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A POSSIBLE MECHANISM REGULATING NOCTURNAL STRATOCUMULUS DECKS IN WEST AFRICA

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1. Introduction

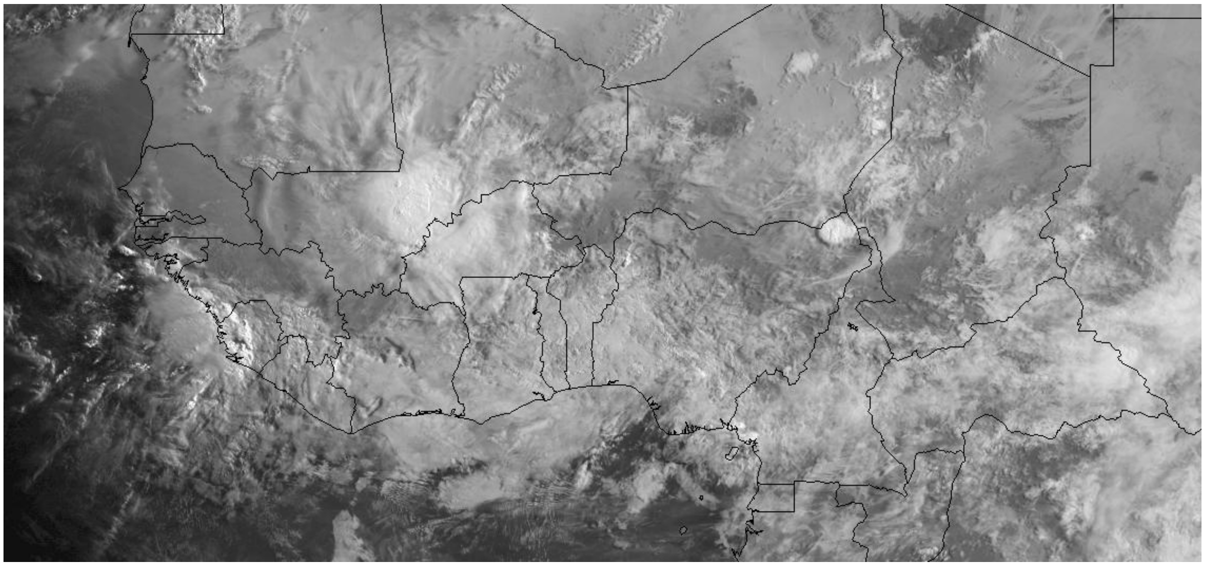
Often at night during the summer monsoon, the Soudanian (9°N-12°N) and Guinea Coast (south of 9°N) climate zones of West Africa are blanketed in a shallow layer of clouds that are stratocumulus. These low, shallow cloud layers are extremely difficult to detect in an analysis of infrared pictures due to their relatively warm temperatures, but characteristically they persist for a few hours after sunrise, making it possible to identify these "nocturnal" structures with visible satellite imagery.

While these stratocumulus decks seem to be quite common, they are by no means ubiquitous. Often it is the case that large regions of tropical West Africa exhibit extremely clear conditions at low levels at night during the monsoon. Synoptic reports typically contain codes for cirrus clouds on such nights; actual reports of "clear" conditions are extremely rare for stations in these moist zones during the rainy season. (See Fig. 1 for an example.)

Schrage et al. (2007) used a combination of *in situ*, satellite and ECMWF operational analysis data from the summer monsoon season in West Africa in 2002 to compare the structure of the atmosphere on nights in which this stratiform cloud layer formed at a relatively early hour (namely, prior to 0000 UTC) with that observed on nights in which the same region remained relatively cloud-free. That study focused on the radiosonde station at Parakou, Benin (9.35° N, 2.62° E) as this was the only operational upper-air station in the Soudanian climate zone at that time. Composite temperature and moisture profiles revealed key structures in the lower troposphere on these nights. Clear nights were considerably drier than cloudy nights, with much higher static stability observed in approximately the lowest 100 m of the boundary layer. In fact, on some cloudy nights, the near-surface layer was almost statically neutral, whereas strong nocturnal inversions were typically observed on the clear nights.

Schrage et al (2007) demonstrated that the cloudy nights in Parakou were characterized by strong moisture convergence at 925 hPa in the ECMWF operational analyses. This moisture convergence was driven by enhanced speed convergence over the Guinea Coast climate zone, primarily caused by an increase in the impact of friction on the flow near the surface. As a result, the authors suggested that cloudy nights are maintained through a positive feedback in which the clouds contribute to a destabilization of the near-surface layer through their radiative impact; destabilization promoted vertical mixing between the surface and the stronger flow at 925 hPa, coupling that level to the surface and thereby increasing the impact of friction. The flow decelerates, and the resulting moisture convergence and vertical mixing support the ongoing production of flow clouds.

a.)



b.)

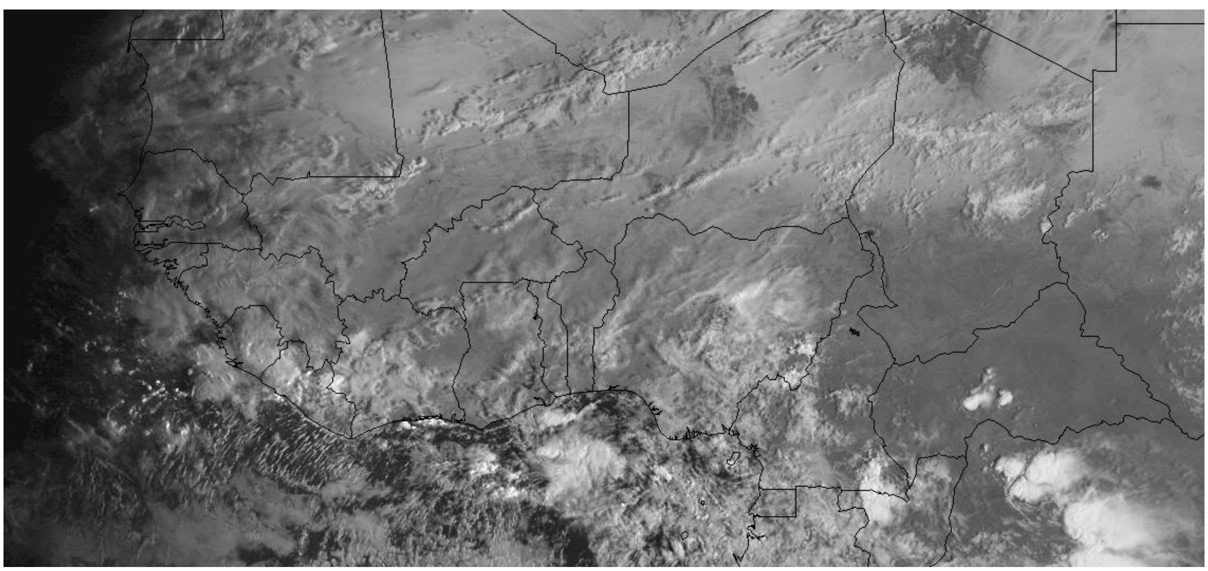


Figure 1: a) Visible satellite image for 0715 UTC August 4, 2006, showing widespread nocturnal stratiform clouds across the Soudanian and Guinea Coast climate zones of tropical West Africa. b.) Visible satellite image for 0715 UTC October 15, 2006, showing clear conditions across Benin.

In contrast, Schrage et al. (2007) proposed that on the clear nights radiative cooling of the surface shortly after sunset decouples the near-surface layer from the surface itself through the formation of a shallow and extremely stable layer in the lowest few tens of meters. Monsoon southwesterlies at 925 hPa in these cases did not experience the surface friction and therefore did not decelerate over the Guinea Coast zone, yielding much less moisture convergence. The drier conditions and reduced vertical motion prevented the formation of clouds, creating a positive feedback that maintained the clear conditions.

The mechanisms hypothesized in Schrage et al. (2007) provide a reasonable mechanism for how the presence or absence of low clouds over sub-Sahelian West Africa is maintained throughout the night. However, that study was not able to address the question of why individual nights either become cloudy or remain clear.

In the present study, a possible connection between a nocturnal low-level jet and the development of these low cloud decks will be explored. The diurnal cycle of the flow over tropical Africa has been explored by a number of previous studies, most notably Farquharson (1939), Hamilton and Archbold (1945), and Parker et al. (2005). These studies and others are in agreement that there is a nocturnal peak in the magnitude of the low-level flow a few hundred meters above the surface, or approximately the 925 hPa level. These winds appear to be consistent with the structure and evolution of a nocturnal low-level jet first suggested by Blackadar (1959). According to this theory, turbulence in the convective boundary layer during the day produces sufficient drag to prevent low level winds from reaching geostrophic velocities; only at night, when a nocturnal inversion decouples the residual layer from the surface can geostrophic or supergeostrophic velocities be achieved. In a monsoon region, the pressure gradient driving the circulation typically maximizes during the day due to the enhancement of the continental-scale heat low, but this pressure oscillation fails to produce enhanced winds until approximately 12 hours after the peak heating. This pattern has been confirmed by a number of more recent studies (e.g., May 1995, Racz and Smith 1999) of the diurnal cycle of the winds of the Australian monsoon.

2. Data

As part of the preparations for their 2006 field campaign, AMMA (African Monsoon Multidisciplinary Analyses) researchers equipped the radiosonde station at Parakou, Benin with a new MODEM ground station (Parker et al. 2008). During different phases of the AMMA field campaign, nominally 2, 4 or 8 radiosondes were launched daily. As is standard practice, the balloons are actually launched approximately one hour prior to the official synoptic time; since the balloon crosses the levels of interest within the first few minutes of the flight, the radiosonde data from Parakou is being attributed to the hour at which the balloon was actually launched, rather than the synoptic hour to which the data would normally be ascribed in a data assimilation.

Estimates of the static stability of various layers of the atmosphere were calculated from temperature retrievals made by a HATPRO (Humidity and Temperature PROfiler) microwave profiler operating at Nangatchori, Benin (9.65°N, 1.73°E, or approximately 100 km west-northwest of Parakou) in 2006. The physical configuration and geographic setting of this profiler has been described in detail by Pospichal and Crewell (2007). The instrument resolves the temperature structure of the atmosphere every 15 minutes. For the present study, calculations were based only on the temperature estimates at levels at or below 1000 m above ground level, and the vertical resolution of the data in this layer is 25 m. Crewell and Löhnert (2007) have shown that the temperature estimates obtained from such a profiler are accurate to less than 1 K in the lowest 1.5 km of the atmosphere and that these data are particularly useful for assessing the height and strength of low-level temperature inversions. The platform also featured a temperature and relative humidity sensor that these measurements are employed in Section 6 of the current study.

During 2006, a lidar ceilometer was collocated with the aforementioned microwave profiler. Pospichal and Crewell (2007) provide a description of the configuration of this instrument, which had a temporal resolution of 15 seconds and a vertical resolution of 30 m. This instrument was used in the present study to determine the exact time of onset of stratocumulus clouds in the region near Nangatchori.

As part of the AMMA field campaign, a UHF wind profiler was installed at Djougou, Benin (9.70°N, 1.68°E, or less than 10 km northwest of Nangatchori). The use and interpretation of data from such a wind profiler has been summarized by Jacoby-Koaly et al. (2002). Wind profilers were derived from the instrument approximately 14 times per hour on average at 13 levels below 1000 m above ground level (nominally, 74, 149, 224, 299, 374, 449, 524, 599, 674, 749, 824, 899, and 974 m). This instrument was put into operation in April of 2006, but most of the data gathered prior to August 7, 2006 are missing due to a variety of technical problems on-site.

Additionally, visible and infrared satellite images, as well as traditional synoptic reports of weather conditions, were used to verify the conditions on the dates chosen as cases for this study.

3. August 9 case

Figure 2 depicts the meteorological events of an example of a night in which

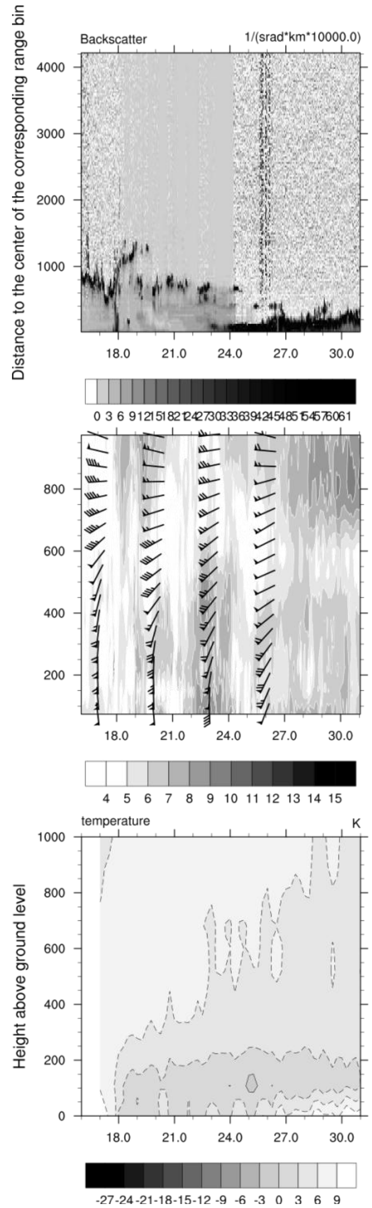


Figure 2: Time-height cross-sections of various meteorological parameters for the period between 1600 UTC August 8, 2006 to 0700 UTC August 9, 2006. The top panel depicts the ceilometer observations at Nangatchori. The middle panel depicts wind speed as determined by the profiler at Djougou (shading) and the winds as observed at Parakou (wind barbs, multiplied by 10.) The bottom panel depicts the static stability $-\partial T/\partial z$ based on temperature retrievals from the HATPRO profiler at Nangatchori.

stratocumulus clouds formed very early. While the ceilometer detected some scattered clouds below 1000 m in the first half of the night, a thick layer of low clouds was observed started shortly before 0000 UTC, and these clouds persisted for the remainder of the night. The wind profiler at Djougou identified the formation of a nocturnal low level jet that peaked with a magnitude of more than 8 m s^{-1} at about 400 m a.g.l. around 2300 UTC on August 8. The available radiosonde observations from Parakou provide supporting documentation that these were the strongest low-level winds of the night and that they represented an enhancement of the flow in the monsoon layer.

Throughout most of the night (including the periods both before and after the onset of low clouds), the Richardson number (not shown) was below 0.25 for the layers below 200 m a.g.l. The fact that the lowest layers of the atmosphere did not get particularly stable on this night probably both reflects the fact that turbulent mixing continued throughout the night and provides an explanation for the low values of that parameter. Overall, this night strongly resembles the patterns discussed by Schrage et al. (2007), which is not surprising as that study focused on the nights in which the stratocumulus clouds were already noted in the 0000 UTC synoptic reports in the Soudanian zone.

4. September 22 case

Very low stratocumulus clouds were observed at Nangatchori starting around 0445 UTC on the morning of September 22, 2006 (Fig. 3). Prior to that time, the ceilometer shows that the sky had been clear below 4 km since approximately 2000 UTC. The onset of a nocturnal low level jet around 0000 UTC lead to

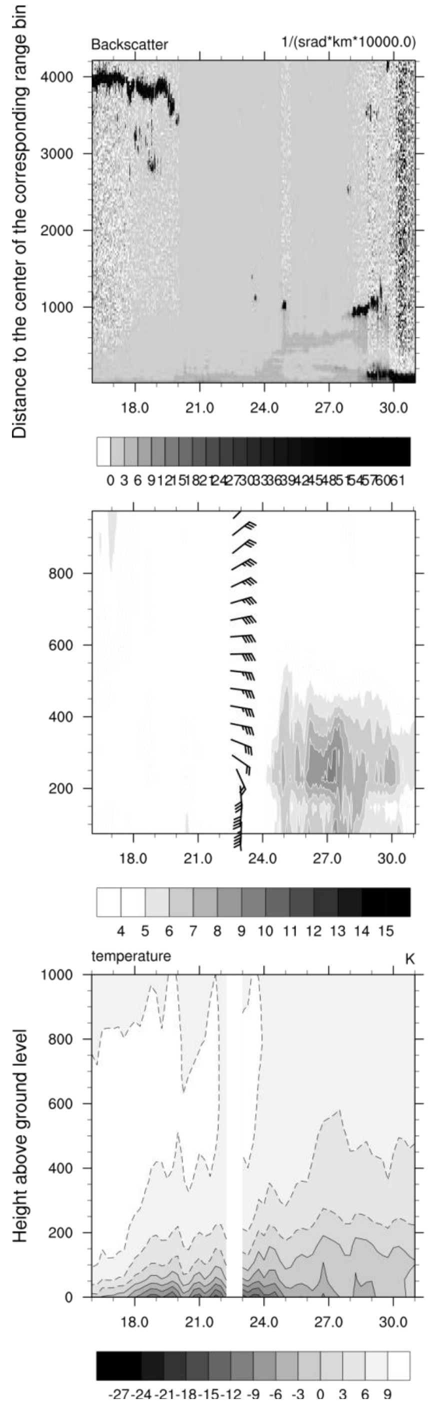


Figure 3: Same as Fig. AUG9, except for the period between 1600 UTC September 21, 2006 to 0700 UTC September 22, 2006.

a peak wind of more than 9 m s^{-1} around 0315 UTC at about 225 m above ground level at

Djougou. The only available radiosonde wind profile at Parakou (namely, 2300 UTC) is in agreement that the jet had not yet become established at that early hour. From sunset until approximately 0000 UTC, the near-surface layer became increasingly stable, with lapse rates of less than -12 K km^{-1} detected for several hours. The periods of decreasing stability corresponded well with periods of low Richardson number (not shown), suggesting that the mechanical mixing caused by the nocturnal low level jet was contributing to the destabilization of the near-surface layer.

5. August 25 case

Although it is true that the ceilometer at Nangatchori did detect the passage of some scattered clouds below 1500 m a.g.l. on the night of August 24-25, 2006 (see Fig. 4), these clouds were both too isolated and too high to be examples of the kinds of nocturnal stratocumulus decks that are the subject of the present work. In the minutes before sunrise, the ceilometer did register some lidar returns from some very low hydrometeors, but it is not clear whether these are stratocumulus clouds or simply fog. Unfortunately, the synoptic reports from nearby stations at this hour are missing, but it seems reasonable to state that long-lasting stratocumulus decks did not form that evening over Nangatchori.

The wind profiler data in Fig. 4 are not "missing", despite the lack of contours on that panel. Rather, the winds on that night were sufficiently weak that at no time during the night did any of the winds reach 5 m s^{-1}—the threshold used for shading and contouring on the plot. The superimposed wind observations from Parakou support the idea that the flow in the monsoon layer did not experience the expected nocturnal enhancement. These winds combined with the moderately strong stability of the near-surface layer to limit the circulation to laminar flow on this night.

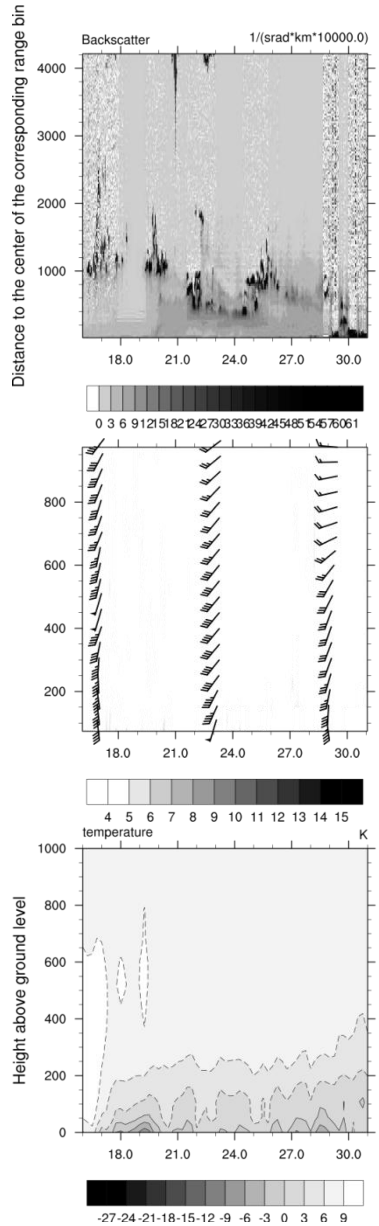


Figure 4: As in Fig. 2, but for the period from 1600 UTC on August 24, 2006 to 0700 UTC on August 25, 2006.

6. Discussion and Summary

In the present study, it was shown that the formation of nocturnal stratiform cloud decks is preceded by a period of destabilization in the near-surface layer. This destabilization seems to occur on most nights in which a nocturnal low level jet forms in the lowest 200 to 400 m of the boundary layer. This jet produces mechanical turbulence and vertical mixing that tends to reduce the static stability of the layer. This process naturally competes with the tendency of the near-surface layer to become increasingly stable over the course of the night by purely radiative mechanisms. The timing and magnitude of the jet and the radiative processes becomes critical for the formation of nocturnal stratocumulus clouds. If too much radiative cooling of the surface happens before the onset of the jet, the near-surface layer exhibits high Richardson number flow and the resulting destabilization of the layer by vertical mixing is much too slow to produce stratocumulus clouds before sunrise. On the other hand, if the jet sets up early enough or is strong enough to overcome the stability produced by the radiative processes, a low Richardson number flow is established promoting strong vertical mixing. This mixing and resulting destabilization typically lead the onset of stratocumulus clouds by more than one hour.

However, if the mechanism that is both destabilizing the near-surface layer and providing vertical mixing is the nocturnal low level jet, then it remains to be explained why this jet is observed on so many nights that remain clear. Broadly, three hypotheses seem plausible:

1. It is possible that the near-surface air is sufficiently dry that the mechanical mixing and lifting produced by the nocturnal jet simply isn't able to lift air parcels to their Lifted Condensation Level. The extremely low cloud bases of these stratiform cloud decks are typically at or below 200 m a.g.l., implying that that the near-surface dewpoint depression would need to be no more than about 2 K in order to be consistent with the clouds that were observed. However, the environmental relative humidity sensor at the HATPRO profiler showed that the lowest layer was saturated, even on the clear nights. (The only exceptions were a few clear nights in the pre-monsoon season.) Therefore, this possibility seems not to have been the case on the observed nights in 2006, but it remains possible or even likely that some nights in other years are clear for this reason.

2. It is possible that the nocturnal low level jet forms on some nights only once the nocturnal inversion is so strong that, even in the presence of strong vertical wind shear, the flow remains laminar. In some ways this seems reasonable, as the decoupling of the surface from the residual layer is a fundamental reason why a nocturnal low level jet forms (e.g., Blackadar 1957). It may be the case, therefore, that the rate of cooling in the first few hours after sunset is critical to this scenario. Once the residual layer decouples from the surface in the evening, acceleration of the low level begins, but this process takes time and is probably not strongly dependent on the stability of the air below jet level. If during this period of the acceleration of the winds, strong diabatic cooling is occurring near the surface, it may be possible that the vertical shear associated with the nocturnal low level jet forms in an environment that is strongly stably stratified.

3. One can also envision situations in which the onset of the nocturnal low level jet occurred simply too late in the night for stratocumulus cloud decks to form. In this

scenario, the jet and its associated vertical wind shear are able to produce mixing and destabilization of the near-surface layer, but the resulting $\partial\Gamma/\partial t$ simply does not have enough time to sufficiently destabilize the boundary layer to promote cloud growth. The day-to-day variability of the nocturnal low level jet was beyond the scope of the present study. However, an investigation into the processes that determine the timing and magnitude of this jet may well prove to be a fruitful line of research.

References

Blackadar, A.K., 1957: Boundary layer wind maxima and their significance for the growth of nocturnal inversions. *Bull. Am. Meteorol. Soc.,* **38**, 283–290.

Crewell, S. and U. Löhnert, 2007: Accuracy of boundary layer temperature profiles retrieved with multifrequency multiangle microwave radiometry. *Geoscience and Remote Sensing, IEEE Transactions*, **45**, 2195-2201.

Farquharson, J. S., 1939: The diurnal variation of wind over tropical Africa. *Quarterly Journal of the Royal Meteorological Society*, **65**, 165-184.

Hamilton, R. A., and J. W. Archbold, 1945: Meteorology of Nigeria and adjacent territory. *Quarterly Journal of the Royal Meteorological Society,* **71**, 231-264.

Jacoby-Koaly, S., B. Campistron, S. Bernard, B. Bénech, F. Ardhuin-Girard, J. Dessens, E. Dupont, and B. Carissimo, 2002: Turbulent dissipation rate in the boundary layer via uhf wind profiler doppler spectral width measurements. *Boundary-Layer Meteorology*, **103**, 361-389.

May, P. T., 1995: The Australian nocturnal jet and diurnal variations of boundary-layer winds over Mt. Isa in north-eastern Australia. *Quarterly Journal of the Royal Meteorological Society* , **121**, 987-1003.

Parker, D. J., R. R. Burton, A. Diongue-Niang, R. J. Ellis, M. Felton, C. M. Taylor, C. D. Thorncroft, P. Bessemoulin, and A. M. Tompkins, 2005: The diurnal cycle of the West African monsoon circulation. *Quarterly Journal of the Royal Meteorological Society*, 131,, 2839-2860.

Parker, D.J., A. Fink, S. Janicot, J.-B. Ngamini, M. Douglas, E. Afiesimama, A. Agusti-Panareda, A. Beljaars, F. Dide, A. Diedhiou, T. Lebel, J. Polcher, J.-L. Redersperger, C. Thorncroft and G.A. Wilson, 2008: The AMMA radiosonde program and its implications for the future of atmospheric monitoring over Africa. Bull. Amer. Met. Soc., **89**, 1015-1027.

Pospichal, B. and S. Crewell, 2007: Boundary layer observations in West Africa using a novel microwave radiometer. *Meteorologische Zeitschrift,* **16**, 513-523.

Rácz, Z. and R. K. Smith, 1999: The dynamics of heat lows. *Quarterly Journal of the Royal Meteorological Society,* **125**, 225-252.

Schrage, J. M., S. Augustyn, and A. H. Fink, 2007: Nocturnal stratiform cloudiness during the West African monsoon. *Meteorology and Atmospheric Physics*, **95**, 73-86.