Variability in boundary layer structure during HAPEX-Sahel wet–dry season transition

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Abstract

The variability of the Sahelian boundary layer has been studied with streamline analyses, rainfall measurements, and upper air soundings during its transition from wet to dry season. The 1992 rainy season ended prematurely because of the early arrival of westerly troughs over West Africa. The change in the circulation is related to global-scale atmospheric circulations as successive westerly troughs over this region can be traced back upstream on a planetary scale. Once the upper level easterlies changed to westerlies, the large-scale circulation brought the surface northeasterly flow southward, which led to the retreat of the Southwest Monsoon in Niger. The boundary layer responded quickly to this transition of synoptic events from wet to dry seasons. During the wet period, the boundary layer was relatively cool and moist because evapotranspiration dominated, keeping the surface cool and preventing significant direct sensible heating of the boundary layer. During the transition period, less extensive showery weather allowed the boundary layer to have more time to recover from rainfall episodes, leading to a warming and drying trend. During the dry period, soil moisture contents dropped rapidly. With more sensible heat flux made available for boundary layer heating and energetics, the boundary layer reached its maximum temperatures and minimum moisture contents during the Hydrological Atmospheric Pilot Experiment in the Sahel (HAPEX-Sahel) intensive observational period. Budget calculations indicate that the horizontal advection and vertical flux divergence terms were most important during the wet period, whereas during the dry period, the subsidence and vertical flux divergence terms were most important. From

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wet to dry seasons, the vertical wind shear of the zonal wind was reduced from 23 m s\(^{-1}\) to 16 m s\(^{-1}\), consistent with vertical wind shear differences between wet and dry years as reported in the literature. Similarities and differences with the First ISLSCP Field Experiment (FIFE-89) boundary layer are also examined. It is hypothesized that the retreat of the southwesterly monsoon could be upheld by a sustained secondary circulation if the wet season rainfall pattern imprints an organized south to north soil moisture gradient maintaining a concurrent reverse gradient in surface sensible heating. The boundary layer circulation that would be established in response to the heating gradient would reinforce surface southwesterlies, as well as reinforcing mid-level easterlies of which the African Easterly Jet is a part, and thus help uphold the intrusion of westerlies and the monsoon retreat. Such a mechanism, whose effectiveness would be a function of how distinct the south–north soil moisture gradient develops from the wet season precipitation pattern, could help explain the large interannual variability of rainfall over the Sahel.

### 1. Introduction

Over recent decades, much of West Africa has experienced dramatic climatic variability in which changes in land surface processes are thought to be partly responsible for the fluctuations. Since the land surface is a sub-system of the planetary boundary layer, the variable nature of boundary layer dynamics and its motions in response to changing surface energy fluxes, as well as feedbacks, are inherent to a full understanding of how land surface changes influence climate. In this investigation we examine the variability and interrelationships of the surface energy budget, planetary boundary layer (PBL) depth, and shear instability over the Hydrological Atmospheric Pilot Experiment in the Sahel (HAPEX-Sahel) study-area during the intensive observing period (IOP) which took place from August 15 to October 9, 1992. Midway during this period, a dramatic shift in surface conditions took place in response to the large-scale transition from wet to dry seasons. Our principle objective is to understand how boundary layer structure and energetics within the study area responds to this transition. The 1992 rainy season in the West African Sahel region concluded with below normal rainfall for the fifth consecutive year (Halpert et al., 1993), continuing a long-term decline in rainfall since 1965 (Sivakumar, 1992). At one time, the decline in Sahelian rainfall had been linked to the local effect of desertification and its associated higher surface albedos; see Charney (1975) and Charney et al. (1977). However, analysis of historical records of Sahelian rainfall (Nicholson, 1989; LeBarbé and Lebel, 1997), and accumulated studies of atmospheric circulations (Kidson, 1977; Kanamitsu and Krishnamurti, 1978; Newell and Kidson, 1984) and sea surface temperature anomalies (Folland et al., 1986; Owen and Ward, 1989) now indicate that desertification is unlikely to be the direct cause of long-term decline in rainfall.

To uncover the mechanisms that explain the observed rainfall pattern in this semi-arid region, it is necessary to examine the hydrological process and its relation to such mechanisms. Wallace (1994) proposed a conceptual framework which explains dryland degradation by a combination of (1) anthropogenic, (2) land surface–atmosphere feedback, (3) hydrological cycle, and (4) external climatic change factors. Each mechanism interacts with the others through feedback links. For instance, over-cultivation and deforestation by humans, and overgrazing by animals will initiate the degradation process by reducing vegetation (Mechanism 1). The loss of vegetation results in increased albedo and less evapotranspiration (Mechanism 2), which may result in a reduction of local convective rainfall and thus to decreased soil moisture storage (Mechanism 3). The loss of vegetation also increases runoff and erosion which is another cause of reduced soil moisture storage. Decreased soil moisture will produce less vegetation. Consequently, there is additional movement of soils by wind and water. Suspended dust particles in the atmosphere will then change the planetary albedo and rainfall chemistry. The land surface changes will ultimately force atmospheric circulation changes beginning with the planetary boundary layer (PBL) and then the free atmosphere.

Superimposed on the internally forced changes, are external forcings from such processes as the impacts of increased concentrations of atmospheric greenhouse gases, sustained sea surface temperature anomalies, and large-scale deforestation operations (Mechanism 4). Eventually, altering the atmospheric circulation can change the basic patterns of heat and moisture transport from the tropics, which can lead to such phenomena as the delayed northward progression of the intertropical convergence zone (ITCZ), or its premature withdrawal during the summer monsoon (Lamb, 1978; Motha et al., 1980). Thus, part of the process of understanding the global significance of long-term decline in Sahelian rainfall is to understand its influence on boundary layer dynamics and atmospheric circulations.

From Wallace's conceptual model, it is evident that the four mechanisms can work together to produce higher surface albedos, warmer surface temperatures, lower soil moisture, and less vegetation in the Sahel after a drought is in place. Therefore, it has been debated whether the deterioration of Sahelian land resources has been initiated by poor land management practices or by the changing atmospheric circulations driven by external phenomena or internal coupled processes (Hastenrath, 1985). Sensitivity studies of the Sahelian climate with global numerical models indicate that the predictions are quite sensitive to how surface processes are parameterized (Cunnington and Rowntree, 1986; Bounoua and Krishnamurti, 1993). By the same token, deficiencies have been noted in the parameterization schemes of soil moisture (Bounoua and Krishnamurti, 1993; Betts et al., 1993), surface sensible heat fluxes (Beljaars and Holtslag, 1991; Wai and Smith, 1995), and evaporative fluxes (Dolman et al., 1993), so the overall validity of the sensitivity studies remains in question. Several observational studies of the surface energy balance in the Sahel (Durand et al., 1988; Wallace et al., 1990; Gash et al., 1991) and its surface temperature (Lloyd et al., 1992) have provided insight into how these surface processes. However, the data from these studies were not gathered on space and time scales inherent to GCM models, which makes it difficult to apply such data to help improve the formulation and calibration of the parameterization schemes within the models. Therefore, there is an ongoing need for measurements that describe accurately the energy and water balance of the Sahelian region over space and time scales comparable to those intrinsic to GCM scales.

HAPEX-Sahel (HS), which was conducted over a 1° square study-area surrounding Niger, West Africa, offered an improved data set to tackle some of the modeling problems. The original scientific objectives of this experiment were to examine processes...
and feedbacks within the hydrological cycle and the land–atmosphere energy exchange relationships over various space and time scales, to better understand the global implications of surface changes that have and continue to take place within vast areas of the Sahelian zone. Goutorbe et al. (1994) and Prince et al. (1995) have highlighted the main biophysical components of the HS study-area, the relevant space and time sampling strategies, and the four main measuring programs involving (1) precipitation data, (2) surface meteorological flux data, (3) upper air data, and (4) remotely sensed aircraft and satellite data. These data sets are essential to help understand the relationships between changes in the surface energy balance and the atmospheric circulations.

During the HS IOP, there was a dramatic shift in surface conditions in response to the large-scale transition from wet to dry seasons. The main goal of this study is to understand how the PBL structure and budgets of temperature and moisture within the study-area changed in response to this transition. In particular, we are interested in examining the variability and interrelationship of the depth, shear instability, advection, and turbulence heating taking place within the Sahelian PBL. Section 2 briefly discusses the data sets used in the analysis. Section 3 describes the precipitation pattern during the IOP in conjunction with the synoptic forcing, followed in Section 4 by an analysis of the response of the boundary layer to the transition of synoptic events. A discussion of the budgets of potential temperature and mixing ratio is given in Section 5. A comparison of the HAPEX-Sahel PBL to that found over the study-area of central Kansas based on the somewhat similar First ISLSCP Field Experiment (FIFE) is given in Section 6. The conclusions are presented in Section 7.

2. Data sets

Three types of data sets have been used for this study. The first is US National Hurricane Center streamline analysis at the surface, 850, and 200 mbar used to obtain an overview of the synoptic situations during the IOP. The second is rainfall obtained from the EPSAT-Niger raingage network (see Lebel et al., 1992, 1996, 1997). These data are used to verify and characterize the transition of synoptic events during the IOP. Rainfall measurements from six of the sites, which are located with or very near to the six HS radiosonde sites, have been analyzed individually. These consist of the raingage stations at: (1) Danguey Gorou, (2) Niamey Airport, (3) Gagare, (4) II Jacher, (5) Gourmarcy, and (6) II Plateau. All are within 1.5 km of a radiosonde launch site except the Gagare station, which is 8.2 km from Hamdallaye (the site of the CNRM station). The raingage–radiosonde station layout is shown in Fig. 1.

The third type of data set is the upper air sounding used to examine the boundary layer structure within the study-area, and the budgets of the potential temperature and mixing ratio. The upper air data were obtained from the Niger government’s radiosonde station at the Niamey Airport (DMN), from the high resolution CNRM station at Hamdallaye, and from the Institute of Hydrology’s (IH) mobile system operated at one of four sites: Danguey Gorou (DG), Dallol Bosso (DB), Southern Site Fallow (SSF), and Southern Site Tiger Bush (SST). Fig. 1 shows the relationship between these sites to the six raingage sites identified above.

In the following we provide a brief discussion of the upper air data; Bergue and Bessemoulin (1993) provide a more comprehensive description of the radiosonde instrumentation and sampling intervals for the three radiosonde systems. The local meteorological service at the Niamey Airport (DMN) routinely launches two radiosoundings (~10.30 and 24.00 GMT) in compliance with the rules of the World Meteorological Organization. DMN collected 77 soundings during the IOP. During the daytime and also depending on the weather conditions, the Centre National de Recherches Meteorologiques (CNRM) launched five to seven radiosondes at Hamdallaye, which is located about 20 km northeast of the Niamey Airport. On rainy days, the launch schedule was either canceled or reduced to one sounding. On September 25 and 26, 19 radiosondes were launched at a 2-h interval beginning at 05.00 GMT on September 25 and ending at 17.00 GMT on September 26. Altogether CNRM obtained 159 soundings. By operating their radiosonde system on a similar schedule to CNRM, the Institute of Hydrology (IH) launched its daily radiosondes at one of four sites: Danguey Gorou (DG), Dallol Bosso (DB), Southern Site Fallow (SSF), and Southern Site Tiger Bush (SST). IH obtained a total of 83 soundings.

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**Fig. 1. Schematic layout of the six radiosonde–raingage stations within the HAPEX-Sahel study-area.** The Niamey Airport (DMN) radiosonde station is a permanent Niger government facility. The Hamdallaye radiosonde used the portable high resolution system from CNRM. The remaining four radiosonde sites at Danguey Gorou (DG), Dallol Bosso (DB), Southern Site Fallow (SSF), and Southern Site Tiger Bush (SST), used the IH mobile sounding system. The raingage stations at Danguey Gorou, Niamey Airport, Gagare, II Jacher, Gourmarcy, and II Plateau were colocated or very near the radiosonde sites.
throughout the course of the IOP. On any given launch day, three radiosonde sites were active, DMN, CNRM, and the IH mobile system at one of four places.

Upper air measurements normally include measurements of pressure, temperature, humidity, plus wind speed and direction. The soundings from both DMN and IH did not include the wind speed and direction in the HS database because of tracking problems with the theodolites. By the same token, CNRM and IH provided additional thermodynamic and wet variables such as potential temperature, mixing ratio, and dew point temperature. The sampling rates and characteristic pressure spacing between samples for the DMN, CNRM, and IH soundings ranged from 5 to 10 s and 2 to 10 mb.

In this study we have used potential temperature, mixing ratio, and wind speed-direction data. Prior to the PBL analysis and budget calculations, we checked the consistency of the wind velocity, temperature, and mixing ratio profiles. We removed any spikes in the soundings and replaced unrealistic values of temperature and mixing ratio by linear interpolation. Boundary layer top heights were first determined objectively by characteristics of the vertical potential temperature gradient (nominally, at least a 2°C increase in θ), then verified with a corresponding increase in dew point depression and evidence of positive growth tendency of boundary layer depth during the day. A verification-correction procedure was necessary because elevated inversion layers persisted throughout the last portion of the IOP, and mixing of air parcels at the internal boundary layer top plus vertical smoothing of the temperature profiles during data reduction tended to reduce the potential temperature gradient across the boundary layer top. Therefore, use of the vertical potential temperature gradient test alone would have mislocated the actual boundary layer top on several occasions, resulting in non-physical intermittent growth of boundary layer depth. Once a boundary layer height (Zb) was determined, we then calculated surface and mixed layer wind velocities (Vw, Vm), potential temperatures (θw, θm), and mixing ratios (qw, qa), along with the vertical shear difference (ΔU) between the maximum values of boundary layer zonal wind and upper level easterly wind. For purposes of these calculations, the surface layer was taken as 10% of Zb.

Because of the presence of the African Easterly Jet (AEJ) and the Tropical Easterly Jet (TEJ) in the troposphere, we also examine stability in strong shear regions on specific days by calculating profiles of bulk Richardson number, taken as:

$$R_{b}(z) = \frac{\left(g/\bar{\theta}\right)(\Delta \bar{\theta}(z) \Delta u(z))}{(\Delta u(z)^2 + \Delta \theta(z)^2)}$$

where g is the gravitational acceleration constant, Δu, Δθ, and Δz are taken over the high resolution radiosonde levels from the CNRM soundings, and θ indicates a layer.
average. We examine the $R_b$ profiles for September 25 and 26 in the discussion on PBL structure during the dry period, because the high temporal resolution of the soundings on these days provides a means to follow the diurnal evolution of boundary layer properties.

3. Precipitation pattern

Fig. 2 shows the daily accumulated rainfall time series over the HS study-area during the course of the IOP based on all reporting gages. The rainfall sequence indicates a wet period from August 21 to September 15, followed by a dry period from September 16 to October 12. Fig. 3 shows the individual daily station accumulation sequences from the six raingages collocated with the radiosonde sites. This diagram reveals a gradual, southward migration of the time of changeover from wet to dry conditions (the locations of the stations in this diagram are arranged from north to south and from west to east). In this regard, it is more reasonable to consider the latter part of the wet period as a transition period. In this division, the redefined wet period is the interval of steady and extensive rainfall which took place between August 21 and August 30. The transition period is the interval of less extensive rainfall (showers) which took place between August 31 and September 15. Between August 31 and September 7, more rain fell in the northern part

Fig. 4. Boundary layer structure at Hamdallaye during HAPEx-Sahel IOP: (a) surface (bottom) and mixed layer (top) wind velocities; (b) surface (S) and mixed layer (M) mixing ratios; (c) surface (S) and mixed layer (M) potential temperatures; and (d) boundary layer height (wide bars) and region of wind shear layer (narrow bars).
of the study-area; between September 10 and 15, more rain fell in the south. The southward shift in the rainfall pattern was related to the premature withdrawal of the ITCZ in Niger (Halpert and Ropelewski, 1992). The monsoon rainfall ended in most parts of Niger shortly after September 15 (World Climate Programme, WCP, 1992). At the end of the rainy season, a maximum in accumulation over the study-area of 792 mm occurred over the Southern Supersite (Prince et al., 1995). Over the Central-East Supersite, the total accumulation was 410 mm, suggesting a rainfall gradient had been established across the study-area. This is substantiated by the analysis of the EPSAT-Niger data reported in the article of Goutorbe et al. (1994). Their Fig. 4 indicates a wet season variability pattern highlighted by a higher rainfall band in the southern portion of the study-area with rainfall totals exceeding 700 mm and a lower band in the north with totals below 500 mm.

4. Boundary layer structure during synoptic transitions

In this section, we describe how the boundary layer changes in response to the retreat of the ITCZ. Although the launch times at the three active radiosonde sites were not always synchronized, the three sets of soundings basically showed similar boundary layer structures during the IOP. To facilitate this discussion, we first summarize the synoptic
conditions in each of the three sub-intervals of the IOP period and the general features of boundary layer structure. We then discuss the response of the boundary layer to synoptic transitions based on the areal means of morning (10.30 GMT) and mid-day (15.00 GMT) soundings. By determining how the boundary layer changed, we are able to estimate how fast the boundary layer takes to recover after a rainy event. Similar to the study of the FIFE-89 boundary layer, we define recovery as the time required for the post-rainy boundary layer to adjust to its pre-rain boundary layer temperature and mixing ratio conditions. Furthermore, we examine the spatial variations in boundary layer temperature and mixing ratio during the wet and dry periods, linking the temperature and mixing ratio gradients to the rainfall pattern.

(A) Boundary Layer Structure (IH Sites)

The four panels in each part of Fig. 4 show the time series of $V_n, V_n, q_m, q_m, \Theta_n, \Theta_m,$ and $Z_i$ at the CNRM site. Times and dates are shown below the bottom panel. Fig. 5 shows similar information from the mobile system at the four IH sites, whereas Fig. 6 shows the information for the permanent system at the Niamey Airport. Note that the IH mobile site location on any given day is shown below the date in Fig. 5. These latter two sets of figures do not include wind measurements because of reasons discussed in Section 2.

4.1. Wet period (August 21–30)

From August 21 to 30, 1992 an anticyclonic vortex broke off from the 200-mb
anticyclone over West Africa and drifted westward. In conjunction with this vortex, a train of shortwave disturbances propagated along 10°N towards the Atlantic ocean. The mid-latitude westerlies were confined to the north side of 20°N. At the surface, a cyclonic vortex formed near the Greenwich meridian on August 22 and moved westward. The surface convergence line (ITCZ) remained in the vicinity of 24°N during the wet period.

Over the HS study-area, southwesterly winds prevailed in the boundary layer. There were two tropospheric easterly jets. The core of the AEJ was found between the 700- and 500-mb levels with speeds of 10–20 m s⁻¹. This jet was sustained by mid-level warmer tropospheric air to the north and cooler tropospheric air to the south associated with the surface temperature gradient between the Gulf of Guinea and the Sahara Desert. Burpee (1972) has examined the physical aspects of the AEJ. The TEJ was located near 150 mb, and represents an extension of the upper tropospheric easterly jet over south India. Similar to the AEJ, the speed of the TEJ ranged from 10 to 20 m s⁻¹. The mean value of ΔU was 23 m s⁻¹ during the wet period. With the passage of an upper level disturbance, the upper level winds fluctuated between northeasterly (a ridge situation) and southeasterly (a trough situation).

The disturbance brought extensive rainfall to the HS experimental domain. According to Goutorbe et al. (1994), soil moisture measurements at the Southern Supersite indicated that the event replenished the soil moisture content from 106 mm to 120 mm in the upper 1.5-m layer. Soil moisture contents then decreased to 116 mm by August 26. Steady and
extensive rainfall returned on August 27, and the soil moisture content reached its maximum value of 164 mm towards the end of the wet period. The wet surface kept the boundary layer moist and relatively cool. The boundary layer tended to remain stable except in the late afternoon, when it transformed to weakly unstable. Differences in the morning and mid-day boundary layer temperatures and mixing ratios were small. The daytime boundary layer seldom penetrated into the layer of strong vertical directional wind shear.

At 10:30 GMT, the mean surface layer and mixed layer temperatures and mixing ratios (based on an average from the three soundings over the wet period) were: $\theta = 300.9 \, K$, $\theta_m = 300.4 \, K$, $q_a = 16.3 \, g \, kg^{-1}$, and $q_v = 14.7 \, g \, kg^{-1}$. Five hours later, the boundary layer warmed and dried slightly: $\theta = 301 \, K$, $\theta_m = 301 \, K$, $q_a = 15.8 \, g \, kg^{-1}$, and $q_v = 14.6 \, g \, kg^{-1}$. The boundary layer height was then about 800 m. During the wet period, the morning boundary layer over the IHSS site was 1-2°C cooler and 2-3 g kg$^{-1}$ wetter than the other two sites, while the morning boundary layer at Niamey was warmer than the other two sites. Mid-day boundary layer temperatures differed little at the three sites.

Similar to what has been found in the First ISLSCP Field Experiment (FIFE), the speed of recovery of the boundary layer partially depends on soil moisture content (Wai and Smith, 1996). Immediately after a rainfall event, the recovery process of the boundary layer begins with vigorous evapotranspiration, which removes excess water from the upper soil layers over a period of 1-2 days. During this period, the Bowen ratio does not exceed 0.5, denoting that latent heat consumption most of the surface net radiation, with little energy available for direct boundary layer heating. As upper level soil moisture decreases, Bowen ratios increase and heating of the boundary layer becomes more efficient. For example, on August 24, the PBL temperature at Niamey increased 4-5°C, the soil moisture content in the top 1.5 m soil layer at the Southern site decreased to 116 mm, and the Bowen ratio at the West Central site increased to 2.0. All factors are indicative of boundary layer recovery. An additional notable boundary layer warming event occurred on August 29, immediately prior to the major rain event of August 30.

Notably, the horizontal temperature and mixing ratio gradients correlate positively to the rainfall pattern, confirming that soil moisture content helps maintain the southwest-northeast gradient in boundary layer temperature and moisture. Effectively, the larger soil moisture contents in the southwest keep the boundary layer relatively cool and moist with evapotranspiration consuming most of the solar heating, while larger sensible heat fluxes and smaller moisture fluxes in the northeast arising from drier soil lead to elevated boundary layer temperatures and reduced mixing ratios. In essence, soil moisture content is considered a critical parameter in the surface energy balance because it can induce such gradients.

4.2. Transitional period (August 30–September 15)

The 200-mb disturbance gradually dissipated on August 30 after producing major rains over most of the study-area. An anticyclone re-formed over the western portion of North Africa, but never re-established itself and underwent several changes in its circulation pattern. When the anticyclone was strong, the upper level winds were easterly or south-easterly. When it was weak, it disintegrated into a vortex and moved off the West African coast. The mid-latitude westerlies remained north of 20°N. The locations of the AEJ and TAEJ changed very little as compared with events during the wet period. At the surface, the boundary layer winds were southweste when the anticyclone was strong, changing to northeasterly or northwesterly when the upper-level anticyclone was weak. The surface streamlines during this period indicated a gradual southward migration of the ITCZ. By September 15, the ITCZ was located near 16°N.

As the shortwave disturbances became less active, the steady rains changed to showers beginning on August 31. Showery weather lasted through September 15. In between, the rain events were followed by approximately 2-day dry periods (August 31–September 1, September 4–5, and September 8–9). The layer of vertical wind shear frequently descended into the boundary layer. During the daytime period, the boundary layer began to recover from the large surface evapotranspiration rates. The 1.5-m soil moisture contents at the Southern Supersite fell continuously to 100 mm by September 9. The last major rainfall on September 15 replenished the upper level soil moisture content to 130 mm. As rainfall became less frequent afterwards, the boundary layer underwent a gradual warming. As reported by Goutorbe et al. (1994) at the Central-West site, the Bowen ratio was about 0.6 on September 6 indicating that sensible heating had become relatively important. During the transition period, boundary layer temperatures and mixing ratios exhibited larger daily and diurnal fluctuations than observed during the wet period.

At 10:30 GMT the mean temperature and mixing ratio conditions for the transition period were: $\theta = 304.3 \, K$, $\theta_m = 303.6 \, K$, $q_a = 15.4 \, g \, kg^{-1}$, and $q_v = 13.9 \, g \, kg^{-1}$. By 15.00 GMT, the boundary layer heated 2-3°C, and dried: $\theta = 306.0 \, K$, $\theta_m = 305.9 \, K$, $q_a = 13.7 \, g \, kg^{-1}$, and $q_v = 12.7 \, g \, kg^{-1}$. During the transitional period, the boundary layer was about 4-5°C warmer and 2 g kg$^{-1}$ drier than in the wet period. In contrast to the wet period, the mid-day surface layer over the IH site was slightly warmer and wetter than other two sites by 0.5°C and 1.5 g kg$^{-1}$. The morning boundary layer temperatures at the IH and Niamey sites were cooler than those at Hamdallaye.

Because of decreased rainfall during the transition period, the boundary layer underwent more efficient heating than during the wet period, tending towards the more classic unstable boundary layer structure. As the soil moisture dropped, the rate of sensible heating rose rapidly. The increase in significant warming of the boundary layer began on September 2 and progressed on throughout the transition period despite continuous showers.

4.3. Dry period (September 16–October 12)

On September 14, the upper level winds at 14 km indicated the intrusion of a mid-latitude westerly trough into the Sahel zone. A layer of westerly wind then appeared between 10–15 km on September 16. Normally, the southward penetration of mid-latitude westerlies coincides with the beginning of the winter season (Adolfahat, 1974). The mid-latitude westerlies pushed the 200-mb anticyclone south to the Guinea coastal area. For several days, successive westerly troughs over the western region of North Africa could be traced back upstream on a planetary scale. The upper tropospheric westerly flow persisted throughout October 7. The downward intrusion of the westerly winds penetrated as low as 7 km. During the period of maximum southward penetration, the speed of the westerly jet
The AEJ was located between 2–4 km. The mean value of $\Delta U$ was 16 m s$^{-1}$, about 7 m s$^{-1}$ smaller than during the wet period. Notably, the mid-September appearance in West Africa of strong upper level westerlies was not an isolated synoptic event. Relative to the climatological mean, there was a corresponding reduction in the maximum winds in the core of the TEJ in September and October of some 5–10 m s$^{-1}$.

As the upper level westerlies began to readjust to the large-scale circulation pattern, the surface wind in the HS study-area frequently underwent shifts from northeasterly to southwesterly and vice versa. During this period, the surface convergence axis oscillated between 17 and 10$^\circ$N, finally positioning itself near 14$^\circ$N on October 12. With the penetration of the westerlies to West Africa, the rainfall abated on September 16 and showery weather gave way to dry weather. At that point, the boundary layer went through a 3-week dry down period. Boundary layer winds became more variable with more instances of easterly flow. On certain days, the entire boundary layer was dominated by northeasterly flow. The boundary layer height often exceeded 2 km, especially at the end of the IOP.

The 1.5 layer soil moisture contents at the Southern Supersite dropped to about 60 mm by the end of the IOP. Accompanying the drop of soil moisture was a gradual increase in Bowen ratio. Daytime Bowen ratios remained around 0.3 at the beginning of the dry period, increased to 0.4 by September 24, and then increased rapidly after October 1. At the end of the IOP, the Bowen ratio was 1.1, signifying over 50% of the surface available heating was transformed directly into boundary layer heating. During the daytime, diurnal fluctuations of boundary layer temperature and mixing ratio were largest over the IOP.

Fig. 7 shows all 19 profiles of $R_b$ from 05.00 GMT on September 25 to 17.00 GMT on September 26. The closed circles represent unstable conditions ($R_b < 0$), the bars represent mechanically driven turbulence ($0 < R_b < 0.25$), and the blank spaces indicate stable conditions ($R_b > 0.25$). A value of 0.25 is chosen for the cutoff in separating turbulent from stable flow (Businger, 1982). This diagram shows the diurnal cycle of the nocturnal stable and convective daytime boundary layer. During the unstable period, mechanically driven turbulence by vertical wind shear is an important production mechanism for turbulent kinetic energy.

The regions of strong stability inside the core of the AEJ (near 5 km) and the upper level jet (15 km) are indefinite stable as $\Delta \mu / \Delta z$ approaches zero and the temperature lapse rate remains positive. These regions act as a rigid lid preventing turbulent transport of heat and moisture from below, even though the regions above and below the core of the jet are themselves turbulent. Therefore, when the meridional temperature gradient is large at the lower levels, the core of the AEJ can reach to a higher level aloft by the thermal wind relationship. On the other hand, as the meridional temperature gradient weakens, the core of the jet intrudes to a lower level. A rigid lid will influence the depth of the boundary layer and exchange of heat and moisture between the surface and free atmosphere. The vertical shear layer also determines the intensity and depth of convective clouds. If the updrafts are weak and the vertical shear is large, the upper regions of the clouds are sheared off and do not survive for very long. Therefore, the growth of large convective clouds requires the intensity of surface heating to match the intensity of the vertical shear of the wind (Asnani, 1993).
At 10.30 GMT the mean temperature and mixing ratio conditions for the dry period were: $\Theta_1 = 306.3 \, K$, $\Theta_n = 305.9 \, K$, $q_s = 14.4 \, g \, kg^{-1}$, and $q_n = 12.8 \, g \, kg^{-1}$. In the mid-afternoon (15.00 GMT), the boundary layer warmed up considerably: $\Theta_1 = 309.2 \, K$, $\Theta_n = 309.5 \, K$, $q_s = 11.0 \, g \, kg^{-1}$, and $q_n = 9.8 \, g \, kg^{-1}$, about 9°C warmer and 4–5 g kg$^{-1}$ drier than the wet period. The depth of the boundary layer extended 3–4 km during mid-day, meaning that the layer of strong directional vertical shear becomes part of the boundary layer structure.

The vertically averaged boundary layer potential temperature and mixing ratio ($\Theta_s$ and $q_s$) over the dry period at the IH SSF site remained cooler and wetter than the other two sites by 1–2°C and 2–3 g kg$^{-1}$. Here the vertical average is taken between the surface (s) and the boundary layer top ($Z_t$). Further analysis shows that the surface temperature and mixing ratio at the IH SSF site indicated a similar pattern. We have concluded that the cooler and wetter boundary layer at the IH SSF site is independent of vertical averaging of the sounding profiles. This can be attributed to its location within the area of minimum rainfall during the IOP. From the wet to dry periods, the boundary layer changed from a relatively stable conditions to convectively unstable. The determining factor was the evapotranspiration which regulated the soil moisture content and the amount of surface sensible heat flux available for boundary layer heating and energetics.

5. Potential temperature and mixing ratio budgets

Further insight into the physical processes at work in the boundary layer can be gained by examining the terms in the temperature and mixing ratio budget equations. We have calculated the terms in the budget with a five-point grid centered over Hamadallaye. The grid distance is 66 km. In the budget calculations, only $u$ and $v$ measurements from the Hamadallaye soundings were used to ensure the greatest possible accuracy, since the quality of the CNRM winds was significantly better than that of either the operational or IH winds, as well as the European Center for Medium Range Weather Forecasts (ECMWF) global analyses. To preserve consistency and continuity of atmospheric features, we require that the soundings were launched at approximately the same time. There were ten sets of soundings satisfying this condition: August 21 and 29, September 9, 10, 12, and 25, and October 3, 7, 8, and 9. Nine launches were made 10.30 GMT while one was at 22.20 GMT on September 25. To enrich the data coverage, we have calculated the temperature and mixing ratio at five additional locations surrounding the sounding stations by the following extrapolation procedure:

$$\Theta_2 = \Theta_1 + (\Delta \Theta / \Delta x) x_1 + (\Delta \Theta / \Delta y) y_1$$

$$q_2 = q_1 + (\Delta q / \Delta x) x_1 + (\Delta q / \Delta y) y_1$$

where subscripts 1 and 2 indicate the location index, $\Theta_1 / \Theta_2$, $\Delta \Theta / \Delta x$, $\Delta \Theta / \Delta y$, $\Delta q / \Delta x$, $\Delta q / \Delta y$ are the horizontal gradients of potential temperature and mixing ratio, and $(x_1, y_1)$ are horizontal distances. The profiles of the horizontal gradients were first calculated from the ECMWF optimal analysis products at 14 levels over a domain of 7.5 $\times$ 5°. Complete profiles were then obtained by linear interpolation. Profiles of vertical velocity ($w$) were also determined from the ECMWF data.

When the profiles of $\Theta$ and $q$ are known, the associated profiles at the four grid points are calculated by the following polynomials:

$$\Theta(x, y, p) = a(p)x + b(p)y + c(p)$$

$$q(x, y, p) = d(p)x + e(p)y + f(p)$$

where $x$ and $y$ indicate the location of a grid. Coefficients $a, b, c, d, e, f$ at each pressure level are determined by a least squares method. This approach has been described by Yanai et al. (1973) in detail.

Once the profiles of $\Theta$ and $q$ are known at the grid points, the profiles of budget terms were calculated from the following finite difference equations:

$$\frac{\partial \Theta}{\partial t} = - \nabla \cdot \mathbf{q} - \rho_0 \frac{w}{\gamma}$$

$$\frac{\partial q}{\partial t} = - \nabla (\nabla \cdot q) - \rho_0 \frac{w}{\gamma}$$

where $\partial \Theta / \partial t$ and $\partial q / \partial t$ are the local rate of changes, $\nabla \cdot \mathbf{q}$ and $\nabla q$ the horizontal advection terms, $\rho_0 \frac{w}{\gamma}$ the vertical advection terms, $F_u$ the flux divergence term of potential temperature including radiation and turbulence, and $F_q$ the flux divergence term of moisture including turbulence and cloud microphysical processes. Here the flux divergence terms are taken as the residuals from the rest of the budget terms.

For consistency checking, the sign of horizontal advection is checked against the thermal wind relationship, while vertical motion is checked by comparing the sign of the divergence ($\nabla \cdot \mathbf{q} = - \rho_0 \frac{w}{\gamma}$) with the sign of vertical motion produced by the ECMWF model. In the divergence relationship, $\rho$ is the meridional velocity, $\phi$ the latitude, and $\gamma$ the mean earth radius ($6.37 \times 10^6$ m). Divergence is obtained by simplifying the potential vorticity equation:

$$\frac{d(\zeta + f)}{dt} = - (\zeta + f) \nabla \cdot \mathbf{q} + (\zeta + f) \nabla \cdot \mathbf{q} - \phi_0 \frac{w}{\gamma}$$

by assuming that the tilting term and relative vorticity term ($\zeta$) are zero. In the vorticity equation, the first term on the left hand side is the total time rate of change of potential vorticity, while the right hand side contains the divergence and lifting terms. Since divergence is proportional to the speed of the northerly wind and to decrease in latitude, convergence is proportional to the speed of the southerly wind and increase in latitude. Therefore, subsidence increases as air moves towards the equator while ascending motion increases as air moves away from the equator.

The budget profiles from Eq. (5) and Eq. (6) are embedded with various features of differing length scales which makes the interpretation somewhat difficult. Therefore, we have stratified the budget profiles into three layers: (1) the entire boundary layer, (2) from boundary layer top to 550 mb, and (3) from 550 mb to 300 mb. We then vertically average the budget terms within the three layers. Summaries of these results are presented in Table 1 (Table 2). The following presents individual discussions for budgets of 5 days best highlighting the changes.
Temperature advection in the middle layer, cold temperature advection compensated with temperature, mixing ratio, and wind dropping for a local maximum of inversion.

5.1. August 21 case

The morning boundary layer (07:00 GMT) was capped by an inversion layer between 650 m and 1600 m with a layer-mean temperature ($\Theta_a$) of 301.7 K. The boundary layer wind was southwesterly and veered through the inversion layer to northeasterly above the inversion top. In the free atmosphere, the wind had a strong easterly component. The AEJ was positioned near 4 km with a speed of 13 m s$^{-1}$. The TEJ was found near 15 km with a speed of 14 m s$^{-1}$. The boundary layer-mean mixing ratio ($q_b$) was about 15.9 g kg$^{-1}$, dropping to 8 g kg$^{-1}$ at the inversion layer. It then decreased gradually with height except for a local maximum of 3 g kg$^{-1}$ near 6 km. By 13:00 GMT, the boundary top had grown to 1200 m. The $\Theta_a$ and $q_b$ quantities were 306 K and 14.6 g kg$^{-1}$ at this time. A thin cloud layer formed between 5500–5600 m. Above the boundary layer, the basic profiles of temperature, mixing ratio, and wind speed underwent little change.

In terms of boundary layer heating, flux divergence ($0.71 \times 10^{-4}$ K s$^{-1}$) and warm temperature advection ($0.55 \times 10^{-4}$ K s$^{-1}$) were the major source terms. The advection of drier air and flux divergence also produced net drying of the boundary layer. In the middle layer, cold temperature advection compensated with flux divergence, producing small cooling. Vertical temperature advection was also small. The downward transport of moisture from the local maximum near 6 km contributed to moistening of the middle layer. In the upper layer, the effects of subsidence and flux divergence overcame cold temperature advection, resulting in a net warming. The horizontal advection of moisture balanced vertical advection.

5.2. August 29 case

On August 29, the temperature profile indicated that the entire atmosphere was stable. However, the dew point depression indicated that two cloud layers were between 3400–8966 m and 4180–4260 m. The surface temperature was 296 K while the mixing ratio was about 14.5 g kg$^{-1}$. The boundary layer wind was south-southwesterly, backing to easterly at 2 km. The AEJ was located at 4 km with a speed of 18 m s$^{-1}$. During the next 3 h, both cloud layers were lifted upward as the boundary layer grew. By 13:00 GMT, the low-level cloud deck dissipated while the upper level cloud was positioned in a layer between 5470–5550 m. The boundary layer depth was 1500 m with $\Theta_b = 303.2$ K and $q_b = 15$ g kg$^{-1}$. The
upper level temperatures and mixing ratios did not vary significantly. The speed of the TEJ positioned at 16 km was 20 m s\(^{-1}\).

In the boundary layer, the flux divergence terms for potential temperature (1.69 \(\times\) 10\(^{-3}\) K s\(^{-1}\)) and mixing ratio (0.35 \(\times\) 10\(^{-4}\) s\(^{-1}\)) produced most of the increase in heating and moistening. In the middle layer, warm temperature advection of dry air was compensated by flux divergence (evaporative cooling of the cloud layer). In the upper layer, warm temperature advection was also balanced by flux divergence. On the wet days, the warming effects of temperature flux divergence and warm temperature advection overcame the cooling effects of vertical temperature advection and therefore led to a net boundary layer warming. Similarly, both flux divergence and vertical advection of moisture were important for boundary layer moistening. Above the boundary layer, flux divergence was almost balanced by horizontal advection. Therefore, transports of heat and moisture by boundary layer circulations and turbulence should be considered important forcing mechanisms.

5.3. September 10 case

The morning boundary layer (07.00 GMT) on September 10 was less than 600 m deep. Near 5 km, a boundary layer top from the previous day persisted. The \(\Theta_h\) and \(q_v\) terms were 300.0 K and 15.2 g kg\(^{-1}\), respectively. The surface southwesters winds backed to northeasters near 2 km. The core of the AEJ was rather broad, centered between 2–3 km. By 13.00 GMT, a deep cloud layer formed between 2800–5200 m. The sub-cloud layer temperature and mixing ratio were 307 K and 14.0 g kg\(^{-1}\).

Flux divergence (2.72 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) accounted for most of the boundary layer warming. In the middle layer, the cooling was also produced by flux divergence (1.10 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) as the cloud top underwent strong radiative cooling. The increase in mixing ratio stemmed from turbulent moistening. Because the upper cloud deck was included in the top layer, subsidence (0.72 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) could not balance flux divergence (1.13 \(\times\) 10\(^{-4}\) K s\(^{-1}\)), and thus a net cooling resulted. Turbulent moistening (0.74 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) was also dominant.

5.4. September 25 case

A stable layer was found at 400 m with \(\Theta_h = 301.5\) K and \(q_v = 15.8\) g kg\(^{-1}\). An upper level inversion was located at 4 km from the previous day. The wind profiles during the day were complicated. The surface southwesters winds shifted to northeasters near the top of the stable layer. In the free atmosphere, northeasters were found between 1500 and 8000 m associated with a broad core of the AEJ (9 m s\(^{-1}\)), while a region of southwesters between 8–16 km peaked at 13 m s\(^{-1}\) near 14 km. By 13.00 GMT, the boundary layer had grown to 2100 m with \(\Theta_h = 309.5\) K and \(q_v = 12.4\) g kg\(^{-1}\). The winds were mainly northeasters, shifting to more easterly above the boundary layer top. The core of the AEJ was lifted to 5 km with a speed of approximately 10 m s\(^{-1}\). The axis of the upper level westlers shifted to 15 km. In the boundary layer, flux divergence largely accounted for heating (1.31 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) and moistening (0.88 \(\times\) 10\(^{-4}\) s\(^{-1}\)). Both horizontal and vertical advection terms were relatively small and tended to balance each other. In the upper layer, changes in potential temperature and mixing ratio were small.

Towards the evening (17.00 GMT) of September 25, the boundary layer merged with the mid-level stable layer near 5 km with \(\Theta_h = 311.5\) K and \(q_v = 9.3\) g kg\(^{-1}\). The vertical positions of the core of the AEJ and the axis of the westerslies differed very little. By 19.00 GMT, a stable surface layer appeared within surface southerlies returning with a speed of 2 m s\(^{-1}\). The AEJ was now more northeasters. The basic features of the westerslies changed very little. By 01.00 GMT on September 26, the stable layer extended to about 1 km and the boundary layer winds were dominated by southwesterlies. Above the stable layer, the temperature and mixing ratio were quite uniform. The basic profiles of wind velocity, temperature, and mixing ratio were quite similar to those found at 07.00 GMT on September 25.

The strong boundary layer and middle layer cooling rates mostly resulted from flux divergence (0.4 \(\times\) 10\(^{-4}\) K s\(^{-1}\)). The advection of cool air contributed to about 20% of the total temperature change. Drying resulted from subsidence (1.3 \(\times\) 10\(^{-4}\) s\(^{-1}\)) and flux divergence due to stratification (0.68 \(\times\) 10\(^{-4}\) s\(^{-1}\)). In the upper layer, flux divergence was also dominant (1.8 \(\times\) 10\(^{-4}\) K s\(^{-1}\)).

5.5. October 8 case

On October 8, a strong stable boundary layer was confined to below 700 m with \(\Theta_h = 299\) K and \(q_v = 11\) g kg\(^{-1}\). An upper level inversion was noted near 5 km, left over from the previous day. The surface winds were quite variable. Above the stable surface layer, northeasters winds prevailed. The influence of westerly winds diminished with their axis confined between 13–14 km. The strengths of the AEJ and westerlies were also reduced to 6–8 m s\(^{-1}\). By 13.00 GMT, the boundary layer height was 1700 m while the upper level inversion base did not change; at the inversion \(\Theta_h = 300.9\) K and \(q_v = 9.8\) g kg\(^{-1}\). Boundary layer winds were southwesterly. The wind systems above the boundary layer basically remained unchanged.

On this day, boundary layer advection was small, and flux divergence due to surface sensible heating (2.02 \(\times\) 10\(^{-4}\) K s\(^{-1}\)) produced most of the diabatic heating. (The observational study of Dolman et al., 1997, using a mixed layer model and analysis of the boundary layer winds, suggests the horizontal advection term on this day was more significant. However, the residual term of their study lumped horizontal advection with both the subsidence and net radiation terms, thus resolving this difference concerning the role of advection cannot be ascertained directly from the two sets of budget analyses.) As a check on the flux divergence term for heat on this day, we calculated its vertical integral minus a term approximating the radiative flux divergence, thus obtaining the surface sensible heat flux (i.e. the \(\partial\)-derivative of the vertical divergence of vertical eddy heat transport is surface sensible heat flux, ignoring the horizontal eddy flux transports are negligible). This gave a surface heat flux of 215 W m\(^{-2}\), which was well within the range of the 205–240 W m\(^{-2}\) observations reported on October 8–9 at mid-morning by Kebat and Goutorbe (1995) at five different monitoring sites throughout the study area.

The net drying that took place resulted from mixing between the stable layer and moist air aloft. In the middle layer, there was net cooling as a result of the mixing of cooler surface air and warmer air aloft. However, there was a net moisture gain as a result of the mixing of moist air aloft and dry air below. In the upper layer, the changes of temperature
and mixing ratio were small. During the dry period, the flux divergence term acts as either a source or sink in the boundary layer temperature and moisture budgets. Furthermore, when horizontal advection produces drying in the boundary layer or in the free atmosphere, turbulence moistening from the surface and advective transport of moisture between vertical layers becomes the only moisture available for cloud formation or rain processes.

6. Comparison between HAPEX-Sahel and FIFE boundary layers

The boundary layer structure and large-scale atmospheric circulations in HAPEX-Sahel and FIFE have both similarities and differences as indicated in Table 3. During FIFE-89, the low level wind system (below 4 km) was a deep, warm, and humid southerly or southwesterly flow originating from the Gulf of Mexico. Under undisturbed conditions, there was little vertical wind shear in the boundary layer except for a short period during an isolated frontal passage. The upper level flow was mostly westerly. As with HAPEX-Sahel, the FIFE boundary layer was also subjected to showers, convective rainfall, and frontal rainfall.

The rainfall in both field experiments established well-defined soil moisture gradients in the experimental domains. During FIFE-89, the rainfall established a northwest-southeast soil moisture gradient with coincident cooler surface and air temperatures, smaller Bowen ratios, and lower surface wind speeds in the southeastern moist sector (Smith et al., 1994). The result of the soil moisture gradient and an accompanying vegetation cover gradient, was the maintenance of a secondary circulation within the boundary layer with daytime convergence and rising motion over the warm sector. During HAPEX-Sahel, the maximum rainfall was found in the west-southwestern sector where the surface layer at the IH SSF site was cooler and wetter than the other two radiosonde sites by 1–2°C and by 2–3 g kg⁻¹ during the IOP.

The HAPEX-Sahel boundary layer showed a similar response to rain events as found in

<table>
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<tr>
<td><strong>Comparison between FIFE and HAPEX-Sahel boundary layers</strong></td>
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<tr>
<td>FIFE</td>
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<td><strong>Date of IOP</strong></td>
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<td><strong>Site location</strong></td>
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<td><strong>Prevailing boundary layer wind</strong></td>
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FIFE; see Wai and Smith (1996) for the FIFE analysis. After a rain event, evapotranspiration accounted for 55–65% of the surface net radiation on a time scale of 1–2 days with little sensible heating available for boundary layer heating and energetics. The speed of the boundary layer recovery primarily depended on antecedent soil moisture conditions, surface wind speed, and surface wind direction (along with the associated temperature of the advected air mass).

Of significance is the possibility of sustained secondary circulations within the Sahelian boundary layer analogous to those in FIFE. Although these would be difficult to detect and characterize with the limited HAPEX-Sahel measurement network, which was not designed for such investigations, we can invoke the same arguments used in our FIFE study (Smith et al., 1994). That study established the presence of such circulations in response to well organized surface gradients of moisture and sensible heating, using observational analysis.

The more riveting question concerns the significance of such circulations, particularly if they can be maintained over extended periods as we have seen in FIFE; see Wai and Smith (1996). In Fig. 8, we provide with a schematic diagram, a conceptual model of a plausible secondary circulation process within the Sahel, that can help explain why its interannual rainfall variability is so large. The diagram is a modification of an original diagram from Haywood and Oguntoyinbo (1987) which illustrates, in a climatological sense, the annual progression and retreat of the southwesterly monsoon into North Africa. The various panels from February to December indicate the northward penetration and changing depth of the surface southwesterlies with a thick solid line along a cross-section from the Gulf of Guinea in the south (−1°N) to the Algerian region in the north (−25°N). Upon the original diagram we have superimposed schematics of the characteristic convective systems that accompany the migration of the southwesterlies into and across the Sahel, systems that are consistent with the weather types described in the figure caption and discussed in the Hayward and Oguntoyinbo atlas.

If the progression were smooth, such as in a climatological process, then the precipitating storms would leave behind a negative south to north soil moisture gradient and the conditions for a sustained south to north aligned secondary circulation would exist. This would result with the southern zone experiencing the most exposure to storms maintained by monsoon southwesterlies; the anticipated soil moisture gradient from this process is indicated at the bottom of Fig. 8. Such circulations are schematically illustrated in the August and October panels (the vertical scale of these circulations are exaggerated for clarity). Through Coriolis effects these circulations would turn to the right leading to a northeast-southwesterly alignment with lower level southwesterlies and upper level north东北lies. Such circulations would reinforce the large-scale monsoon circulation, and in essence help maintain the monsoon against the intrusion of the northwesterlies which leads to the eventual retreat of the monsoon (see Newell et al., 1972). Given this situation, the wet season would be extended and additional rainfall would fall. On the other hand, if the locations of the storms during the progression of the monsoon were random, and an organized south-north soil moisture gradient were not established, the large-scale monsoon circulation would not be reinforced by a secondary circulation and the monsoon would be expected to retreat sooner. Such an explanation is only hypothetical at this point and would require a carefully designed modeling experiment to substantiate, but it is a
physically consistent and fully plausible explanation of why the duration of the wet season and the accumulated rainfall can undergo year-to-year variability. Clearly, we do not presume that this is the only process leading to interannual rainfall variability over the Sahel. However, it is an intriguing mechanism and provides another dimension to the debate over the cause of flood and drought cycles that frequent the Sahel.

7. Conclusions

The variability of the Sahelian boundary layer within the HAPEx-Sahel study-area has been examined during a premature withdrawal of the inter-tropical convergence zone. The 1992 rainy season ended prematurely because the upper level atmospheric circulations changed from easterly to westerly. Normally, the arrival of the westerlies coincides with the beginning of the winter season. The circulation brought the surface northeasterly flow southward and led to the retreat of the Southwest Monsoon. The rainy season ended in most parts of Niger shortly after the arrival of westerlies. From the wet to dry period, the ITCZ shifted from 21°N to about 14°N. At this time, we do not have a complete explanation for the early arrival of the westerlies. However, the change in the atmospheric circulation over the western region of North Africa is clearly related to global-scale circulations because successive westerly troughs over West Africa can be traced back upstream on a planetary scale.

The boundary layer responded quickly to the transition from wet to dry seasons. During the wet period, evapotranspiration was the dominant term in the surface energy balance, continually removing excessive moisture from the upper soil layers. Little surface energy was available for sensible heating of the boundary layer during that time. Therefore, the boundary layer generally remained cool (300.5 K) and moist (15.5 g kg\(^{-1}\)). Morning and mid-day differences in potential temperature and mixing ratio were small.

During the transitional period, less extensive rainfall provided more time for the boundary layer to recover from surface wetting. The Bowen ratio ranged from 0.2–0.3 on wet days to 0.6 on the dry days. More surface sensible heat flux was available for boundary layer heating which led to its warming and drying (305 K and 14 g kg\(^{-1}\)). Differences in the morning, and mid-day temperatures and mixing ratios were larger than those in the wet period.

During the dry period, the role evapotranspiration decreased after the last rain event on
September 15 as the Bowen ratio increased from 0.3 to 1.1 by the end of the IOP. As soil moisture decreased, even more sensible heat flux was available for boundary layer heating and energetics than in the prior two periods. During this portion of the IOP, the boundary layer reached its warmest and driest conditions (308 K and 12 g kg\(^{-1}\)). Moreover, the mean depth of the boundary layer at 15.00 GMT was at its maximum, exceeding 3 km, as compared with 900 m for the wet period and 1 km for the transition period.

The budget calculations indicate that horizontal advection and vertical flux divergence were the most important terms during the wet period in maintaining energy balance. During the dry period, subsidence and the vertical flux divergence terms were the most important. From the wet to dry periods, the vertical shear difference (\(\Delta U\)) of the zonal wind was reduced from 23 m s\(^{-1}\) to 16 m s\(^{-1}\). The reduction of vertical zonal wind shear across these two periods is consistent with the characteristics of wet and drought years as described by Kanamitsu and Krishnamurti (1978) and Newell and Kidson (1984).

A key issue is the significance of what we have found involving the relationship between Sahelian rainfall and boundary layer dynamics as controlled by land surface processes. Here we can only speculate, as we do not yet have sufficient data to test our hypothesis that has emerged from this investigation. As shown in the Smith et al. (1994) observational FIFE study and a number of mesoscale modeling studies, well established and persistent land surface gradients of soil moisture and vegetation cover can give rise to and maintain, through differential sensible heating of the boundary layer, persistent secondary circulations embedded as a perturbation in the large-scale flow. Wai and Smith (1996) have proceeded to show with FIFE analysis how these perturbation circulations can be maintained, even in the presence of intermittent synoptic events which disrupt the background flow in the boundary layer. In the case of the Sahel region, during the course of the IOP, we have found evidence in a south–north gradient in soil moisture possibly correlated to an associated gradient in vegetation cover. The data indicate the northern portion of the HS study-area begins drying before the southern portion. According to the precipitation measurements from the EPSAT-Niger network, this gradient is preserved over a relatively lengthy time scale, and since the resistance to a perturbation circulation by the balanced flow is relatively small at Sahelian latitudes, the conditions are ideal for an ongoing secondary circulation. Given a south to north soil moisture gradient, a north–south aligned secondary circulation would persist with the upward branch in the northern Sahel (in the zone of relatively low soil moisture and higher sensible heat transfer to the boundary layer), and the downward branch somewhere over the southern Sahel–rain forest zone (where surface conditions are moister, cooler, and less capable of directly heating the boundary layer). In a simple zonally symmetric situation, as shown schematically in Fig. 8, the perturbation circulation would both support the maintenance of the low level monsoon southwesterlies across the Sahel due to Coriolis turning of the near-surface flow, and help maintain mid-level easterlies over the Sahel associated with the African Easterly Jet.

Embedded in the large-scale circulation is the south to north positive surface temperature gradient from the relatively cool Gulf of Guinea and African tropical rain forest, to the hot and dry Sahara desert. This surface gradient, partially imposed onto the atmosphere above, is what establishes and maintains the mid-tropospheric AEJ, through thermal wind dynamics. A secondary circulation superimposed on this basic state, reinforcing the low level southwesterlies and mid to upper level northeasterlies, would be expected to resist the southern retreat of the monsoon as the dry season wears on and the central Sahara begins to cool. This leads to the conditions for a surface driven process which could modulate the retreat of the monsoon on an interannual basis, dependent upon the intensity of the soil moisture gradient across the Sahel. In other words, a wet season rainfall pattern that built up a strong cross-Saharan soil moisture gradient would have the effect of preserving the strength of the monsoon at its preferred late summer position, which in turn would resist the southerly advance of the of the deep-layer westerlies and re-establishment of dryer northerly flow over the Sahel. In effect, a pre-existing strong soil moisture gradient at the time of the onset of the Sahel dry season would hold up the retreat of the easterlies and the advance of the westerlies, which climatologically speaking favors more humid conditions. Such a hypothesis may be testable with a long year-to-year time series of precipitation data and mid-troposphere winds in conjunction with a limited area model. At this juncture, we can only speculate that a persistent perturbation boundary layer circulation, forced by differential soil moisture conditions over the Sahel, has the possibility of altering rainfall by extending the rainy season, with the additional possibility that this is an interannually modulated phenomenon dependent upon the magnitude and intrinsic organization of the soil moisture gradient pattern.

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