The Harmattan over West Africa: nocturnal structure and frontogenesis

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This paper presents observations and model simulations of the low-level nocturnal structure of the atmosphere over West Africa. The measurements are taken from the dry-season Special Observing Period (SOP-0) of the African Monsoon Multidisciplinary Analysis (AMMA), at Niamey, Niger, on January 23rd and 24th 2006. During this time, mesoscale structures in the atmospheric aerosol loadings were observed. The available observations indicate that these mesoscale features at Niamey are consistent with the passage of gravity currents or bores in the northerly Harmattan winds.

Model simulations at resolutions down to 1 km indicate that the mesoscale structures are caused by nocturnal frontogenesis in the baroclinic zone to the south of the Sahara in the winter months. This frontogenesis is a continental-scale phenomenon, which has significant implications for the uplift and transport of dust and biomass-burning aerosols in the region. An accompanying frontogenetic feature appears further south in the model simulations, associated with the winter inter-tropical front. The frontogenesis is possibly linked with the development of structures showing characteristics of canonical mesoscale phenomena, including internal bores and gravity currents. Representation of these features in different models (the UK Met Office Unified Model and the Weather Research and Forecasting (WRF) model of the US National Center for Atmospheric Research), and at different resolutions (from 12 km to 1 km), is discussed. Copyright © 2011 Royal Meteorological Society

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1. Introduction

During the months of November to March, the climate of the West African region (see Figure 1) is dominated by north-easterly trade winds. These winds, commonly referred to as the Harmattan, arise due to synoptic-scale pressure gradients that align north-south across the Saharan desert. Although the Harmattan is one of the major wind systems of Africa, it has received relatively little attention in the literature, probably due to its dryness, and therefore its lack of association with the critical rain-bearing storms.
of the Sahel. However, for the local populations and for the global climate system, the Harmattan remains extremely important. A significant quantity of mineral dust is advected southwards across the region by these winds, leading to a major impact on local visibility, agriculture, and health. Furthermore, the Harmattan is relatively cool at this time of year, when the warmest temperatures in the lower troposphere are found around 8 °N, and synoptic variability in its southward extent can lead to intrusions of Harmattan winds, and cold events in the countries of the Guinea coast (Hamilton and Archbold 1945). Such cold events can cause hardship and even mortality to vulnerable local people. Finally, winter-time rains in the region have a disproportionate economic impact, influencing the early development of crops and in some cases causing stored foods to rot (Knippertz and Fink 2008).

Figure 1. The West African region showing political boundaries. The outer domain represents the model domain for the 12 km simulation; the 4 km and 1 km simulation domains are shown by rectangles. Niamey is marked with an asterisk. Basemap created by ArcMap.

Schematic summaries of the climatic state in these winter months have typically focused on the inter-tropical front (ITF), which separates the southerly winds flowing off the Gulf of Guinea from the Harmattan. Hamilton and Archbold (1945) show the typical structure of the ITF, sloping southwards with height, and describe the diurnal cycle of winds close to the ITF. More recently, making use of airborne and radiosonde and lidar data from the African Monsoon Multidisciplinary Analysis (AMMA) dry season special observing period (SOP-0), Haywood et al. (2008) have shown how the atmospheric state over West Africa in winter typically displays two frontal zones: the ITF is located, as described by Hamilton and Archbold (1945) between 5°N and 8°N, with the frontal zone ascending towards the south, while to the north there is a separate “Harmattan front” located around 20°N at the surface, and ascending towards the north. The Harmattan front appears to separate the cool air in the low levels in the Sahara from the warmer air to the south. Haywood et al. (2008) also showed how dust uplift in the Harmattan winds appears to lead to fronts in the aerosol loading. Currently it is not clear how such dust fronts are formed, whether by local or large-scale processes, and it is one goal of this article to shed light on this issue.

The collection of AMMA SOP-0 observations showed remarkable structure in the atmospheric aerosol loadings in January 2006 (Haywood et al. 2008; Marticorena et al. 2010). The presence of these aerosol is an important factor in global and regional climate.

A number of past studies have shown that the winds in this region have a marked diurnal cycle (Parker et al. 2005, Washington et al. 2006, Todd et al. 2008, Lothon et al. 2008, Abdou et al. 2010). This diurnal cycle will have a significant impact upon dust uplift from the surface and aerosol distributions in general. Put simply, by day the turbulence in the convective boundary layer retards the large-scale atmospheric circulations and, while turbulent eddies may be strong, the mean winds are relatively light. Around sunset, when the convective turbulence decays, the winds accelerate down the pressure gradient, and a nocturnal jet may form. The stability of the nocturnal surface layer means that the flow can be decoupled from the surface, and commonly there is no signature of the nocturnal
jet in the mean surface winds (Lothon et al. 2008); however, after sunrise the growing convective boundary layer can mix down momentum from the jet, leading to the diurnal peak in surface winds (May 1995, Parker et al. 2005) at which time dust uplift may reach a maximum (Washington et al. 2006). These studies have revealed that the diurnal cycle plays a crucial role in West African monsoon dynamics, and has a key impact upon the aerosol distribution.

This paper aims to explore the processes of frontogenesis which have been inferred from the observations of Haywood et al. (2008) and to examine the nocturnal dynamical structure associated with the observed dust uplift. We make use of modelling of events taken from AMMA SOP0, and some associated measurements. Section 2 outlines lidar measurements of dust events that act as a primary motivation for the current study. Section 3 describes the numerical models which have been used to simulate the diurnal variability of the Harmattan, and the following section uses the models to explain the dynamical characteristics of the system.

2. Dust observations 23rd-24th January, 2006

During the period 23rd to 24th January 2006, measurements from the Lidar Aérosols UltraViolet Aéroporté (LAUV A) backscatter lidar (Chazette et al. 2007) situated at Niamey Airport show evidence for an overnight passage of a dust plume (Figure 2, reproduced from Haywood et al. 2008). This feature is consistent with reports from Niamey Airport of a coherent nocturnal feature experienced regularly in this region, that is accompanied by enhanced concentrations of dust. The work presented here examines the dynamical nature of this dust plume and using high-resolution model simulations presents evidence for a frontal feature at the leading edge of the Harmattan flow.

Closer inspection of Figure 2 shows that overnight on the 23rd to 24th January, a number of specific dust events occurred over Niamey. Of particular note is the event starting at 2100 UTC which raises dust to a height of approximately 700 m above mean sea level, equivalent to around 500 m above ground level (AGL), by 2200 UTC. Subsequent events at 2330, 0030 and 0200 UTC can also be identified. From this one-dimensional image we can make a few general statements about the arrival of “dust fronts” at Niamey on this night. Typical events were around 500 m or 600 m in vertical extent. Each event shows a coherent plume of dust which can be traced from the surface to a higher level. Ascent rates of the dust uplift have been estimated by Haywood et al. to be of the order 10 cm s$^{-1}$ to 30 cm s$^{-1}$.

We suggest that the events observed on the night of 23rd-24th January 2006 correspond to the passage of a gravity current around 2100 UTC, followed by internal bores in the stably-stratified nocturnal flow. Bores propagating on nocturnal inversions have been observed at other sites in the world, such as Australia (Smith 1988, Goler and Reeder 2004) and the USA (Koch et al. 2008). A bore is a nonlinear, solitary wave which propagates through the fluid, leading to a permanent upward displacement of air parcels. Unlike a gravity current, in which fluid is materially carried into the convergence line from behind, the bore moves as a wave, with rearward relative motion of air parcels. Therefore (as in Koch et al. 2008) while a gravity current exhibits abrupt cooling in surface station data, a bore generally does not, and may exhibit surface warming due to enhanced mixing in a nocturnal inversion. Unfortunately we lack profiles of winds before and after this event to evaluate the relative motion of winds, but in the events after 2100 UTC, the relatively weak upward motion, the lack of abrupt cooling at the surface (as revealed in the relevant pressure-temperature-humidity sonde data for Niamey (not shown)), and the results of numerical modelling studies (Section 4 below), are consistent with the wavelike behaviour of a bore.

To bring Figure 2 into a wider context, Figure 3 shows observations taken from the ‘Ceilometer’ lidar, also situated at Niamey Airport. Figure 3(a) shows the vertical profile of backscatter from 24th January and, as with the LAUV A lidar, shows a prominent dust event passing over the region at around 0200 UTC. This event raises dust to
approximately 500 m above ground level, as shown also by the LAUV A lidar. Figure 2 also shows evidence of already lifted dust concentrations at 0000 UTC that is consistent with Figure 2.

Figure 3(b) shows the average backscatter across January and February 2006 from the same instrument. The data indicate that the dust features observed in both Figure 2 and Figure 3(a) have a significant effect on the mean aerosol on the period, and seem to occur on a regular basis: the latter is confirmed by the anecdotal reports of scientists working at the site. The most striking aspect of Figure 3(b) is the increase in dust concentrations in the period around 0100 UTC, which suggests that a coherent dynamical mechanism may be operating on a regular basis at similar times of the night. Another peak in dust loading may be observed at around 0900 UTC: this is associated with the daily maximum in wind speed at Niamey (May 1995, Parker et al. 2005), but is outside the scope of this paper.

A visual inspection of all of the ARM profiler data for the 51-day period from 9 January to 28 February 2006 (paying attention to any abrupt changes in low-level aerosol concentration below 1000 m) suggested that similar dust events were recorded on at least 20 nights, and possibly another 7 nights. In addition, on 3 to 6 other nights in this period there was evidence of a rapid, non-dusty airmass change.

In conclusion, the nocturnal dust events observed on the night of 23rd-24th January 2006 are consistent with the arrival of a series of convergence lines carried in generally northerly winds. These features appear to be important in the lifting of dust from the surface and its elevation above the atmospheric surface layer, and there is evidence that such events occur on around half the nights in January and February at Niamey. In the next sections we describe their representation in high resolution atmospheric models, and the ambient conditions which cause them.

3. Analysis of frontal features in the model results

3.1. Models used

In this study we make use of nested model simulations using the Met Office Unified Model (UM). The model is initialised using the global UM operational analysis for 22nd January 2006, and then is nested down to subdomains at 12 km, 4 km and 1 km resolutions, with 50 vertical levels. In order to supplement the UM runs, a limited set of simulations using the Weather Research and Forecasting (WRF; Skamarock et al. 2008) model was performed. In these runs, WRF was initialised with analyses (synchronised with the UM analyses) derived from the US Global Forecast System of the National Center for Environmental Prediction. In a similar manner to the UM runs, the WRF simulations were nested down to 4 km
3.2. Convergence lines in the 4 km and 12 km models

Figures 4 (a) to (f) show the wind field at 350 m altitude as well as regions of convergence, at 2-hourly intervals during the night of 23rd - 24th January, 2006, with data taken from the 4 km model. At 2200 UTC (Figure 4(a)), the wind field north of 13°N is predominantly from the north and reaches velocities of up to 10 m s\(^{-1}\). In the southern part of the model domain, south of approximately 10°N, winds are from the south with velocities, on average, marginally less than those in the north of the domain. At the same time, there is a marked region of much calmer conditions separating the two opposing flows. The region of relatively calmer winds corresponds to the monsoon trough—the local maximum in temperature and minimum in pressure—at this time of year (e.g. Lavaysse \textit{et al.} 2009).

Figure 4 (a) also shows that at this time there are two lines of convergence, or fronts, in the 4 km model data. A northernmost front lies at the head of the northerly flow, around 12°-13°N over longitudes 4°W to 4°E, and, with further justification, we will call this the “Harmattan front”. A southern front, representing the ITF, lies at 9°-10°N across the entire domain, and separates monsoon southerlies to the south from light northerlies. The Harmattan front is less coherent than the ITF at this time, and occurs around the leading edge of the enhanced northerly winds, as a seemingly disorganised band of convergence lines. Among these is a convergence line extending north-eastwards from 13°N, 3°E to the foot of the Air mountains (shaded region at top right of the domain). This line appears, at least in part, to be a consequence of the mountain itself with winds being channelled around the mountain into an easterly flow that eventually collide with the northerly flow.

Figures 4(b) to (f) show how the wind field and regions of convergence evolve overnight. By 0200 UTC (Figure 4(c)), the region of northerly winds has moved significantly southwards with the Harmattan front now stretching into northern Benin. The convergence lines associated with this flow have now organised themselves into a more contiguous structure that stretches across much of the model domain. Simultaneously, the monsoon flow has moved slightly northwards whilst maintaining a coherent frontal band at the ITF.

An image of the $\theta$ field at 0300 UTC at approximately 300 m above ground (Figure 5) confirms that both the southerly monsoon flow and the northerly Harmattan flow have distinct temperature gradients at their leading edges, which are coincident with the convergence lines in the wind field. Figure 5(b) shows sharp gradients in potential temperature co-located with the regions of convergence highlighted in Figure 4, with coherent east-west extent. Both the Harmattan front and the ITF represent a flow of cool air intruding into warmer air, from the north and south respectively. Figure 5(a) shows a transect of potential temperature aligned north-south through the centre of the domain at 500 m altitude. The band of air separating both flows lies between approximately 11°N and 11.5°N, with a potential temperature reaching 312 K. On either side of this a significant drop in potential temperature occurs: for instance, across the northern front a drop of 2 K occurs over a distance of approximately 50 km. Figure 5 also highlights the thermally-direct nature of the model atmosphere during this time. Relatively cool flows from the north and south are associated with radiatively cooled air from the Sahara, and an air mass originating over the Gulf of Guinea, respectively. These air masses encroach at low levels into a region of warmer air.

By 0400 UTC (Figure 4(d)), the opposing flows have now collided to make one band of convergence. Figure 6 (a) shows a Hovmöller-type diagram of horizontal wind convergence and vertical velocity for the period covering Figure 4, again using data from the 4 km model. In this figure, the Harmattan front can be identified as a band of convergence moving from around 14.5°N at 1800 UTC on 23rd January (hour 42 of the model run) to collision with the ITF around 11.3°N at 0400 UTC (hour 52). It is remarkable
from this figure that both of the convergence lines—the Harmattan front and the ITF—appear to persist after the collision and continue at something close to their original speed at least until 0900 UTC (Figures 4(e) and 4(f)). The speeds of both the Harmattan front and the ITF are calculated from this diagram, by observing the movement of these lines of convergence in time, i.e. by calculating the time-latitude gradient.

This kind of behaviour, with features persisting though a head-on collision, is characteristic of a collision of solitary waves, but also may occur if gravity currents of differing densities collide (e.g. Goler and Reeder 2004, Thomsen et al. 2009).
It is instructive to consider the scalar frontogenetic function (see, e.g. Markowski and Richardson 2010, pages 122-123). The full scalar frontogenetic function calculates the rate of change of the horizontal gradient of potential temperature, with the gradient taken along a line which is normal to the front, in our case taken to be the $y$-axis (northward). Unlike Markowski and Richardson (2010), we here define positive frontogenesis to represent an increase in the gradient of potential temperature with positive $y$.

The frontogenetic function $F$ contains contributions from horizontal shear, confluence, tilting and the horizontal gradient of diabatic heating. In the present work the gradient of diabatic heating has been neglected since its contribution is negligible compared to the other terms. A function such as $F$, based as it is upon various derivatives, is by nature noisy, and so $F$ has been smoothed in the north-south direction using a 13-point (corresponding to approximately 50 km) box filter. This removes perturbations on the order of 20 km. $F$ is plotted in Figure 6 (b). Slight care must be taken in interpreting this figure, since the direction of the cold air is different for the northern and southern air masses, with warmest low-level air around 11°N-12°N. In particular, for the Harmattan front in the north of the domain, where temperature is decreasing towards the north, positive $F$ is frontolytic, while negative $F$ is frontogenetic. With this in mind, $F$ is seen to be frontogenetic along and ahead of both the Harmattan front and the ITF, along the lines of convergence seen in Figure 6 (a). Analysis of the individual terms in the frontogenetic function (not shown) suggests that at the lower levels illustrated in Figure 6 (b), confluence is the dominant term. At slightly higher levels, e.g. 600 m, tilting leads to overall frontolysis. Area averages of $F$ for the whole domain, for the region 0°-2°E, and separately for the regions north and south of 12°N, show that the large-scale average frontogenesis function has small positive values, and therefore that the frontogenesis is dominated by the local circulation and not the large-scale forcing. Indeed, the large-scale average of $F$ is slightly frontolytic for the Harmattan front. For these reasons we infer that the frontogenesis in the Harmattan front is not caused by large-scale frontogenetic forcing, but by mesoscale overturning of the temperature gradient at night, as described by Linden and Simpson’s (1986) laboratory experiments.

Figure 4 has shown that a frontal feature occurs at the leading edge of the Harmattan between approximately 10°N and 14°N. Wind convergence in the 12 km simulation (i.e. on a larger domain, not shown), confirms that this frontal feature is also well represented in the lower-resolution simulation, and is of significant horizontal extent, identifiable between around 10°W and 8°E.

In the WRF simulations, similar features are observed. For example, in Figure 7 several coherent convergence lines can be seen, of similar strength to those simulated by the UM, and in similar locations. Time series of the movement of these convergence lines in the 12 km model simulations (not shown) display the same general development to that seen in the UM: southerly- and northerly-moving flows collide and, what is more, persist after the collision.

We conclude that the mesoscale features identified at the Niamey site are part of a continental-scale circulation. A simulation with the regional topographic features removed showed consistent convergence zones (not shown), confirming that these zones are generated from the continental-scale atmospheric dynamics, rather than from topographic forcing.
3.3. Frontal structures at high resolution

In order to investigate the details of the fronts, or convergence lines, the 1 km model results are presented in Figures 8 and 9, which show a north-south orientated vertical slice through the region of colliding fronts from the 1 km model.

At 0100 UTC, the existence of the “monsoonal” flow from the south is quite apparent (Figure 8). The head of this monsoonal flow, travelling at close to 5 ms\(^{-1}\), is situated at approximately 10.85°N with a temperature contrast (relative to the air mass to the north) of the order of 1 K and a significant convergence, leading to upward motion. A few kilometres behind this is a more distinct thermal contrast at around 10.7°N. At each of these features the relative air flow is northerly (blue line in lower panel of Figure 8) meaning that these features are consistent with the presence of internal bores, propagating into the ambient stable stratification (see Koch et al. 2008). Ahead of the monsoon flow is a slightly opposing flow of broadly cooler air around 11°N. To the north, around 12.2°N, is the head of the Harmattan flow, represented by a cool surge of around 1-2 K and strengthening northerly winds. The meridional wind component relative to this feature (red line in bottom panel) shows that there is relative inflow on both sides of the convergence line, implying that the line is characteristic of a gravity current.

By 0200 UTC these features have all become more distinct (Figure 9), with a strong separation between the two convergence lines in the ITF, now at 10.6 and 11.1°N. The existence of the two separate convergence lines in the ITF can be more properly associated with subcritical flow (in the sense defined by Haase and Smith 1989), leading to a detached head. This is qualitatively similar to the subcritical flow cases discussed in Thomsen et al. 2009.

Significant undulations in the potential temperature contours in this region are indicative of resonant waves in the low-level flow, as described by Goler and Reeder (2004) for an Australian case of colliding sea-breezes.
4. Linking Simulations and Observations

As mentioned above, it is problematic to compare model results with observations in this case. Routine observations are sparse in this region, which affects the quality of the analyses used to drive the models. Little surface-based data exists, and this is difficult to interpret in light of the decoupling of the surface from the atmosphere aloft, as occurs at night. Additionally, there is no dust module in the UM, and comparing modelled dust loadings with those observed is impossible. It is possible, however, to investigate certain model fields serving as proxies for dust uplift.

Figure 11 (a) shows the modelled potential temperature at Niamey Airport, plotted as a time-height diagram, together with the lidar observations (previously shown in Figure 2). As can be seen, there is a remarkable concordance between the motion of the lifted dust and the isentropes. In particular, the principal lifting episodes coincide with enhanced tilting of the isentropes aloft. This can be seen at 0000 UTC where the tilting is apparent throughout the lowest 1.5 km. A similar structure is seen at approximately 0200 UTC when the isentropes are again tilted upstream. In addition, between these episodes of isentropic tilting, there are periods (for example at 700 m AMSL at 0130 UTC) when the horizontal $\theta$ (and hence density) gradients at lower levels are horizontal; at these times the dust also maintains a constant altitude. In general, the dust signal follows the isentropes remarkably well. Observed small scale features are not shown in the model results (in part due to the frequency of model output and the model resolution) and

Harmattan convergence line remains distinct, near 11.8°N, despite the thermal contrast across the feature remaining relatively weak, of the order of 1 K.

After the collision between the two fronts, the leading edge of the Harmattan front flows over the detached head of the ITF front, allowing the convergence lines (at 350 m) to propagate at the same speed as before. This can be seen in Figure 10 where the movement of the air masses (again at 350 m) is indicated by arrows.
the above analysis of course relies upon Taylor’s hypothesis being applicable.

The simulated vertical velocities are shown in Figure 11 (b). As would be deduced from the isentropes shown in Figure 11 (a), there are three principal episodes of lifting. The first occurs at 2130 UTC, with vertical velocities of approximately 2 cms$^{-1}$; the second at approximately 2330 UTC, with values of vertical velocity of 4 cms$^{-1}$; and the third at 0145 UTC with vertical velocities of 5 cms$^{-1}$. These episodes of enhanced vertical velocity in the model coincide very well with the times when the dust was observed to be lifted, although the vertical velocities are lower in magnitude than those estimated by Haywood et al. 2008.

The similarities in Figures 11 strongly suggests that the model is capturing the dynamical effects that can be inferred from the lidar observations.

5. Discussion and conclusions

Haywood et al. (2008) presented the idea of the Harmattan front, based on AMMA SOP-0 observations. Here we have explored the behaviour of this front in more detail, shedding light on its structure, frequency of occurrence and dynamical properties. Further observations from the ARM Mobile Facility at Niamey confirm the existence of the Harmattan front and suggest that it occurs quite regularly at night in the months of January and February. Estimated uplift speeds of the observed aerosol layers suggests that the embedded mesoscale features or convergence lines within the front can resemble an internal bore, or a gravity current, at least for the event of 23rd-24th January. Model simulations of the event at a range of resolutions, down to 1 km, show that the Harmattan front is a continental-scale feature in the model. This front exhibits internal gravity current and bore-like structures which are broadly consistent with the observations. These features are generated from the continental-scale atmospheric dynamics, rather than being forced by local topography in the vicinity of Niamey. The Harmattan front and ITF may be seen as regionally-significant weather features in the winter months over West Africa, which can lift dust from the surface and elevate aerosols (dust and biomass-burning products) to higher levels in the lower troposphere (Haywood et al. 2008).

The limited nature of the observations available to us in this Sahelian case, lacking measurements of winds and thermodynamics above the surface, means that we can only make qualitative comparison with the model results. Although all of the simulations clearly show the presence of mesoscale convergence lines, the details vary between simulations, and therefore must be treated with caution. All model runs show the possibility of formation of a gravity-current structure in the ITF and Harmattan front, but the presence of a bore or solitary wave in each shows
(an expected) sensitivity to resolution and initialisation. However, it has been shown that model proxies (potential temperature and vertical velocity) for dust uplift agree remarkably well with lidar observations. At this stage it can be said that consistency exists between model and observations to the level that mesoscale frontogenesis is observed in the measurements and the model and in each case, gravity currents and bores may be inferred. These conclusions are consistent with theory, so we can conclude with some confidence that internal bores and gravity currents occur within the nocturnal boundary layer over the Sahel in January and February, and that they are responsible for dust uplift.

Interpreting specific features of the observed data (e.g. Figure 2) in the light of the modelled response is difficult due to the fact that a spectrum of phenomena exist between bores and pure density currents (Simpson 1997). The attempt at any such interpretation is further impeded by the likely temporal offset between models and observations, and indeed the sparsity of ground-based observational data for this case. It is clear, however, that in both models and observations there can be seen bursts of activity, occurring with spatial and temporal variability, which is consistent with the existence of complex density-current and bore-like behaviour.

Further understanding of these mesoscale systems probably requires more observations, to measure the wind and thermodynamic structures above the surface in detail, and to categorise the different flows on different nights. Observations are notably lacking from the gap between the AMMA 2006 meso-sites at Ouémé (around 13.5°N) and Niamey (around 13.5°N)—exactly the zone in which the opposing fronts are most intense and collide, in our model. Pilot balloons, sodars, a tethered balloon and nocturnal ground observations would help to describe the convergence lines at relatively low cost.

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