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The diurnal cycle of the West African monsoon circulation

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SUMMARY

Using numerical model analyses, it is shown that there is a coherent diurnal cycle of the West African monsoon winds. As has been observed in previous studies of arid and semi-arid areas, the winds are at their weakest in the afternoon when the convective boundary layer (CBL) is deep, and intensify overnight when the boundary-layer turbulence is much weaker. This diurnal cycle is maximized in the northern part of the monsoon layer, where the meridional pressure gradient and the diurnal cycle of the CBL are both strong.

The diurnal cycle can also be resolved in surface and upper-air data, which show how the nocturnal meridional circulation acts to stratify the lower part of the monsoon layer. In contrast, mixing in the daytime CBL acts to maintain the baroclinicity, as has been observed in laboratory flows. This pattern has implications for the efficiency of the monsoon circulation in the continental water budget, as well as in mixing of trace gases and aerosols between the surface layer and the free troposphere. Vertical mixing occurs by day, while meridional advection, with isentropic upgliding and downgliding, is most efficient at night.

Finally, high-resolution observations from the JET2000 experiment are used to show that there is mesoscale structure in the diurnally varying monsoon circulation. In the nocturnal flows, local circulations have been observed and appear to represent a response to recent deep convective events. In contrast, the daytime CBL properties at these scales have been shown in a previous study to map closely onto patterns of soil moisture, with horizontal advection playing a weaker role.

KEYWORDS: Convective boundary layer Nocturnal jet Sahara Sahel

1. INTRODUCTION

The West African monsoon (WAM) is an important climatological system which arises during the boreal summer. As the zone of peak solar heating moves over the Sahara in June, there is a shift in the zone of peak rainfall, or intertropical convergence zone (ITCZ), inland from the Guinea coast (e.g. Sultan and Janicot 2000; Le Barbé *et al.* 2002). The WAM impacts a wide region over the continent from the West African coast right through central Africa to Ethiopia, with a remarkably zonal orientation. Understanding the WAM is, therefore, essential for understanding the climate of sub-Saharan northern Africa.

It has become apparent in recent years that global weather and climate models have systematic difficulty in simulating and predicting basic characteristics of the WAM rainfall on timescales ranging from diurnal (e.g. Yang and Slingo 2001) to annual (e.g. WCRP 2000). Perhaps not surprisingly, these models also poorly represent the key dynamical features of the WAM. For example, Thorncroft *et al.* (2003; hereafter Th03) have shown an example of a case in which both the Met Office Unified Model and the European Centre for Medium-Range Weather Forecasts (ECMWF) model were able to represent the African easterly jet (AEJ) accurately in an analysis field, but failed to locate the AEJ in a five-day forecast for the same validation time. Evidence was presented to suggest that this problem is related to errors in the boundary-layer fields

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in the northern Sahel, but the reasons for such errors were not established. Indeed, it is apparent that more process studies are required that are concerned with improving our basic understanding of the WAM and, in particular, the interactions between the dynamics and diabatic processes. This paper focuses on the interactions that take place during the diurnal cycle. We show that the WAM over the land is characterized by a distinctive diurnal cycle in regional circulations and we believe that this offers a test bed for our basic understanding of the WAM that may reap benefits for our understanding and modelling of longer timescales.

The dynamics of the WAM circulation are intimately linked to the existence of the AEJ; a mid-tropospheric flow with peak amplitude of around 10–20 m s^{-1} at an altitude of around 600–700 hPa and a latitude of around 15°N in August. The AEJ is known to be in approximate thermal-wind balance with the thermodynamic contrasts from the Guinea coast to the Sahara (Burpee 1972; Thorncroft and Blackburn 1999; Cook 1999; Parker *et al.* 2005) and is, therefore, a basic part of the momentum balance of the WAM circulation. Apart from its role (and origin) in the balance of the WAM thermodynamic state, the AEJ has a meteorological impact on the patterns of rainfall in the Sahel. The lower-tropospheric easterly component of wind shear with height is favourable for the generation of intense and long-lived cumulonimbus systems (Takeda 1971; Moncrieff and Miller 1976; Thorpe *et al.* 1982; Houze and Betts 1981), which dominate the rainfall distributions of the region (Mathon *et al.* 2002) and may have a significant impact on vertical transport of momentum (Moncrieff 1992; LeMone and Moncrieff 1994).

Basic models of the maintenance of the AEJ system (notably Thorncroft and Blackburn 1999) have identified the importance of convective processes. Dry convection in the deep Saharan boundary layer and moist convection in the ITCZ region further south set up different thermodynamic profiles, and the latitudinal temperature contrasts forced by these differences lead to the thermal wind that defines the jet. The dry and moist convective processes also introduce stress terms in the momentum balance that are not well understood, and which are difficult to parametrize in models.

The ageostrophic meridional flow (see Cook 1999) is responsible for the advection of humid air at low levels into the continent, and therefore feeds the moist convection in the Sahel summer months. Burpee (1972) suggested that a thermally-direct mean circulation, with equatorward flow in the AEJ, must exist in order to explain the maintenance of the AEJ against the downgradient barotropic fluxes of zonal momentum associated with African easterly waves (AEWs). This argument, while consistent with the existence of this mean meridional flow, does not explain its physical origins. In this paper we argue that, in order to understand the meridional circulation associated with the AEJ and the WAM, it is necessary to consider the balance between convection and geostrophic effects, and this balance is most strongly modulated on diurnal timescales.

In discussing the maintenance of the AEJ, Thorncroft and Blackburn (1999) stressed the importance of the mean-state heat low over the Sahara, which contributes to the strength of the monsoon trough. In attempting to explain the meridional circulations driven by this heat low, it is possible to draw on previous work from other parts of the globe. A number of recent papers have explored the dynamics and evolution of the heat low over central Australia, and its interaction with cold fronts. Most significantly, it has been seen how the development of the nocturnal jet can be a continental-scale phenomenon (May 1995) and can introduce a strong diurnal cycle to the low-level winds associated with the heat low.

An idealized study by Racz and Smith (1999) (hereafter RS99) has been able to explain the observed cycle of low-level winds over Australia. RS99 showed that on the continental scale there is a phase shift in time between the diurnal peak heat low and the diurnal peak heat-low vortex. During the day, the heat low is closely linked to the thermal heating over the continent by hydrostatic balance, and therefore the pressure minimum occurs in the afternoon. At this time, although the low-level geostrophic wind is at a maximum, the actual wind is very weak, due to the effects of turbulent mixing in the convective boundary layer (CBL). Once the sensible heating at the surface falls in late afternoon, the turbulence diminishes rapidly and the flow is able to respond to the heat-low pressure-gradient force. At this point, the winds accelerate towards low pressure and come under the influence of the Coriolis force, essentially forming a nocturnal jet (Blackadar 1957). For a low-latitude continent, the wind speeds and the heat-low vortex are maximized in the early morning, almost exactly out of phase with the peak heat low itself. It is worth noting at this stage that Linden and Simpson (1986) have demonstrated a similar interaction between turbulent and baroclinic flows in the case with negligible Coriolis force, through laboratory investigations on gravity currents.

In contrast with the Australian monsoon, over West Africa the monsoon trough moves much further inland—some 15° or more in latitude from the Guinea coast—and, therefore, the pressure gradients over much of the Sahel are directed from the humid region in the south towards the dry north. This means that the effect of a nocturnal increase of the heat low convergence may be important to the moist advection into the Sahel. Relative to the idealized study of RS99, the Sahara–Sahel region has a smoother mean gradient of surface properties, but much higher small-scale surface variability (e.g. Taylor and Lebel 1998). The WAM is also characterized by intermittent deep-convective events, which are (quite naturally) not represented in the idealized model of RS99. There is known to be a strong diurnal cycle of deep convection over the land (e.g. Houze and Betts 1981; Duvel 1989) which adds a mesoscale diurnal forcing to the WAM. Recent studies (Diongue *et al.* 2002; Redelsperger *et al.* 2002) have highlighted the significance of scale interactions between such systems and their synoptic environment.

Some former studies from the Sahara and Sahelian region have noted the existence of the diurnal cycle in low-level winds. Farquharson (1939) analysed data at Khartoum to show that the low-level winds in the morning (0400–0700 UTC) were systematically stronger than the winds in the afternoon (1100–1500 UTC), associated with weaker wind shear in the lower troposphere in the afternoon. Similar results were obtained by Solot (1950). Hamilton and Archbold (1945) also noted the systematic diurnal changes in the winds, especially in the baroclinic zone immediately south of the intertropical front (ITF), which they attributed to mixing with higher-level winds as the CBL develops during the day. In attempting to explain the nocturnal maximum in deep-convective rainfall in the Sahel, McGarry and Reed (1978) analysed 600 m and 900 m pilot balloon winds at the four main synoptic hours. Although somewhat inconclusive due to the sparsity of observations, McGarry and Reed's (1978) results showed a diurnal cycle in the winds and corresponding divergence field, with peak convergence in late afternoon. More recently, using data obtained from HAPEX–Sahel*, Dolman *et al.* (1997) stressed the importance of advection in the CBL budget in the Sahel during the night and the early morning, as a result of the tendency of the south-westerly monsoon winds to restore themselves during the night. From nocturnal soundings at the Hamdallaye site near Niamey, they showed that cooling rates of 100 to 200 W m^{-2} occurred, and noted the likely importance of cold advection (in addition to long-wave cooling) in the budget. The emphasis placed by Dolman *et al.* (1997) on the importance of meridional advection at night, and convective turbulent processes by day, is in close accord with the models of RS99 and Linden and Simpson (1986). Unfortunately, beyond these few

* The Hydrology–Atmosphere Pilot EXperiment in the Sahel.

observational studies, analyses of the WAM have tended to focus on the mean state and on composite patterns, which eradicate diurnal variability. This lack of diurnal studies has been compounded by a lack of high-frequency upper-air data in the region.

The objectives of this paper are to describe and to explain the diurnal cycle of the WAM circulation at the continental scale and the mesoscale, using data obtained from model analyses and local observations. From this we are able to infer some fundamental dynamical characteristics of the WAM system of significance to its synoptic and climatological state and to its transport properties. Section 2 introduces the datasets used in this study. In section 3 global-model analyses, which are relatively well resolved in time and space, are used to show comprehensive estimates of the diurnal variations of the WAM at the continental scale. In section 4, ground-based observations are examined in an attempt to provide more insight on the nature of the diurnal cycle and its seasonal variation and to evaluate the conclusions drawn from the global model analysis in section 3. Such data are sparsely distributed, spatially and temporally; in section 5 high-spatial-resolution profiles obtained using dropsondes during the JET2000 campaign are used to make a near-synoptic assessment of the WAM circulation and to relate this to diurnal patterns of forcing. The final section of the paper discusses the implications of the diurnal cycle for regional climate.

2. Data sources and analysis methods

The data used in this paper are taken from the periods of two recent field campaigns in West Africa: HAPEX–Sahel in 1992, and JET2000 in August 2000. The model data used here consist of ERA40* data at 1.125° resolution for the period 1979–2001, as well as a set of special analyses and 24-hour forecasts using the ECMWF operational model (at 0.5° resolution) for the JET2000 period in August 2000. Th03 and Taylor *et al.* (2005) have indicated that the ECMWF analyses have useful skill in representing the basic synoptic structure of the AEJ as well as the diurnal and synoptic variability of low-level thermodynamic patterns. Tompkins *et al.* (2005) have evaluated the impacts of different streams of data assimilated into the ECMWF model to show that satellite-derived data and *in situ* data from the global upper-air network make major contributions to the quality of the analyses over West Africa. It is known that there are discrepancies in the variability of different model analyses at different timescales (e.g. Annamalai *et al.* 1999). In this paper we use a single model to explore the physical mechanisms of diurnal variability in the WAM, and do not attempt any model intercomparisons.

For the 1979–2001 period, the wettest and driest years have been selected for composite analysis of the model fields through analysis of the GPCP† rainfall dataset (Adler *et al.* 2003). The rainfall was averaged over August in the domain 7.5°W–17.5°E, 12–20°N, from which 1994, 1999, 1988, 1989 were the wettest years and 1984, 1990, 1983, 1986 the driest.

Upper-air data over West Africa is sparse, and radiosonde observations away from the main synoptic hours of 0000 UTC and 1200 UTC have been rare. The only high-frequency radiosoundings through a full diurnal cycle that we know to have been made in the West African region is the set of two-hourly HAPEX–Sahel soundings from 25–26 September 1992 (Dolman *et al.* 1997). In addition, the HAPEX–Sahel experiment obtained a good set of automatic-weather-station data within the degree square around Niamey (13–14°N and 2–3°E), which are employed here for analysis of the low-level

* European 40-year re-analysis—a dataset constructed by the European Centre for Medium-Range Weather Forecasts.

† Global Precipitation Climatology Project.

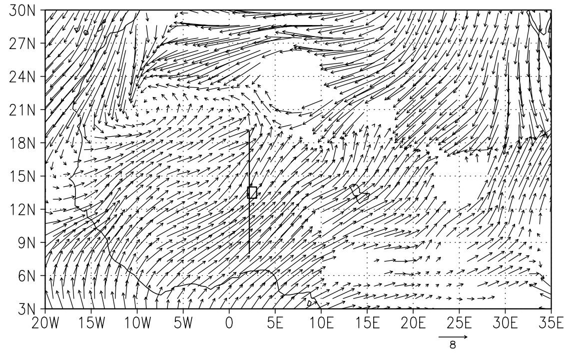

Figure 1. ERA-40 data for August 1992, showing mean winds at 925 hPa averaged over the four synoptic hours of 0000, 0600, 1200, and 1800 UTC. Blank areas are topographic regions where the mean surface pressure is below 925 hPa. The HAPEX–Sahel square (in which Niamey is located) is also marked, as is the track of the JET2000 flight described later (see Fig. 10).

structure of the diurnal cycle. Pilot-balloon data, from the HAPEX–Sahel year 1992, have been obtained for several stations in West Africa, and observations at the main synoptic hours of 0000, 0600, 1200 and 1800 UTC are included in this sample. For each time of day, missing data have been filled by linear interpolation, and a nine-point running mean has then been computed in order to filter out the effects of synoptic variability. In the Niamey area, the long-term EPSAT–Niger* network provided hourly rainfall observations; for 1992 there were over 100 stations covering an area of 1° × 1°.

Dropsonde profiles were obtained during the JET2000 experiment, which took place in late August 2000 using the Met Office C-130 Hercules aircraft (see Th03). In this paper we use observations made during the higher-resolution eastern leg of flight 2 on 28 August 2000, on which 22 dropsondes were released at approximately 0.5° intervals between 8°N and 19°N at around 2.2°E and between 0841 and 1110 UTC. Description of the data processing methods has been given by Parker *et al.* (2005).

3. THE DIURNAL CYCLE IN ERA40 DATA

(*a*) *Statistical measures from ERA40*

Figure 1 shows the mean wind field over northern Africa at 925 hPa (averaged over all four synoptic hours) over the month of August 1992, while Fig. 2 shows the average anomalies from this field at each of the synoptic hours. These images are in agreement with the basic diurnal-cycle model that has been discussed by RS99, and are broadly representative of the 1979–2001 August mean winds (not shown). For this month, the peak winds at low levels (the 'monsoon winds') occur at night, generally at 0000 UTC. These south-westerlies can be seen across Africa, between latitudes of 6°N and 18°N. By 0600 UTC the winds in the Sahel around 12–20°N have generally veered to gain a stronger westerly component and a weaker southerly component, consistent with the action of the Coriolis force in the absence of strong turbulent friction. At 1200 UTC there is a dramatic weakening of the south-westerlies in the Sahel between around 10°N and 18°N, and extending towards 21°N around the longitude 0°E, consistent with the action of convective mixing. By 1800 UTC the westerly component of the wind is further reduced in this zone. The zone of strongest diurnal variation in the mean winds is coincident with, or slightly to the south of, the zone of maximum baroclinicity at low

* Estimation des Pluies par SATellite—experience Niger (Lebel *et al.* 1992).

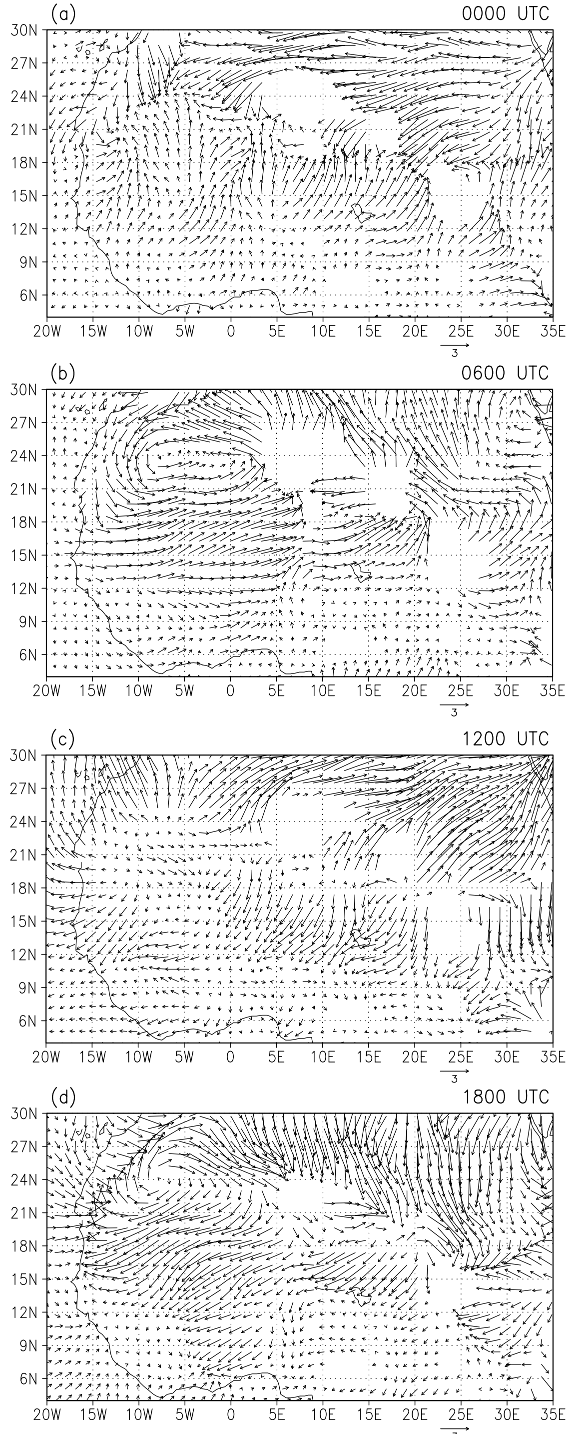

Figure 2. ERA-40 winds for August 1992, at 925 hPa, showing monthly-mean perturbations at the four synoptic hours of (a) 0000 UTC, (b) 0600 UTC, (c) 1200 UTC, and (d) 1800 UTC, relative to the mean of these four times (as seen in Fig. 1). Blank areas are topographic regions where the mean surface pressure is below 925 hPa.

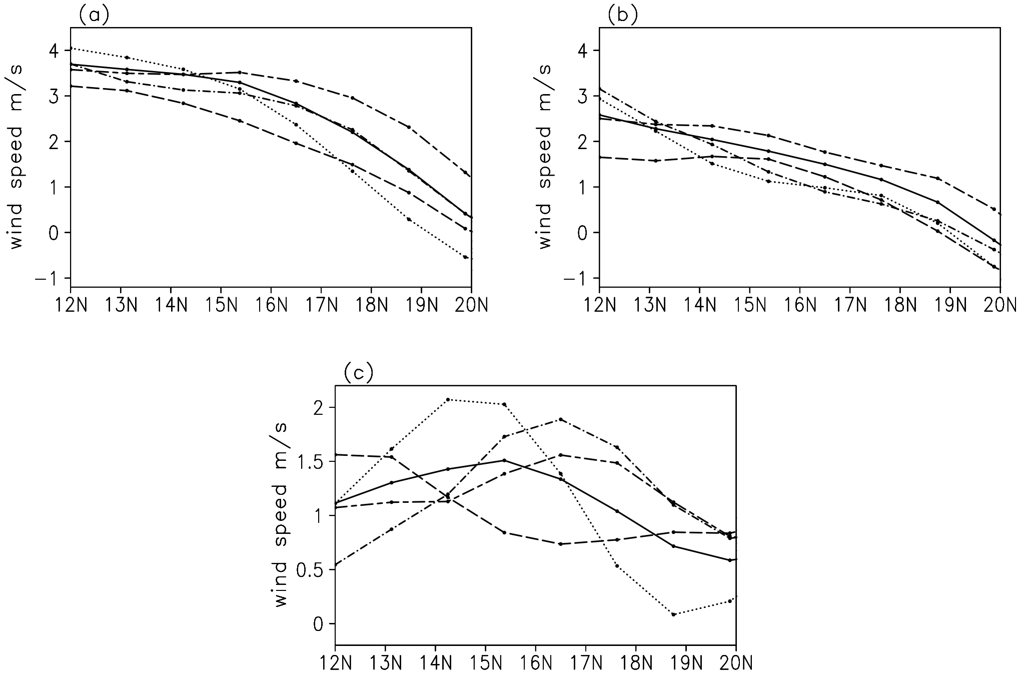

Figure 3. ERA-40 August monthly-mean data, showing meridional winds at 925 hPa averaged over longitudes 7.5°W to 17.5°E, for the 1979–2001 average (solid line), the mean of the four wettest years (long–short dashed line), the mean of the four driest years (dotted line), and the years 1992 (dot-dashed line) and 2000 (dashed line). Panel (a) shows the 0600 UTC winds, (b) the 1800 UTC winds and (c) the 0600–1800 UTC difference.

levels (not shown). Hamilton and Archbold (1945) also suggested that there is a zone immediately to the south of the ITF in which the diurnal cycle of winds is strong (see also Buckle 1996).

Replication of Figs. 1 and 2 using data from the NCEP/NCAR* re-analyses (not shown) indicates very similar patterns of diurnal variability. Indeed, the description of the diurnal cycle in the ERA-40 925 hPa winds given in this section applies equally to the NCEP/NCAR winds. A more detailed intercomparison of the diurnal cycle of these two models in the study region is beyond the scope of this paper.

Figure 3 gives a climatological perspective on this circulation. It can be seen that the mean 0600 UTC meridional winds are greater than those at 1800 UTC in the 1979–2001 mean, as well as in the composites of the four wettest and four driest years. The mean location of the peak in the amplitude of the 0600–1800 UTC difference (Fig. 3(c)) lies at 15.5°N; for the wettest years the peak is further north while for the driest years it lies further south and is stronger. This shift is consistent with a northward advance of the whole monsoon system in wetter years. The year 2000 shows a relatively low 0600–1800 UTC meridional-wind difference, apparently due to relatively low 0600 UTC winds. If we examine the four wettest and four driest years we find that they cluster distinctly (not shown). However, it should be stressed that the high spatial variability in the rainfall distribution over the continent means that there can be considerable differences between the winds analysed in years with comparable mean rainfall.

* National Centers for Environmental Prediction/National Center for Atmospheric Research.

From these six-hourly model analyses we have thus obtained a relatively crude (in temporal resolution) characterization of the diurnal cycle in the low-level winds associated with the WAM. Broadly speaking, this pattern agrees with Farquharson (1939) and Hamilton and Archbold (1945), with stronger winds in the morning than the afternoon.

(*b*) *Diurnal cycle in the ECMWF model for the JET2000 period*

Figure 4 shows a series of latitude–pressure plots from the ECMWF operational model for the period of the JET2000 experiment in late August 2000 (see Th03). We have plotted the model forecast data at the main synoptic hours, with the inclusion of 1500 UTC, which is the time of maximum sensible heating in this three-hourly dataset. It must be stressed that the peak winds seen in these model plots, away from the analysis times (1200 UTC), are the result of model internal dynamics with no observational input. It can also be noted that the analysis fields of Figs. 1, 2 and 3 have very few observational data assimilated at 0600 and 1800 UTC.

Inspection of these images in Fig. 4 presents a very clear and strong diurnal cycle in the winds and thermodynamic fields at low levels, which is in good qualitative agreement with RS99. The model graphically indicates the maintenance of meridional rather than vertical temperature and humidity gradients during the day (e.g. in the mixed layer at 1500 UTC on 27 August 2000) due to strong turbulent mixing whose influence, over the continental scale, is far stronger in the vertical than the horizontal. The model also shows that the meridional winds are very weak during the day, due to the same turbulent processes. After sunset, around 1800 UTC, the turbulence diminishes rapidly and the low-level southerly flow intensifies at a level of around 950 hPa. The southerly flow further intensifies overnight and its peak moves northward. Above the southerlies is a northerly return flow, centred around 700 hPa. With the development of this thermally-direct wind shear, the humidity field can be seen to be overturning, with cool humid air being advected northwards at low levels.

By 1500 UTC on 28 August 2000 there is a dramatic signal of deep convective activity in the model. At this longitude, deep convection was, in fact, observed much earlier on the evening of 27 August (see Taylor *et al.* 2003, hereafter Ta03); model convection is known to be inaccurate in timing and location (Yang and Slingo 2001). In the model (and in reality) deep convection is intermittent, with a return period at Niamey (13.6°N) of about four days. Figure 4 shows the evolution of the modelled monsoon circulation during a 24-hour period of no significant cumulonimbus convection in that model, until the final time-frame.

From these vertical sections, we can see rather graphically how important the diurnal cycle of winds can be in the moisture budget of the region. Nocturnal advection at low levels brings humid air into the continental interior with a dry return flow aloft; by day the low-level humidity falls as dry air from above is mixed down in the developing CBL. We can describe this process as a diurnal ‘flushing’ of the atmospheric monsoon layer, with the nocturnal surge in moisture being broken down by daytime convective mixing.

As remarked above, the model results shown here are really only indicative of internal model processes away from the analysis time of 1200 UTC. In the next two sections, observations are presented which confirm the qualitative diurnal flow patterns seen in the models.

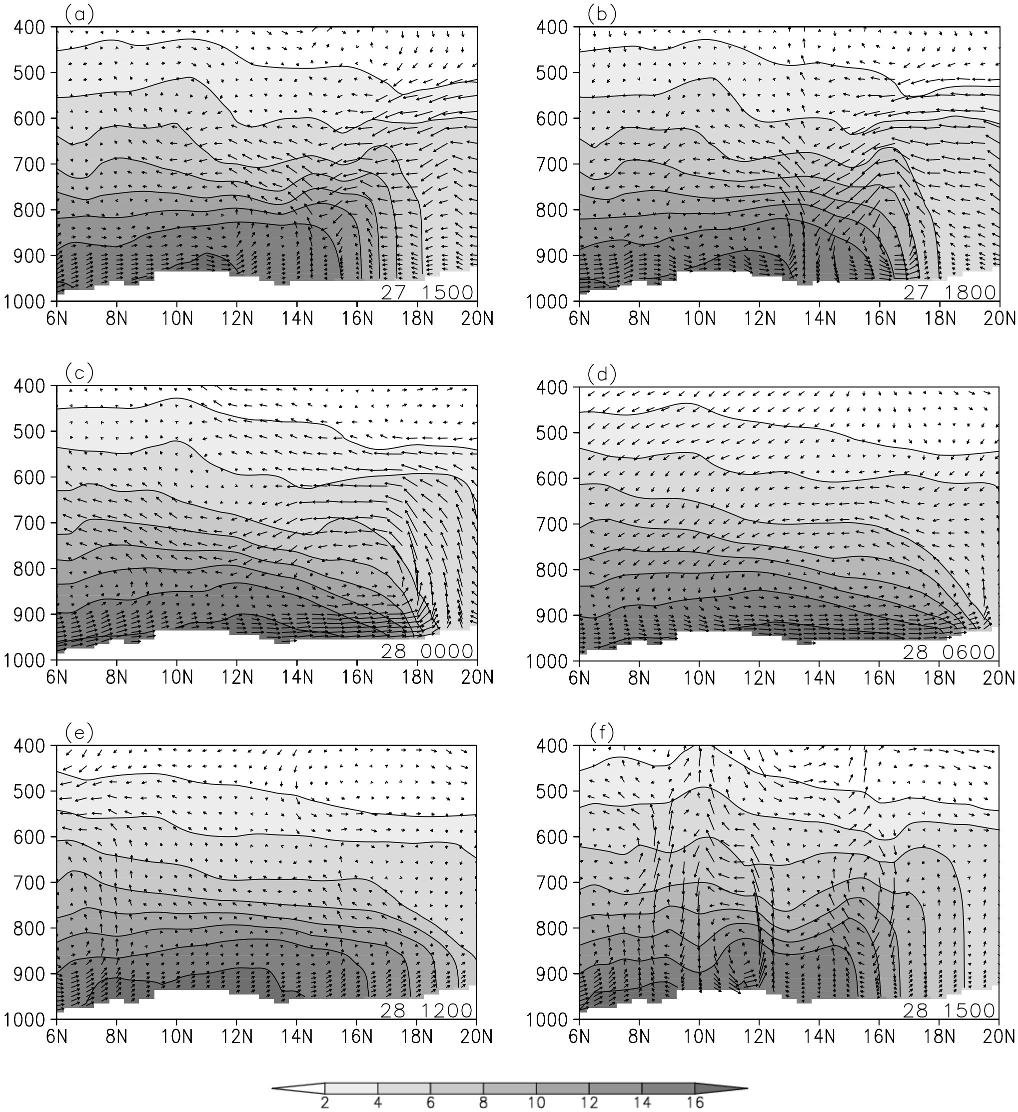

Figure 4. Plots showing the evolution of humidity mixing ratio (with a contour interval of 2 g kg^{-1}, beginning at 2 g kg^{-1}) and winds (vectors, scaled to give three-hour advection distance) in a latitude–pressure (in hPa) section from ERA40, averaged over 2.5°W–7.5°E. Fields are taken from ECMWF analyses and forecasts for the period 1500 UTC 27 August to 1500 UTC 28 August 2000; each panel is labelled with the date (in August) and time (UTC). White areas represent the topography.

4. Observations of the monsoon diurnal cycle from ground-based systems

(*a*) *Diurnal radiosonde data*

An example of the vertical structure of the diurnal cycle during the latter stage of HAPEX ten days after the last significant rain of the year in the Niamey region is shown in Fig. 5. A Hovmüller plot of the meridional wind component from this set of profiles (Fig. 5(a)) shows a relatively simple diurnal cycle which is also in excellent qualitative agreement with the model of RS99. As the CBL grows during the day (Fig. 5(b)), the vertical shear of the winds is reduced almost to zero.

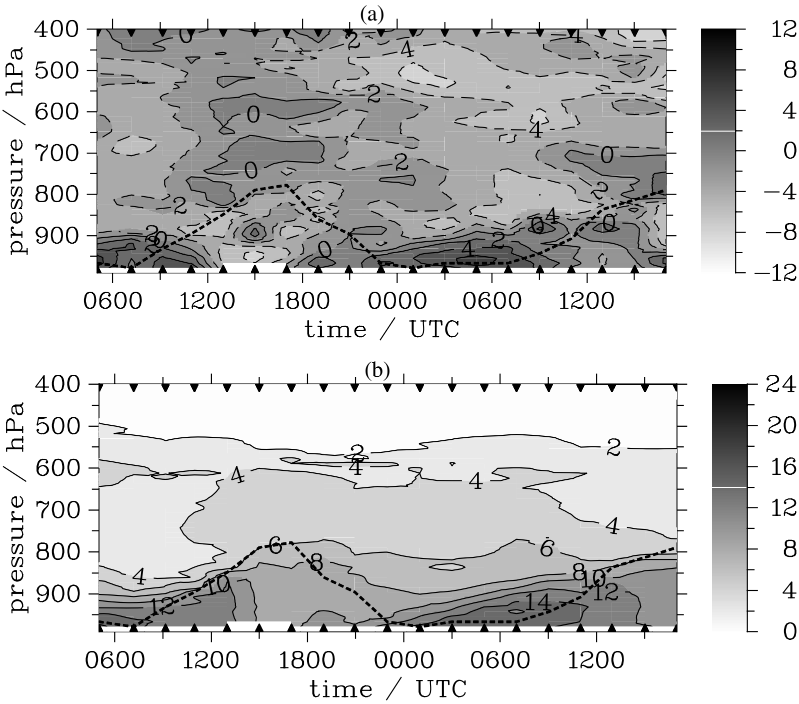

Figure 5. Time–pressure Hovmüller plots of the HAPEX radiosonde data obtained from the Hamdallaye site in the period 25–26 Sept 1992 (days 269–270), showing (a) meridional wind (contoured at 2 m s^{-1}, with negative contours dashed) and (b) humidity mixing ratio (contoured at 2 g kg^{-1}, with the top of the dry convective mixed layer denoted by a dashed line). Timings of the soundings are indicated by triangles on the time axis.

At night, when the turbulence drops, the winds begin to increase, with a low-level maximum at between 900 hPa and 950 hPa (as seen in the model results for August 2000, Fig. 4) just before 0600 UTC. The cycle in these observations is particularly striking in the meridional component, which is expected to account for most of the moisture transport in this region of strong north–south gradients. When the CBL is deep, fluctuations in the wind in this layer are probably due to convective eddies (e.g. at 1500–1700 UTC on 25 September), and generally the vertical average of these winds in the CBL is small. Note that the line of zero meridional wind, which is a monsoon layer-depth measure commonly used by forecasters to indicate the likelihood of cumulonimbus initiation and maintenance, shows significant diurnal variation, being deeper after midnight and in the early morning.

In Fig. 5(b), the increasing stratification of the boundary layer during the night is apparent. Dolman *et al.* (1997) quantified the heat and moisture budgets for these nocturnal profiles and stressed the importance of both advective and diabatic processes in the low level cooling, with the statement that ‘the cooler moister air from the south is the dominant factor in the evolution of the nocturnal boundary layer’ (as well as noting the difficulty of separating the terms in the budget with such limited data). Lhomme *et al.* (1997) used the HAPEX–Sahel daytime radiosonde data in an attempt to estimate regional energy and chemical tracer budgets. Their study found that the advection terms were important and that simple models of the advective fluxes failed as a result of diurnal and vertical variations in the winds, as has been inferred in the present paper.

In a study of the low-level jet over the Southern Great Plains of the USA, Whiteman *et al.* (1997) noted the importance of the southerly jet (which reaches its peak at night)

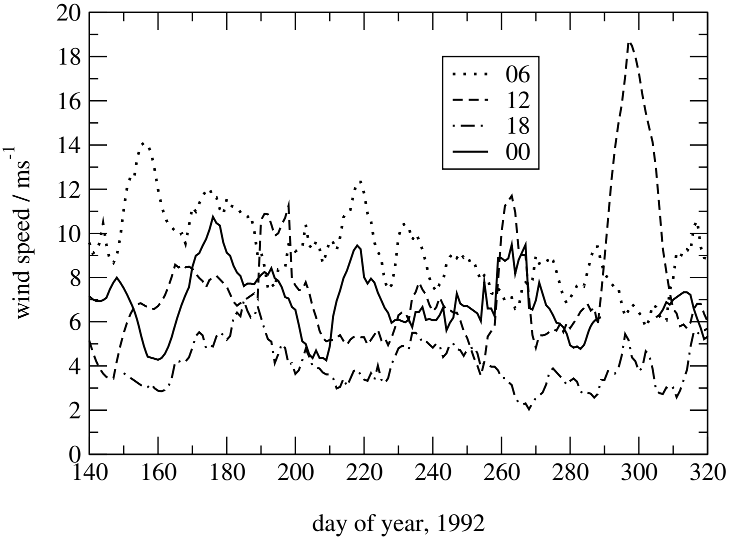

Figure 6. 600 m pilot-balloon data from Niamey in 1992, showing nine-day averages of wind speed at the four synoptic hours of 0000 UTC (solid line), 0600 UTC (dotted line), 1200 UTC (dashed line) and 1800 UTC (dash-dotted line). Note that day 214 represents 1 August 1992.

in the advection of moisture from the Gulf of Mexico. Within the WAM, we regard the diurnal cycle as having similar importance in the continental moisture budget. Humid air advected northwards by night is then mixed into the deeper monsoon layer by convection during the day. Buckle (1996) noted the fall in afternoon humidity in the zone just to the south of the ITF, and this phenomenon is also apparent in Fig. 5(b), resulting from vertical growth of the CBL. This cycle of low-level humidity variations is important in the variability of stability measures for moist convection (Williams and Renno 1993; Parker 2002), and ultimately the likelihood of rainfall.

(*b*) *Wind data from pilot balloons*

Unfortunately, this short set of HAPEX–Sahel profiles, obtained outside the important rainy season of the Sahel, is the only set of high-temporal-resolution radiosonde data available to us. However, many pilot soundings have been made over the continent at the four main synoptic hours. Although, temporally, this is a low-resolution set, strong evidence of the diurnal cycle can be seen in the data. Figure 6 shows the observations of wind magnitude made from Niamey at 600 m; in this figure, the mean winds at 0600 UTC are generally stronger than at other times of day, and always stronger than the 1800 UTC winds—as found by Farquharson (1939) in Khartoum—consistent with the effects of the cycle in boundary-layer turbulence discussed above. Note from this figure that, in using observations from 0000 and 1200 UTC, we would fail to capture the diurnal variability in the winds at this level.

In Fig. 6 there is some evidence of fluctuations in the strength of the diurnal wind variations through the season. Comparison with the EPSAT–Niger mean rainfall (see Fig. 7(d)) suggests that there is no simple relationship between the rainfall and the diurnal wind fluctuations. However, comparison of surface temperature estimates further north into the heat low shows a much more suggestive relationship. Figure 7(a) shows the meridional wind components at 0600 and 1800 UTC from the pilot-balloon dataset (with the same interpolation and averaging procedures), and the difference

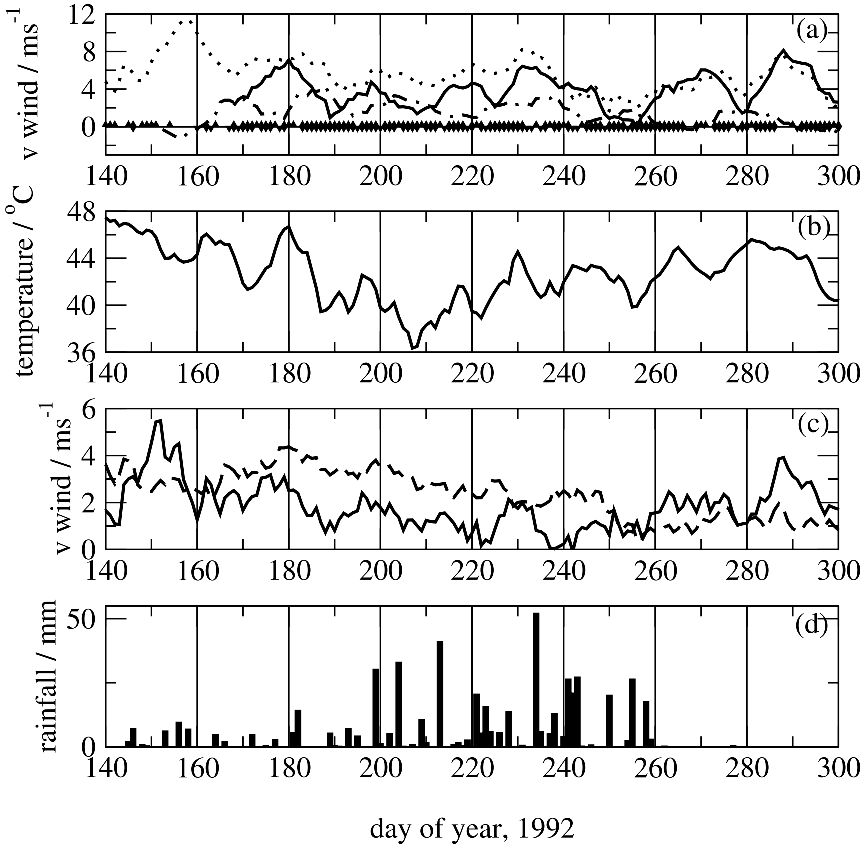

Figure 7. Nine-day averages of observed and model parameters for 1992. (a) 600 m pilot-balloon data from Niamey in 1992, showing meridional winds at 0600 UTC (dotted line) and 1800 UTC (dot-dashed line), with the 0600–1800 UTC difference (solid line). The solid triangles show the data coverage at 0600 UTC (above the axis) and 1800 UTC (below). (b) Meteosat-derived surface-temperature estimate at 21.5°N. (c) 900 hPa ERA-40 meridional wind at (2.5°E, 13.75°N), showing the 0600–1800 UTC difference (solid line) and the mean of the four main synoptic hours (dashed line). (d) The EPSAT–Niger network mean daily rainfall.

between these winds, as a measure of the amplitude of the diurnal cycle. Figure 7(b)) also shows the Meteosat-derived cloud-screened surface temperature (see Taylor *et al.* 2005), averaged between 21–22°N and 0–4°E. There is good evidence in this figure, particularly up to day 260, that the strength of the diurnal cycle is closely related to the surface temperature further north, as might be expected from the heat-low arguments, which explain the origin of the low-level flows. Indeed, when a 31-point running mean is removed from the screened temperature of Fig. 7(b), its correlation with the 0600 to 1800 UTC wind difference in Fig. 7(a) has a value of 0.57 for the period from day 168 to 260. Given the nine-point running mean on the data, there are approximately ten degrees of freedom (e.g. Jenkins and Watts 1968, p. 252), for which the 10% significance level is 0.497. After day 260 the link breaks down, and this is consistent with the southward migration of the monsoon trough after this time. In this later period the thermal and pressure gradients are systematically changing, and this changes the relevance of the 21.5°N temperature to Niamey winds; 21.5°N is then on the wrong flank of the heat low to influence Niamey winds. Finally, Fig. 7(c) shows the diurnal mean and 0600–1800 UTC difference of 900 hPa meridional wind from ERA-40 at (2.5°E, 13.75°N) averaged over a nine-day running mean for this period. Broadly speaking, the model captures the same patterns of variability seen in Fig. 7(a), and the correlation of the 0600 to 1800 UTC meridional-wind difference between model and observations is 0.57 for the period of day 168 to 300. In this case there are approximately 14 degrees of freedom and the 5% significance level is 0.532. It can also be seen that the amplitude of the diurnal cycle is often as strong as the diurnal mean, notably outside the main rainy period (before day 160 and after day 260). Furthermore, the diurnal cycle shows more variability than the mean, suggesting that variability in the monsoon circulation appears in sub-diurnal timescales.

Figures 6 and 7 also indicate the nature of the diurnally varying monsoon circulation around the time of the monsoon onset in late June (around day 180). Sultan and Janicot (2003) have described the dynamics of the changes in the monsoon circulation in this period, as the heat low and monsoon trough make a rapid northward shift over the continent. Figure 7(c) confirms that the mean meridional winds are at a maximum around the onset time; it can also be seen that these changes are associated with a maximum in the cloud-screened surface temperature (Fig. 7(b)), and a correspondingly strong amplitude in the diurnal cycle of meridional winds (Fig. 7(a)). In consequence, it seems that further understanding of the dynamics of the onset—notably the evolution of the humidity structures which control convection as described by Sultan and Janicot (2003)—should take into account the nocturnal peak in the circulation and the very different processes which control the nocturnal and daytime flows.

(*c*) *Surface wind data*

During HAPEX–Sahel, hourly surface wind data were collected at 10 m from the Banizoumbou site throughout 1992. A wavelet analysis of these data (contoured at a 5% significance level using the software of Torrence and Compo 1998) is summarized in Fig. 8. It can be seen from these data that there is a pronounced diurnal cycle through most of the year, but that the diurnal cycle is stronger during the dry season and weaker in the wet season when synoptic activity of longer than a two-day period becomes more significant.

The raw surface wind data show the characteristics of this apparent change in the diurnal cycle between the wet and dry periods. During HAPEX–Sahel, automatic surface station data were collected at 12 sites in the HAPEX square by the CNRM* group, with complete data coverage in the period 16 August to 15 October 1992. As inferred from the wavelet plot (Fig. 8), these wind data also indicate a transition in behaviour between the early and late part of the observational period. In the early wetter part of the experiment (not shown), there is no conspicuous evidence of a simple diurnal cycle in the winds, particularly in the meridional wind component. In contrast, Fig. 9(a) shows the mean of the ten-minute average winds for the 12 CNRM stations for a period during the later part of the experiment, several days after the last rainfall events (Fig. 7(d)). This later dry-period diurnal cycle agrees qualitatively with the model of RS99 and can be interpreted as:

1200–1800 UTC: low winds (even north-easterly) in the afternoon, due to turbulent mixing of momentum in the vertical, which suppresses the wind shear associated with the AEJ and the south-westerly monsoon;

* Centre National de Recherches Météorologiques.

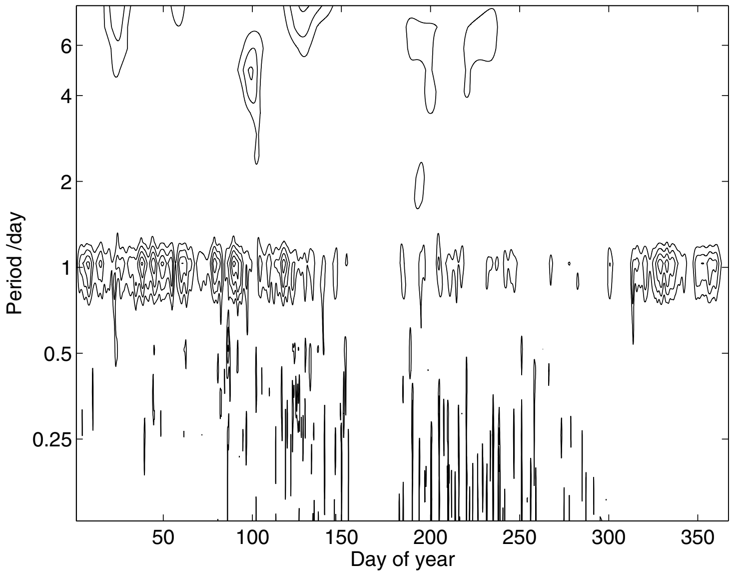

Figure 8. Results of a wavelet transform of hourly wind-speed data. The data were recorded at Banizoumbou during 1992 using a 10 m tower. Contouring is the ratio between the modulus squared of the wavelet and the values at 95% confidence (values greater than 1 are significant at 5%). Contour values are between 1 and 7.

1800 UTC–0600 UTC: increasing meridional winds after sunset as the air flows down the large-scale pressure gradient towards the monsoon trough in the north; increasing zonal winds a few hours after sunset, consistent with the influence of the Coriolis force (related to an inertial oscillation and the nocturnal jet), peaking in the early hours after midnight;

0600 UTC–1200 UTC: rapid increase in the zonal (less clearly the meridional) winds at dawn, consistent with mixing down of momentum from the elevated nocturnal jet maximum; reduction of both components in the wind towards the afternoon, as horizontal momentum becomes mixed more deeply in the vertical.

Figure 9(b) indicates the 10 m winds from ERA-40 for the same location and time-period and, although the temporal resolution is low (six hourly), the pattern is consistent with that of the measurements in Fig. 9(a).

It is particularly interesting that the diurnal cycle in surface observations involves a later peak in the winds than observed aloft in the radiosondes, due to the time taken to mix down the elevated nocturnal jet peak winds. At night the low-level flow in the surface layer (in which surface observations are made) is, to some extent, decoupled from the flow above, due to the stabilizing of the profile and the consequent reduction in turbulence. More detailed inspection of the lower-level winds from the HAPEX–Sahel radiosondes (not shown) is certainly consistent with a morning wind maximum related to mixing-down the nocturnal jet as the peak winds descend to the surface around the time of the 1101 UTC sounding. May (1995) has also noted, in association with the nocturnal jet over Australia, this effect leading to a peak in surface winds around 0900 local time suggesting also that the decoupling of the lower-level winds is affected by cloud cover, which can diminish the net long-wave radiative cooling at the surface. After the morning peak in the winds, subsequent entrainment of air with lower or negative zonal momentum serves to diminish the low-level winds later in the day.

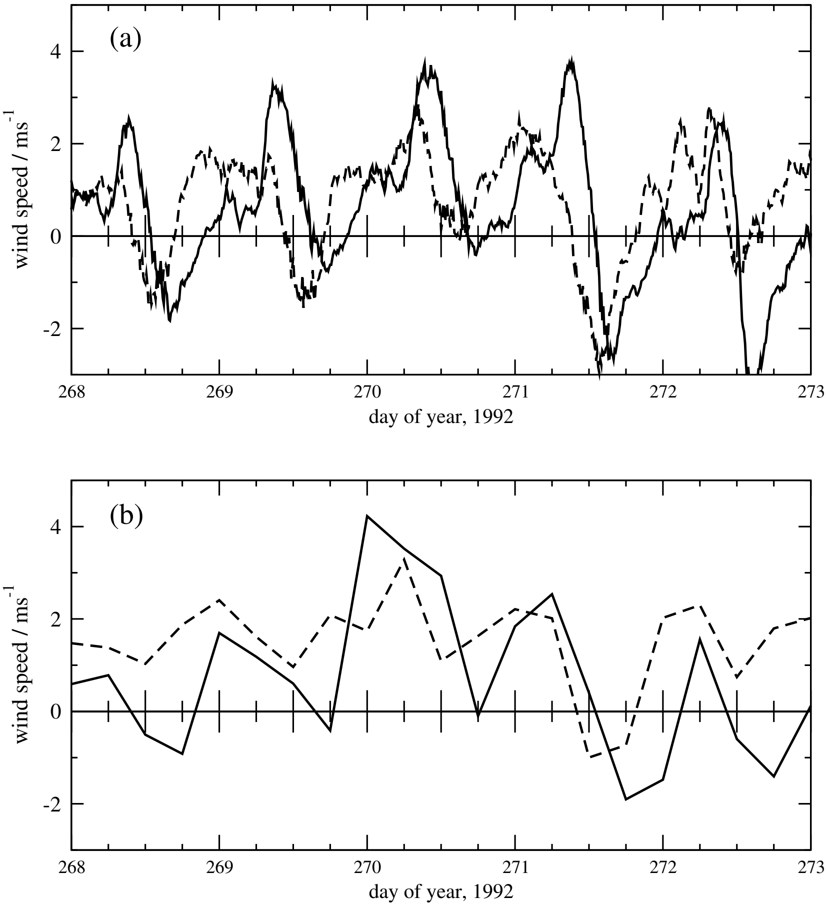

Figure 9. (a) HAPEX–Sahel ten-minute surface wind data averaged over the 12 CNRM stations and (b) six-hourly ERA40 10 m winds at (2.5°E, 13.5°N), each as a function of time in 1992, showing the zonal component (solid line) and meridional component (dashed line).

Consideration of the wind fluctuations in the rainy period (not shown) confirms that intense wind anomalies accompany rain events—such airflows are integral elements in the internal dynamics of the convective systems (e.g. Eldridge 1957; Lafore and Moncrieff 1990). Disturbance to a regular diurnal cycle by these intermittent rainfall events will serve to diminish the coherent diurnal cycle of winds otherwise observed and, in periods when rainfall is heavy, there is evidence that the diurnal cycle tends to be weak. However, there are also some periods of no measured rainfall in which a regular diurnal wind cycle is not apparent at the location of the surface measurements. We surmise that the reduction in the strength of the diurnal cycle during the wet season is not solely a result of the direct modification to the winds during convective-storm events at the observing location. The observational case study described in the next section indicates that some of this temporal variability in the strength of the diurnal cycle in the wet season may be related to mesoscale spatial variability in the wind structures due to recent rain events, and possibly to soil moisture variability.

The seasonal changes in the diurnal signal observed in the surface data may also be linked to variations in the structure of the nocturnal surface-layer inversion in which the measurements are taken. Eddy-correlation observations made during the experiment (as described by Lloyd *et al.* 1997) indicate that, at night, the stability of the air at 10 m is closely linked to factors such as cloud cover and soil moisture, which affect the magnitude of the sensible-heat flux. Variations in these factors, therefore, affect the strength of the coupling between the wind observed in the surface layer and the monsoon flow. However, our inspection of these surface-layer stability measurements (not shown) did not show a clear relationship with the strength of the diurnal cycle of winds at the surface and 600 m. Finally, it is likely that, as the baroclinic zone associated with the ITF migrates further north around the peak monsoon period, the diurnal variability may weaken at a station in the mid Sahel (e.g. Niamey). For all of these reasons, the intraseasonal variability seen in the pilot-balloon data (Fig. 7) is not identified in observations made within the surface layer. The reasons for the reduction in the diurnal cycle in surface-based wind measurements during the wet season can probably only be established by observations at much higher spatial and temporal resolution, or by detailed high-resolution modelling studies.

5. Mesoscale variability in the monsoon diurnal cycle

The period of the JET2000 observations, in late August 2000, appears to have been one of anomalously low rainfall in the Sahel, and certainly in the vicinity of Niamey (Th03; Ta03). Both the radiosonde and the numerical weather-prediction analysis data suggest weak AEW activity during this period and, as a consequence, synoptic variability may be seen to be less significant in the observational period than variability due to mesoscale convective systems (MCSs). Ta03 have described the recent history of deep convection and rainfall in the region of the JET2000 box flights. In short, the boundary layer overflown by the aircraft was characterized by alternate wet and dry land surfaces, and Ta03 have been able to relate these to independent measures of the surface and CBL characteristics—surface temperature, turbulence as measured *in situ* by the aircraft, and shallow convection.

Equivalent potential temperature, θ_e, has been derived from the set of dropsonde profiles, and is plotted in Fig. 10(a) as a latitude–pressure section. As discussed by Parker *et al.* (2005), we can describe the low-level structure in the southern part of this domain as the 'monsoon layer' in which there is a CBL growing from below, and we here define the top of the monsoon layer (and the lower limit of the Saharan air layer, or 'SAL') as a contour of constant virtual potential temperature, $\theta_v = 313$ K. It can be seen that, at the times of the observations (0842–1110 UTC), the CBL is shallow (dashed line in Fig. 10), and the air above this can be said to be in the zone of nocturnal circulation (that is, not yet directly influenced by the day's convective turbulence). In the lower levels there is, broadly speaking, a northerly component of shear with height (Fig. 10(b)), with the peak northerlies along the top of the monsoon layer. Southerly flow is confined to a layer below 900 hPa except in a zone around 12–14.5°N where light southerlies extend up to 800–850 hPa.

Closer inspection of the patterns in Fig. 10 suggests the occurrence of characteristic shallow mesoscale circulations on horizontal scales of 50–200 km and vertical depths of 2 km. The two most prominent examples of such circulations are labelled A and C. These circulations appear to be thermally direct in terms of the mean-state baroclinicity of the monsoon layer, with air ascending from the north over cooler descending air from the south.

Although the A and C circulations are apparently above the convective layer, it can be said that they consist of air that was within the CBL and shallow cumulus layer at

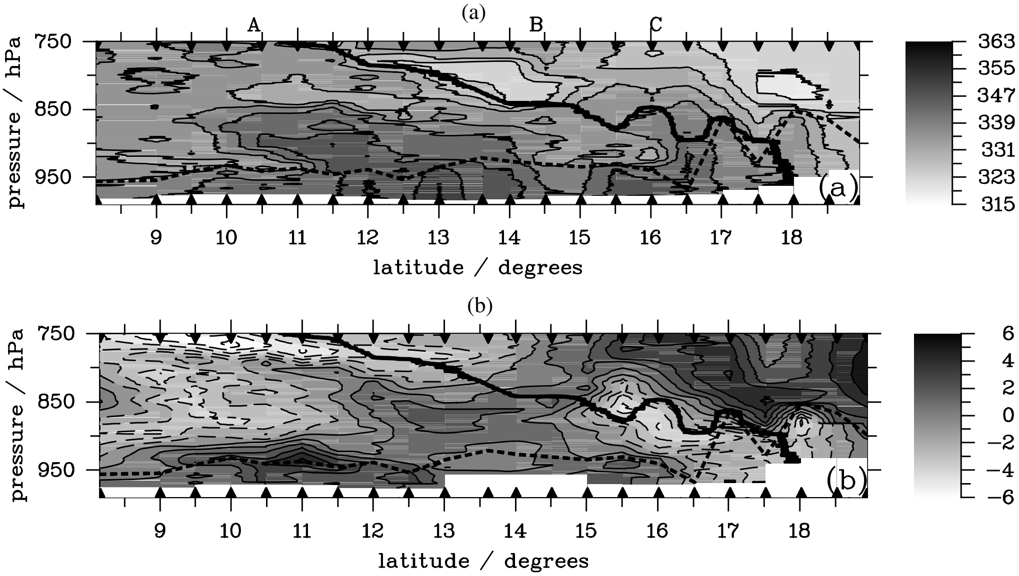

Figure 10. JET2000 sections showing overturning circulations around the 975 hPa level. In (a), equivalent potential temperature is shaded at 4 K intervals, while in (b), meridional wind is contoured at 1 m s^{-1}, with negative contours dashed. The thick dashed line is an estimate of the convective mixed-layer top, defined as the level where the temperature of lifted low-level air (10 hPa above the surface) becomes at least 1 K cooler than its environment. The thick solid line represents the contour of virtual potential temperature equal to 313 K. Triangles at the upper and lower axes mark the dropsonde positions.

the end of the previous day. This conclusion is based on observations of the CBL depth during the western leg of this flight later in the day, and on the relatively uniform profiles of saturated equivalent potential temperature (see also Parker *et al.* 2005).

It is also possible to suggest a reversed circulation at B, with air at 900 hPa moving from the south over drier air at 920 hPa that is moving from the north, relatively. The superposition of this circulation on the larger-scale monsoon flow explains the break in the northerlies at the top of the monsoon layer at this latitude. A simple estimate of the timescale of the A, B and C overturning circulations (taking velocity anomalies of 6–8 m s^{-1} and horizontal scales of 100–200 km) gives values of 4–7 hours. If the motion has accelerated from rest under a constant pressure gradient, we might double this estimate to 8–14 hours. Thus, the A and C flows seem to represent strong nocturnal circulations—locally enhanced monsoon circulations—while the monsoon flow at B is suppressed or even reversed.

All the dropsondes between 14.5°N and 16°N show a layer of dry air between 900 hPa and 925 hPa, and there is evidence of more humid air moving over this layer from the north and the south of this band. Ta03 showed that a band of organized convection passed through this range of latitudes at around 1500–1700 UTC the previous evening; such systems tend to inject dry low-θ_e air into the planetary boundary layer (PBL) through the occurrence of downdraughts (as was apparent in the aftermath of a convective event the following day—see Th03). The layer of dry air around 900 to 925 hPa seems coherent over this range of dropsondes, and coincident with the broad MCS pattern inferred from cloud-top temperature, despite the narrowness of the satellite-inferred rainfall structures at the surface (Ta03). Our basic hypothesis is that the A, B and C circulations represent an adjustment from horizontal inhomogeneities present in the previous day's boundary layer, caused by the deep-convective events which happened at around 1600 UTC the previous day (i.e. 18 hours previously). These convective storms would have injected cool, low-θ_e air into the PBL as well as wetting the land surface and thereby modifying the patterns of surface fluxes.

Overall, these dropsonde observations, along with the analysis of Ta03, suggest significant mesoscale modification to the coarse-resolution pattern of the diurnal cycle seen in the global model fields of section 3. During the day, the CBL properties seem to map closely onto the highly complex patterns of surface moisture which are observed in the Sahel, with relatively weak influence of advection, on scales down to 10 km or smaller (Ta03). At nightfall, with the decay of turbulent mixing, the flow can respond much more strongly to the local horizontal pressure gradients. To the north of a cool anomaly the continental-scale gradients are enhanced (Fig. 3 of Ta03) and a strong nocturnal circulation such as A and C (Fig. 10) occurs; to the south the continental-scale gradients are weakened or even reversed, leading to weak or reversed nocturnal circulation, as at B (Fig. 10). A similar diurnal cycle of surface-forced mesoscale flow has been described, in regard to the sea-breeze circulation, by Reible *et al.* (1993). The pattern of mesoscale control observed in Fig. 10 is further complicated by the occurrence of convective downdraughts, which make a dramatic local impact on the boundary-layer structure. Modelling studies are currently being performed to assess the relative importance of the direct profile modification by the storms and the impact of differential surface heating due to patterns of soil moisture resulting from the storms' rainfall in creating the anomalous profiles, and hence circulations, at A, B and C.

6. DISCUSSION AND CONCLUSIONS

The analysis presented in this paper has highlighted the marked diurnal cycle of lower-tropospheric monsoon circulations and boundary-layer mixing that characterizes the baroclinic region over West Africa. The basic pattern of this cycle is summarized in section 3(a) above, and in the model data of Fig. 4, and is driven by the interplay between the diurnal cycle of the heat low over the Sahara (which drives the circulation) and the diurnal cycle of boundary-layer convective turbulence (which tends to suppress the circulation). This coherent diurnal cycle of transport and mixing is a key process that influences the continental water budget over West Africa (as well as the budgets of trace gases and aerosol). During the day, when convection is active in the CBL, water vapour is well-mixed in the vertical while meridional circulations are suppressed. At night, when the convection in the CBL and associated turbulence becomes far weaker, a circulation develops in which horizontal advection becomes more important and vertical mixing is weak, consistent with the development of vertical stratification. Moist air is advected polewards at low levels in association with an acceleration towards low pressure and the development of the nocturnal jet. A return flow occurs above the nocturnal jet, following the baroclinic variation of the pressure gradient with height. Through this diurnal cycle, air close to the surface can be lifted in the daytime mixed layer and then lifted further by isentropic upgliding at night; inversely, mid-tropospheric air may be taken to the surface over diurnal timescales. It is in this light that Parker *et al.* (2005) have stressed the role of the near-adiabatic layer above the monsoon layer (the SAL) as a zone where air may exchange between the free troposphere and the diurnally controlled CBL and monsoon layer.

The observation that the monsoon-layer circulations have a coherent and significant diurnal cycle raises important issues regarding the way in which time-averaged analyses of the WAM are conducted. If such averages do not include the nocturnal hours, they will not take account of the peak meridional circulations over the continent. Of even greater concern for the quality of such analyses is the fact that the main synoptic hour of upper-air observation over the region—1200 UTC—is close to a node of the diurnal cycle. Observations at 1200 UTC fail to capture the peak monsoon flow; indeed, we would like to suggest that observations at 0600 UTC and 1800 UTC would be of more benefit to understanding and analysing the WAM than most of the currently available upper-air data.

We regard the coherent diurnal cycle seen in our analysis as a fundamental mode of the atmospheric circulation in the WAM. Indeed, since arguably the 'essence' of the ageostrophic meridional circulation within the large-scale WAM is captured at these timescales, the diurnal cycle of winds could be regarded as a key test of the fidelity of general-circulation models (GCMs) used for weather and climate prediction. It is tempting to speculate that improved simulations of the diurnal cycle of WAM circulations could lead to similar improvements in simulations of the annual cycle, but this needs further investigation.

In the JET2000 observations, we are led to the conclusion that, although the overall monsoon layer observations support the continental-scale pattern of diurnal circulation, there is also considerable mesoscale structure within the monsoon circulation. Local pressure gradients associated with mesoscale variations in PBL properties may dominate over the continental-scale pressure gradients, so that the diurnal cycle of winds is controlled by the local rather than the continental gradients. The mesoscale structure of nocturnal circulations (labelled A, B and C in Fig. 10) is quite naturally absent from the GCM fields, since the convective parametrization is not expected to resolve the individual convective systems which we believe have led to the local flows. In this sense, the diurnal cycle seen in GCMs may be more spatially coherent than that which occurs in reality. If this is the case, it is possible that, by lacking the impact of mesoscale circulation patterns, the global models may transport moisture too far north in the nocturnal CBL. Although the link between the continental-scale transport and the mesoscale circulations is complicated by the nature of mixing in the models (parametrized or numerical), this is one possible explanation for the cool and humid bias in the northern Sahel exhibited by some GCMs (e.g. Th03) and should be investigated.

It should be recognized that the importance of the diurnal cycle of the WAM circulations varies with the surface temperature of the heat low to the north (Fig. 7). One time at which this variation may be particularly important is the period leading up to the monsoon onset. Sultan and Janicot (2000) have suggested that the onset may be a rapid event on the continental scale, occurring with some regularity in late June close to the solstice, and they have argued that the heat-low circulation dominates the monsoon in this period (Sultan and Janicot 2003). There has also been notable interest in the development of the cold tongue in the Atlantic Ocean in determining the onset of the WAM. We would point out here that the onset period is also a time of dry land surface in the Sahel and, therefore, a time of regular and robust diurnal cycle (confirmed in the two years of pilot-balloon wind data available to us) during which nocturnal advection of moisture into the continent at low levels may be at its peak. As suggested by Sultan and Janicot (2003), future work on the onset of the WAM should consider both the role of SSTs and the heat low.

There are many questions outstanding from this study, but the primary constraint in understanding the dynamics of the WAM is the lack of observational data. We have argued that upper-air observations at 0600 and 1800 UTC would ameliorate this problem, but there is also a real need for higher spatial and temporal resolution upper-air and land-surface data. The JET2000 results have shown from one synoptic case that coherent mesoscale circulations occur within the monsoon flows, and that these circulations are related to recent rainfall events. If this is generally the case, such circulations may have a great impact on the continental-scale monsoon circulation. In order to explore this hypothesis, we need to observe the atmospheric flows on the mesoscale and the synoptic scale, but also with systematically increased temporal resolution, including observations made during the night (or at least shortly after sunrise) and for several days after significant rainfall events. We hope that the one benefit of the forthcoming AMMA* programme will be better observations of this kind.

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