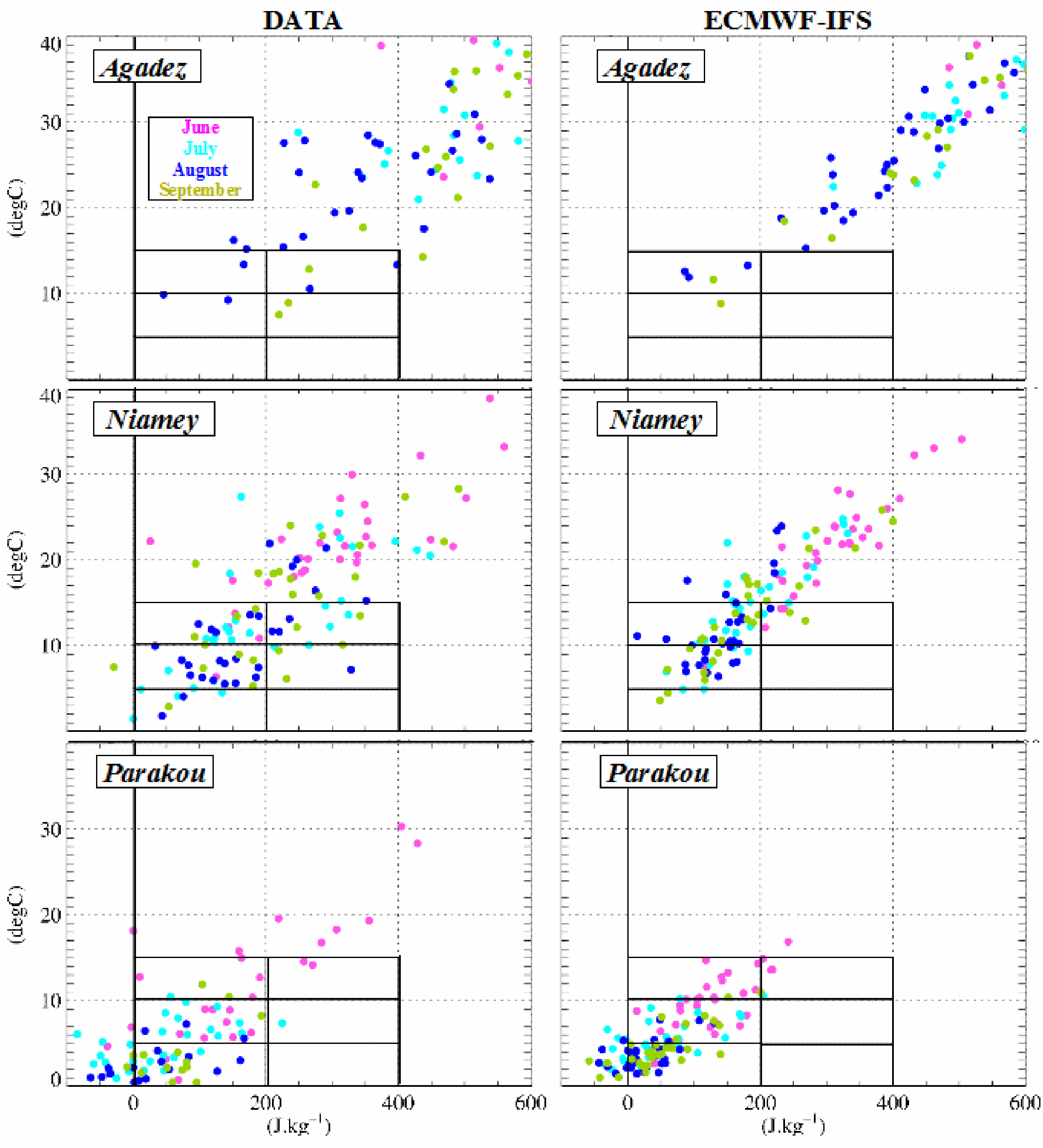


Figure C1 : Scatter diagrams of dryness index versus CTP from soundings (left) and ECMWF analysis (right) in Agadez, Niamey and Parakou in Jun-July-August-Sept 2006, computed from 0600 UTC soundings, colours discriminate between months.

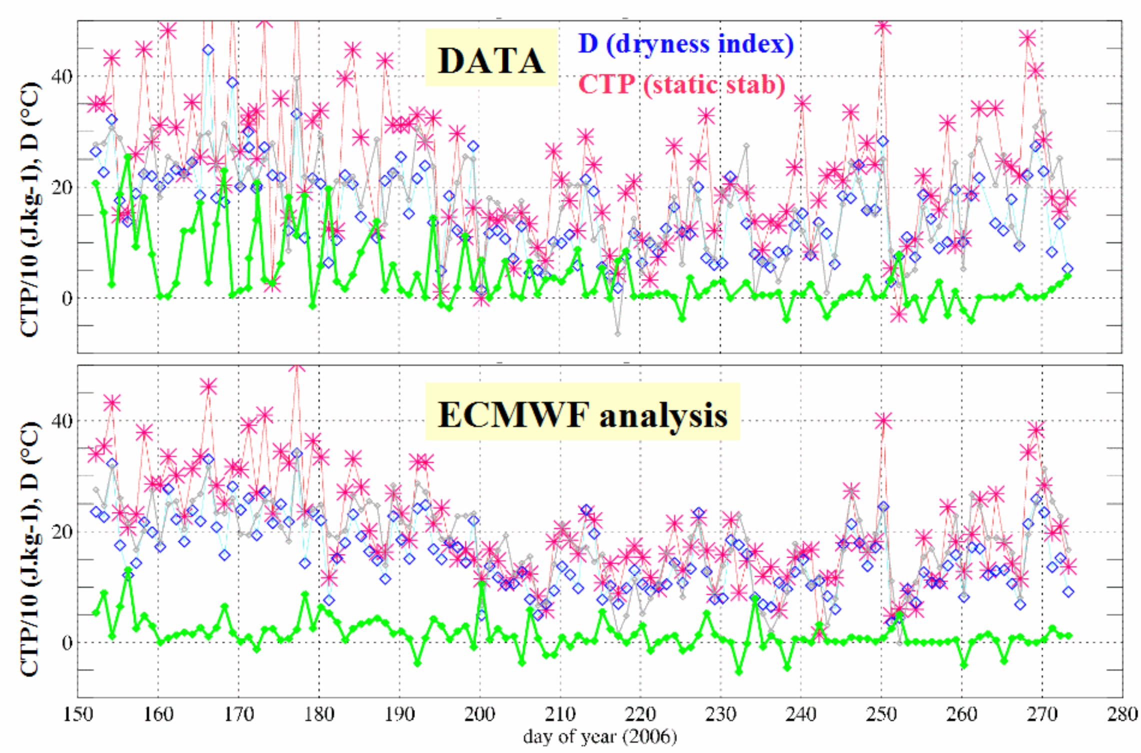


Figure C2 : Time series of dryness index (blue) and CTP at Niamey, from soundings (top) and ECMWF analysis (bottom).

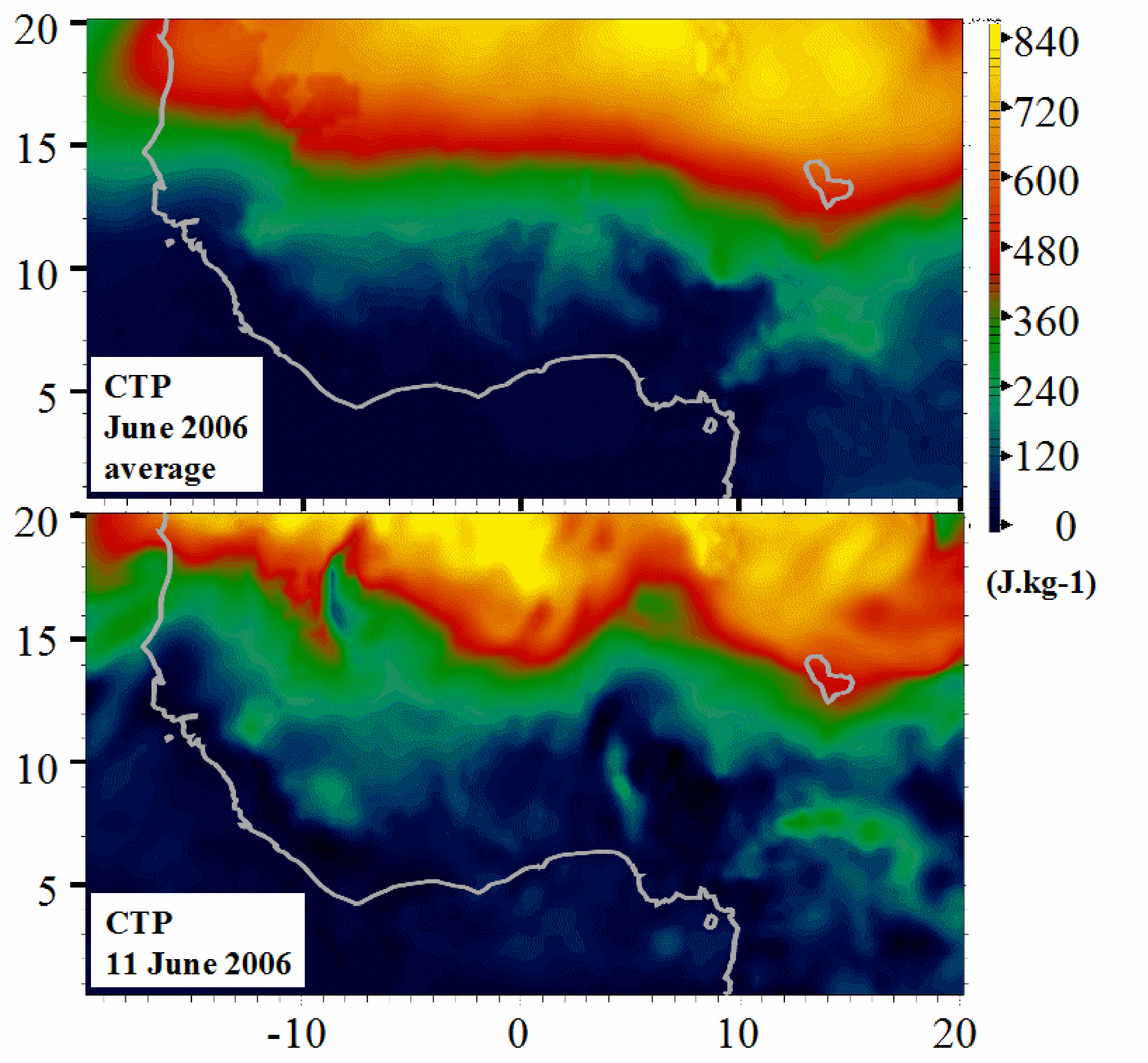


Figure C3 : Maps of CTP computed with ECMWF analysis, June average (top) and 11 June 2006 (bottom).