The Seasonal Evolution of the Atmospheric Circulation over West Africa and Equatorial Africa

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ABSTRACT

This paper examines the mean annual cycle of rainfall and general circulation features over West Africa and central Africa for 1958–97. Rainfall is examined using a 1400-station archive compiled by the first author. Other circulation features are examined using the NCEP–NCAR reanalysis dataset. Important features of the reanalysis zonal wind field are shown to compare well with the seasonal evolution described by the radiosonde observations. In addition to the well-known African easterly jet (AEJ) of the Northern Hemisphere, the seasonal evolution of its Southern Hemisphere counterpart is also described. Thermal wind calculations show that although the southern jet is weaker, its existence is also due to a local reversal of the surface temperature gradient. In the upper troposphere, a strong semiannual cycle is shown in the 200-mb easterlies and a feature like the tropical easterly jet (TEJ) is evident south of the equator in January and February. The paper describes the movement of the rainbelt between central and West Africa. An asymmetry in the northward and southward migration of the rainbelt is evident.

The paper discusses the influence that the jets may have on rainfall and possible feedback effects of rainfall on the jets. Evidence suggests that the midtropospheric jets influence the development of the rainy season, but also that the rainfall affects the surface temperature gradient and in turn the jets. In the Northern Hemisphere, east of 20°E, the axis of the TEJ is located so that it may promote convection by increasing upper-level divergence. However, west of 10°E and in the Southern Hemisphere, the location of the TEJ is consistent with the suggestion that it is the equatorward outflow of convection that produces the TEJ.

1. Introduction

Sahelian West Africa is one area of the world that has undergone a remarkable change in climate. Thirty-year rainfall means have declined by 30%–40% between 1931–60 and 1968–97. The protracted dry spell since the late 1960s has prompted study of the region’s climate, as has the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE), which took place in 1974. Since that time, numerous papers have increased our understanding of rainfall variability in the Sahel. These have shown, for example, that the ENSO signal and tropical Atlantic sea surface temperatures (SSTs) have a strong enough correlation with Sahelian rainfall to be useful predictors on seasonal timescales (Semazzi et al. 1988; Ward 1994; Moron and Ward 1998; Lamb 1978; Rowell et al. 1995). In addition, systematic differences in the basic state between wet and dry years have been noted with some consistency. In wet (dry) years the midtropospheric African easterly jet (AEJ) tends to be weaker (stronger) and more poleward (equatorward), while the 200-mb tropical easterly jet (TEJ) and the monsoonal flow tend to be stronger (weaker) (Newell and Kidson 1984; Fontaine et al. 1995; Kanimitsu and Krishnamurti 1978; Grist and Nicholson 2001). There is less agreement as to the role that these features play in rainfall variability. Similarly, differences in the intertropical convergence zone (ITCZ) between wet and dry years have been noted, but there is some disagreement as to whether variations in its location or intensity are more important in determining rainfall variability (Kraus 1977a,b; Ilesanmi 1971; Miles and Folland 1974; Tanaka et al. 1975; Nicholson 1980; Citeau et al. 1989; Ba et al. 1995). Thus, the fundamental reasons for the region’s unique rainfall variability have not been conclusively determined.

In contrast to the wealth of studies over West Africa, there have been relatively few studies of equatorial Africa (CLIVAR 2000). Our present understanding of meteorological processes in the region is correspondingly weak. Notable studies by Laing and Fritsch (1993) and Mohr and Zipser (1996) have described the spatiotemporal distribution of mesoscale convective systems. Apart from this, little is known about either the synoptic or mesoscale systems that shape the rainy season or about the factors governing interannual and interdecadal variability. Recent Tropical Rainfall Measuring Mission (TRMM) data have shown that the frequency of light-
ning there is by far the greatest in the world. State-of-the-art satellite algorithms overestimate rainfall in this region by a factor of 2 or 3 (McCollum et al. 2000; Nicholson 2000), suggesting a poor understanding of rain processes in the region. In this context it is interesting to note that extensive biomass burning in this region produces high smoke concentrations, a factor shown by Rosenfeld (1999) to suppress convective rainfall. The region clearly merits further study as even basic aspects of the general atmospheric circulation are poorly understood.

Considering the gap in the quantity of research existing between West and central or equatorial Africa, it is ironic that the regions exhibit a number of similar or common circulation features. Thus, much stands to be gained by studying the two regions in parallel. This point has been missed, however, by most previous studies. An example is the midtropospheric easterly jet in the Southern Hemisphere. A number of authors [beginning with Burpee (1972)] have acknowledged the existence of this jet, but neither its annual cycle nor its structure has been described in detail. Likewise, its influence on regional meteorology has not been considered.

In this article we begin to address this deficit in the literature by jointly examining the annual cycles of circulation features over West Africa and central Africa. This allows us to better describe the basic state of the atmosphere in these regions, something that is of fundamental importance to modeling efforts. Our interpretation of the results also offers some clarity to the meteorological processes at work, particularly in central Africa. This can facilitate the interpretation of modeling results and the design of forecast methods.

This article makes several new contributions to the understanding of the meteorology of these regions. Our work provides the first analysis of the annual cycles of the midtropospheric easterly jets and the TEJ over Africa. Since past studies of these features have generally been limited to the local rainy season, usually only the boreal summer (and often only August) has been examined. Here we show, counter to previous assumptions, that the TEJ is neither restricted to the summer months nor to the Northern Hemisphere and that the AEJs exist during most months. We also demonstrate a characteristic weakening of the northern AEJ in midsummer. By using the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis dataset that was recently made available, we are able to describe the important features of the wind and temperature fields in a more spatially and temporally consistent manner than has been done previously. This includes an analysis of the associated wind shears. In recognizing the possible shortcoming of the NCEP reanalysis data, we also use radiosonde observations to confirm some of the results based on NCEP. This combination allows us to relate the evolution of various circulation features and thermal fields to one another and to suggest a physical basis for this evolution and the interrelationships. We also present the first analysis of the seasonal evolution of the precipitation field in the context of the various dynamical features.

The structure of this article is as follows. In section 2 an overview of the dynamical systems affecting West Africa and central Africa is presented. Section 3 describes the data and methodology. Sections 4, 5, 6, and 7 describe the results of the analysis of zonal wind, temperature, meridional wind, and the rainbelt, respectively. In section 8 there is some discussion about the evolution of important dynamical features and their interrelationships. Section 9 presents the summary and conclusions.

2. Overview of dynamical systems influencing the study region

a. West Africa

The area described as West Africa is defined in Fig. 1. The rainfall field over West Africa is characterized by a zone of maximum precipitation that migrates north and south throughout the course of the year. This zone lies to the south of the surface position of the ITCZ and its seasonal excursion roughly parallels the seasonal excursion of the ITCZ. The rainbelt is the loci of disturbances that are dynamically linked to the midlevel AEJ. Westward-propagating mesoscale disturbances, termed cloud clusters, are the dominant convective system. Certain environmental factors, such as vertical wind shear,
buoyant energy, low-level jets, and latitude, determine whether convection organizes into these long-lived systems (Laing and Fritsch 1993). The frequency of cloud clusters and the amount of rainfall associated with them is modulated by transient synoptic-scale African or easterly waves (Houze and Betts 1981; Thompson et al. 1979).

These waves originate as a consequence of a joint baroclinic–barotropic instability associated with the vertical and horizontal shear of the midtropospheric AEJ. The waves, as well as the cloud clusters, are generally confined to a relatively narrow latitudinal zone near and south of the jet (Laing and Fritsch 1993; Norquist et al. 1977; Albignat and Reed 1980; Burpee 1972). This zone corresponds to the region between the axes of the AEJ and the upper-tropospheric TEJ (Tourre 1981; Bounoua 1980). Their growth and development is influenced by several factors, including the magnitude of the horizontal and vertical shear of the AEJ (e.g., Rennick 1981), and latent heat release (Norquist et al. 1977).

Both observations and models have shown that characteristics of the AEJ (particularly horizontal and vertical shear) and the mean zonal flow influence the characteristics of these waves (Burpee 1974; Simmons 1977; Mass 1979; Rennick 1976, 1981; Kwon 1989). Although the shear instability associated with the jet is present throughout the rainy season, the waves appear to contribute to the development of large-scale rainfall systems only during late summer (Reeves et al. 1979; Chen and Ogura 1982; Miller and Lindzen 1992). For the waves to organize rainfall, they must produce moisture convergence and be reasonably close vertically to the moist layer. Past studies have suggested that in the Sahel, these prerequisites are present only in late summer. This fact may help to explain the contrast between the rainfall regimes in June–July and August–September, noted in several studies, and the fact that mesoscale convective complexes are much more frequent during the latter period (Laing and Fritsch 1993), when wave amplitude is larger (Duvel 1990; Grist 2002).

Squall lines, which may develop from isolated convection or from cloud clusters, are also important convective features. They likewise appear to be related to wave activity (e.g., Reed et al. 1977; Payne and McGarry 1977), but the association is inconsistent (e.g., Bolton 1981, 1984; Rowell and Milford 1993). The generation of squalls requires conditional instability, vertical wind shear in the lower troposphere, a shallow moist layer with dry air aloft, and probably a minimum intensity of the AEJ. Growth may be triggered by surface heating, topography or large-scale, low-level convergence, as is found just ahead of the wave trough (Bolton 1981; Mansfield 1977; Rowell and Milford 1993).

Another dominant feature of the circulation during the West African rainy season is the TEJ. During the height of the rainy season it is of the order 20 m s$^{-1}$ at 200 mb and 7°N. Although it tends to be much stronger in wet years, there appear to be some difference of opinions as to its relationship with rainfall. Druyan and Hall (1996) and Reiter (1969) suggest that the increased divergence of its exit region might promote convective rainfall, while Thorncroft and Blackburn (1999) concluded that the TEJ was a response to the convective rainfall.

It is evident that the dynamical features influencing rainfall in the region exhibit strong seasonality. Consequently, a description of the annual cycle should be helpful in understanding rainfall variability in the region.

b. Central Africa

In this article, we use the term central Africa to denote the central equatorial Tropics of Africa, ranging from approximately 5°N to 10°S and extending westward to Cameroon and eastward to the Rift Valley highlands. The geographical limits described here and indicated in Fig. 1 are basically pragmatic. Areas somewhat farther north (Central Africa Republic, Chad, and Sudan) and west (Nigeria and other countries of the Guinea Coast) are meteorologically controlled by the same processes and features that influence Sahelian West Africa and peak rainfall occurs in the boreal summer, when the ITCZ moves to its far northern limits. In contrast, central Africa experiences a strongly bimodal annual cycle, coincident with both the northward and southward passage of the ITCZ. Peak rainfall tends to be in the two transition seasons. This zone may have many common features with eastern equatorial Africa, generally termed East Africa. However, the latter has been comparatively well studied and also contrasts with central Africa in some key climatic features, such as proximity to the Atlantic versus Indian Oceans, the occurrence of rainfall in the boreal summer, and the influence of the Rift Valley highlands and the barrier they provide between the two regions.

Rainfall variability in the region shows some commonalities with that over East Africa. The variance spectra for rainfall throughout the region show very strong peaks on the order of 5–6 yr and generally smaller peaks at roughly 3.5 and 2.3 yr, the same peaks found in the Southern Oscillation (Nicholson and Entekhabi 1986). The variability shows strong links to both the Southern Oscillation and to SST variability in the eastern Atlantic (Nicholson and Entekhabi 1987). The variability is generally out of phase with that in the Sahel [i.e., dry (wet) Sahel corresponds to wet (dry) Central Africa]. Although wave activity has not been identified, rainfall tends to be organized into mesoscale convective systems (MCSs) analogous to those in Sahelian West Africa (Laing and Fritsch 1993; Mohr and Zipser 1996).

Very little is known about the synoptic situations and atmospheric dynamics that control regional rainfall. A few studies concentrating on southern Africa (e.g., Kumar 1978) have shown dynamical teleconnections be-
to gain insight into any longitudinal variation in variables that are attributable to the land–sea contrast in the Southern Hemisphere. Characteristics of interest include the general structure, the intensity and location of the various jet streams, their seasonal development and associated horizontal and vertical wind shear.

In order to highlight possible problems in the reanalysis, key features of the wind field are compared to radiosonde observations. The radiosonde observations were dataset DS353.0 taken from NCAR data archive. A summary of this is included in Fig. 2a. There was sufficient data to enable the 700-mb latitudinal profile for east and west sectors to be compared with the reanalysis for 1962–72. It is noted, however, that there is very poor coverage south of the equator in the western sector. This is important to remember when viewing results from this area. For the 200-mb level, where fewer observations were available, only the annual cycles for key stations could be derived (not shown).

The same sections were used to produce zonal means of the meridional wind and temperature fields for 1958–97. The dynamical features are then related to the development of the rainy season via a comparison of their annual cycles with the mean “structure” of the rainbelt over West Africa. To do this, meridional cross sections depicting rainfall as a function of month and latitude are derived from a rainfall dataset (Fig. 2b) assembled by Nicholson (1986) and recently updated (e.g., Nicholson et al. 2000).

4. The mean zonal wind field

Mean zonal winds for the two sectors are shown in Figs. 3a,b. Three easterly jets are evident over West Africa, an upper-tropospheric easterly jet (the TEJ) around 200 mb and two midtropospheric jets near 650 mb. One of these is the well-known AEJ of the Northern Hemisphere. The second is a Southern Hemisphere jet and has not yet been described in detail. Here we use the terms AEJ-N and AEJ-S to distinguish them. Their annual cycles are also summarized in Fig. 4. Further apparent are the two upper-tropospheric westerly jets of the midlatitudes and a low-level westerly wind maximum near the equator.

In the western sector (Fig. 3a) the AEJ-N is discernible throughout the year, but during the Northern Hemisphere (NH) winter, when it is located around 0°–5°N, it is relatively weak. It strengthens during the NH summer and also shifts to 10°–12°N. Core speeds range from about 8 m s⁻¹ in October–March to 12 m s⁻¹ in June. The altitude of the core also changes during the year. It is at 600 mb in April–September, when the jet is relatively strong. The core is closer to 700 mb during the months of October–March, when the AEJ-N is relatively weak.

The AEJ-S is most clearly defined during September–November. During this time its location is at 700 mb between 5° and 10°S. Mean core speeds reach 8 m s⁻¹
in October and November, but only 4–6 m s⁻¹ in the other months. Hence, it is weaker than the AEJ-N. During December–March there is a low-level (900 mb) easterly maximum at around 20°S, but at other times of the year (April–July) this feature is not very distinct and no semblance of the AEJ-S appears to be a discernible.

In the eastern analysis sector (Fig. 3b) the annual cycle of these jets is similar, although some differences are notable. The AEJ-N is somewhat weaker in this sector, its core speeds barely reaching 10 m s⁻¹ in May and June, then again in September. The jet disappears in November and its location in the NH winter ranges from about 3° to 5°N. Its core is near 600 mb in all months, except October, when it lies near 650 mb.

The AEJ-S is more pronounced in the eastern sector. It is best defined in August–November. Its position ranges from about 5°S in August to 8°S in November. Core speeds range from about 6 m s⁻¹ in August to 10 m s⁻¹ in October. Some semblance of a midtropospheric jet is apparent in December–March, but in April–July it is replaced by a weak lower-level easterly maximum near 800 mb.

The horizontal shear at 600 mb was calculated as the difference in zonal wind (Fig. 3a) between adjacent grid points, which are spaced at 2.5° intervals. The variation of horizontal shear with latitude and month is shown in Fig. 5a. The seasonal migration of the area of maximum shear is clearly associated with the migration of the AEJ-N. Cyclonic shear south of the jet reaches 10–12 × 10⁻⁶ s⁻¹ in the eastern sector. The anticyclonic shear north of the jet, where there is a rapid transition to westerly flow, reaches 14 × 10⁻⁶ s⁻¹ in April and May. The vertical shear associated with the AEJ was calculated as the difference in the zonal wind velocities between 700 and 850 mb. The seasonal migration of the maximum in the vertical shear is determined by the seasonal strengthening and migration of the AEJ-N and the low-level westerlies. It is at its most poleward in August. The vertical shear beneath the jet core is around 5–6 × 10⁻⁴ m s⁻¹ Pa⁻¹ much of the year, but it appears to peak at around 7 × 10⁻⁴ m s⁻¹ Pa⁻¹ in June. Both the horizontal and vertical shears associated with AEJ-S are considerably weaker than the AEJ-N. This reflects both the weaker midtropospheric jet and the weaker westerlies in the Southern Hemisphere.

Figure 6 compares the radiosonde and reanalysis mean annual cycle of zonal wind at 700 mb for both sectors. In general, there is very good agreement between the two. The magnitude, location, and nature of seasonal migration of the AEJ are very close. The good agreement partly reflects the fact that the reanalysis has been produced by the assimilation of radiosonde (and other observations) into the NCEP model. The area of weaker easterlies on the equatorward side of the AEJ-N is also evident in both datasets. It is noticeable that there is increased variability in the radiosonde values at this latitude. This may indicate some longitudinal variation in the wind field. Despite the fact that there are fewer radiosondes in the SH, they still show the formation of the jet around August–September. The radiosondes show the AEJ-S weakening fairly dramatically in November and continuing to move poleward. In the
Fig. 3. (a) Mean zonal wind speed (m s\(^{-1}\)) for Jan–Dec, averaged between 10°W and 10°E for 1958–97. (b) Mean zonal wind speed (m s\(^{-1}\)) for Jan–Dec, averaged between 10° and 30°E for 1958–97.
FIG. 4. Mean intensity and location of jet core (monthly averages) of the (a) AEJ-N and (b) AEJ-S for both sectors. The dashed lines indicate standard deviations from the monthly mean.

FIG. 5. Annual cycle of zonal wind shear. (top) Horizontal shear of the zonal wind at 600 mb in the sector from 10°W–10°E; units are 10^{-2} m s^{-1} and the solid contours are for negative values. (bottom) Vertical shear of the zonal wind, averaged for the layer from 850 to 700 mb in the sector from 10°W–10°E; units are 10^{-2} m s^{-1} Pa^{-1} and the solid contours are for positive values.

eastern sector, the reanalysis depicts this evolution differently with the weakening occurring about two months later than the radiosondes. According to both reanalysis and radiosondes the AEJ-S has all but disappeared by April.

The annual cycle of the magnitude of the TEJ features a distinctive double maximum (Fig. 7). A northern summer maximum occurs in July, when the TEJ is most northward. A second, weaker (around 14 m s^{-1}) maximum occurs in February, when the TEJ is close to its most southern latitude. As it moves north and crosses the equator, its magnitude decreases to around 6 m s^{-1}. At its most northerly point, the magnitude of the TEJ is at around 19 m s^{-1}. The jet weakens as it moves back south. According to Fig. 3b, the TEJ is not a distinct feature in November, but reappears at 8°S and 8 m s^{-1} in December. Through most of the year the TEJ is stronger in the eastern sector.

The apparent appearance of the TEJ in the Southern Hemisphere is interesting. The existence of upper-level easterly jets near (Davies and Sanson 1952) and just south (Bond 1953) of the equator has been noted for some time and they are considered part of the Walker circulation. These Southern Hemisphere easterlies are weaker and not considered part of the TEJ (Reiter 1969). The TEJ is believed to owe its existence to the presence of the Tibetan high and the large meridional temperature gradient that forms between the Tibetan highlands and
the Indian Ocean during the northern summer (Koteswaram 1958). Thus, it is usually considered to be restricted to the NH and the boreal summer (e.g., Hastenrath 1988). However, the steady seasonal migration of the easterly wind maximum in Fig. 7 suggests the same process operates year-round. This and the importance of the longitudinal variation in the jets will be discussed in section 8c.

The equatorial westerlies (Figs. 3a,b) are better developed in the western analysis sector than in the eastern, because they result primarily from the southeasterly trades associated with the South Atlantic high, which become westerly upon crossing the equator. This system is less extensive in the continental sector farther east. The equatorial westerlies are best developed in July–September.

5. Zonal mean temperature

Figure 8 shows a meridional cross section of zonally averaged temperature in the eastern sector (Fig. 8a) and the western sector (Fig. 8b) during each month of the year. Considering the eastern sector, we note that the temperature field is characterized by weak latitudinal gradients in the mid troposphere and upper troposphere from about 20°S to 20°N, and two temperature maxima in the lower troposphere in subtropical latitudes. The maximum in the NH migrates from 10° N in NH winter, when maximum temperatures reach 300 K, to 22°N in NH summer, when temperatures exceed 308 K. The maximum in the SH is generally between to 15° and 20°S. The maximum is most pronounced in October, with maximum temperatures of over 300 K at 15°S. It is least pronounced between April and June when a maximum of less than 300 K is spread over some 20° of longitude.

The NH maximum reflects the intense heating of the Sahara and its southern margin, while its seasonal shift is associated with the seasonal excursion of sub-Saharan rainbelt (see section 7). The temperature gradient resulting from this low-level maximum (Fig. 9) reaches a maximum of 10 K (1000 km)\(^{-1}\) in June, when the jet is most intense and the vertical shear is strongest. Similarly, the maximum in the SH is indicative of the warming of the air above the semiarid Kalahari, prior to the onset of rains.

There is a notable difference between the eastern and western sectors in the SH, which appears to reflect the land–sea contrast. The western region has a low-level temperature maximum at around 10°S. This is probably
Fig. 8. (a) Vertical cross sections of mean temperature (K) for Jan-Dec, averaged between 10°W and 30°E (eastern sector), 1958-97. (b) Vertical cross sections of mean temperature (K) for Jan-Dec, averaged between 10°W and 10°E (western sector), 1958-97.
due to the advection of air warmed by the land to the east. However, the fact that the temperature maximum is elevated to around 850 mb, in August–December appears to indicate that the surface air is also modified by the relatively cool SSTs.

It has been claimed that the AEJ-N forms as a result of the reversal of temperature gradient. We investigated this, and also whether it is the case for the AEJ-S, by comparing the zonal wind shear with the thermal wind \( (U_t) \) between 600 and 925 mb. The zonal wind shear was calculated as the difference in zonal wind between 600 and 925 mb. Values were calculated for all months, for the eastern and western sectors. The monthly means of the basic states from the western and eastern sectors were used in the calculation. The thermal wind was calculated from the following formula:

\[
U_t = \frac{R}{f} \frac{dT}{dy} \ln \frac{600 \text{ mb}}{925 \text{ mb}},
\]

where \( R \) is the gas constant 287 J deg\(^{-1}\)kg\(^{-1}\), \( f \) is the Coriolis parameter at the appropriate latitude, and \( T \) is the mean zonally averaged temperature for the layer between 600 and 925 mb.

If the temperature gradients are responsible for the jets, a strong linear correlation between thermal wind and observed wind shear near the jet location would be expected. However, the values would not be comparable close to the equator, where the Coriolis parameter goes to 0. For this reason, the calculations were restricted to the latitudes 8.75\(^\circ\)–13.75\(^\circ\)N and 8.75\(^\circ\)–13.75\(^\circ\)S. In addition, only winds of magnitude greater than 2 m s\(^{-1}\) were plotted. The average temperature for the layer 600–925 mb was estimated by a polynomial interpolation of the temperature profile to the midpoint (762.5 mb) of the layer.

The upper three panels of Fig. 10 shows the results of the comparison of the wind shear and the thermal wind for the western sector for 1) both the Northern and Southern Hemispheres, 2) just the Northern Hemisphere, and 3) just the Southern Hemisphere. The lower three panels show the same but for the eastern sector. All graphs show strong correlations, with greater than 99% significance. Thus our calculations support the hypothesis that the vertical wind shear (i.e., the AEJ-N) is caused by the temperature gradient. In addition, the calculations further suggest that although the AEJ-S is weaker, it likewise forms as a result of the local reversal of the surface temperature gradient. In light of this it is interesting to note that the slope in Fig. 10 for the western sector in the Southern Hemisphere is much smaller than the others. The western SH section is oceanic. Therefore, the difference in the thermal wind–wind shear relationship may be because the meridional surface temperature gradient is weaker. Alternatively, it may be because the paucity of upper-air observations in the sector make the thermal wind relationship difficult to characterize.

The annual cycle of the AEJ closely resembles that of the maximum temperature gradient (MTG) at 850 mb. Figures 11 and 12 indicate that differences in the surface temperature gradient between the NH and SH are not just in magnitude. Their annual cycles also exhibit different characteristics. Figure 11 shows that in the NH, the poleward migration of the MTG is nearly coincident with its strengthening at the peak of the boreal summer. By contrast, shortly after the southern MTG moves poleward the MTG weakens substantially.

We suggest that the differences in the annual cycles are due to the varying influences of rainfall. Figure 12 shows the annual cycle of rainfall and 850 mb temperature at different latitudes. It is evident in both hemispheres that the rainfall moderates the annual cycle of temperature. This is presumably by increasing the ratio of latent to sensible heat transfer from the surface, although increased cloudiness may also be influential. In the Southern Hemisphere there is more rainfall poleward of the MTG and, hence, more cooling. As a consequence, the MTG (and, in turn, the AEJ-S) weakens. However, poleward of the MTG in the NH (20\(^\circ\)N) there is little rain and little moderation of temperature. Thus, unlike the SH after the onset of the rains, the MTG continues to increase until it is near its maximum poleward location.

6. Mean meridional winds and the ITCZ

Figure 13 shows the mean meridional wind component in the western sector. All 12 months are depicted. In all months a quadrant pattern depicting a Hadley cell circulation is evident. The reanalysis shows that the meridional circulation strengthens in the winter hemisphere, in agreement with previous observations (e.g., Lau 1984). However, this strengthening is less evident for the Southern Hemisphere. Lindzen and Hou (1988) show that this asymmetry in the circulation occurs when thermal heating is displaced from the equator.

The southerly flow near the surface and in the lower atmosphere reflects the southeasterly trade winds on the equatorward flank of the South Atlantic high. It generally extends to 750 or 800 mb and is evident in all months throughout the Southern Hemisphere lat-
Fig. 10. Monthly mean values of zonally averaged vertical wind shear (600–925 mb) against thermal wind for 8.75°N, 11.25°N, 13.75°N, 8.75°S, 11.25°S, and 13.75°S: (a) AEJ-N and AEJ-S, eastern sector; (b) AEJ-N, eastern sector; (c) AEJ-S, eastern sector; (d) AEJ-N and AEJ-S, western sector; (e) AEJ-N, western sector; and (f) AEJ-S western sector.

Fig. 11. (top) Annual cycle of 850-mb MTG for Northern Hemisphere (10°±30°E), location in degrees latitude north (dashed lines) and magnitude (solid lines). (lower) Same for southern 850-mb MTG, only it is shown with a 6-month lag to aid comparison with Northern Hemisphere and the location is in degrees latitude south. Units of MTG are K (1000 km)^{-1}.

7. The structure of the rainbelt

Figures 14a,b show rainfall as a function of latitude and month for the areas 10°W–10°E and 10°–30°E, respectively. These figures are constructed from station rainfall at all stations within the averaging sector. The general characteristics of the tropical rainbelt shown in the cross section are manifestations of the large-scale dynamics governing rainfall. Two main features are evident in both cross sections: the latitudinal excursion of the rainbelt during the course of the year and the asymmetry of the northward and southward portions of this migration. Some other features, however, are representative of regional peculiarities. For example, the maximum near 5°N in June in the western sector is a result of the geographical configuration of the continent: at this latitude, rainfall over the land area is dominated by a small area of peninsulas and highlands in eastern Nigeria and coastal Cameroon that experience extremely heavy rainfall between May and July. Likewise, the relatively low January and February rainfall from roughly 5° to 11°S is related to coastal effects of the Benguela Current and not large-scale dynamics.
The northward excursion occurs during the six months from February to August, when the axis of the rainbelt migrates from roughly 9° or 10°S to 11°N (8°N in the eastern sector). The southward excursion is more rapid, with the axis migrating back to 10°S by November, that is, within 4 months. This asymmetry was noted earlier by Nicholson (1981) and is evident throughout the east–west extent of the continent. In the western sector, the southward excursion is rapidly accelerated from September to October, abruptly ending the rainy season in the central Sahel/Soudan zone. In the eastern sector, the biggest shift occurs between October and November, so that the eastern Soudan zone receives significant October rainfall.

The rainbelt is considerably more “intense” (in terms of rainfall per month) during the southward excursion. In the western sector (Fig. 14a) rainfall is on the order of 350–400 mm month$^{-1}$ from September to November in a large sector within the latitudinal belt 4°–1°N. Rainfall does not exceed 350 mm month$^{-1}$ during the northward excursion. In the eastern sector (Fig. 14b) the area with rainfall in excess of 200 mm month$^{-1}$ is likewise limited to the latter half of the year. As discussed in section 8b this asymmetry may be related to the appearance of both midtropospheric jets during the southward excursion. The intensity suggested by the cross sections in Figs. 14a,b is a function of the strength of the rain-producing processes and the time that a location is affected by the core of the rainbelt. Because of the more rapid southward excursion, there is a shorter, more intense rainy period during the southward excursion than that suggested from the monthly totals at individual stations. Over the tropical Americas, Horel et al. (1989) noticed a similar asymmetry in the timing (but not the intensity) of the march of the convective rains.

The classic interpretation of the seasonal excursion of the rainbelt is that it represents the northward and southward movements of the ITCZ. A comparison with the ITCZ, as defined by wind convergence, is discussed in the following section. A somewhat alternate explanation for the seasonal excursion of the rainbelt and a hypothesis concerning its asymmetry are also presented.

8. Discussion

To examine the relationship between the annual cycle of the AEJ-N and the rainbelt, the mean locations of both are plotted in Figs. 15a,b. These are compared as well with the location of the strongest horizontal temperature gradient and with the ITCZ. The ITCZ location is defined by the shift from northerly to southerly meridional flow.

In the western sector, the seasonal excursion of each of these features is quite similar. The asymmetry of the northward and southward excursion noted earlier for the rainbelt is evident in all of these indices. The jet core lies roughly 4°–5° south of the surface convergence, and the maximum temperature gradient lies in between. The rainfall maximum is about 5° south of the AEJ in most months (it is about 10° to the south for July–September). In the eastern sector (Fig. 15b), the seasonal excursion of each of these parameters is nearly the same as in the western. Although the strong similarity of the seasonal excursions of the wind and temperature parameters in both sectors could be an artifact of the model used to
produce the reanalysis data, the rainfall data are completely independent. The major trends seen in the other parameters, especially the asymmetry of the northward and southward excursions and the small points of contrast between the eastern and western sectors, are also apparent in the rainfall data. Thus, the geographical associations among these parameters appear to be real climatological features of the atmosphere.

a. The ITCZ

The parallel seasonal excursions of the ITCZ, AEJ, and rainbelt suggest a fundamental relationship between all of these parameters. Although Figs. 15a,b show parallels, in the mean, between the ITCZ (i.e., the surface convergence) and the location of the rainbelt, there is clear evidence that the two are decoupled on interannual timescales (Grist and Nicholson 2001). The location of the ITCZ shows little contrast between the wet and dry year composites analyzed in that study, while the location of the core of the AEJ-N varies greatly, as does the location of the rainbelt. Both are displaced equatorward during dry years over West Africa. Attributing the position of the ITCZ as a control on rainfall also fails to account for the asymmetry of the rainbelt during the northward and southward excursion, with the most intense rainfall occurring from August to November, and for the preferential occurrence of large and well-organized disturbances (MCCs) late in the rainy season (Duvel 1990; Laing and Fritsch 1993).

b. African easterly jets

It is interesting to note that the rainbelt is most intense when both the AEJ-N and the AEJ-S are well developed. During this time, late in the Sahelian rainy season, there is a preferential occurrence of large convective events and the rainbelt lies roughly between the two axes of the jets (Figs. 14a,b). There is thus circumstantial evidence that the juxtaposition of the two jets plays a role in determining the dynamics of the rainfall regime and the development of the rainbelt during August–November. Although it should be acknowledged that the wind shear associated with the AEJ-S is relatively weak, the most intense mesoscale convective systems in Southern Hemisphere Africa are on the equatorward side of this jet (Laing and Fritsch 1993; Mohr and Zipser 1996) and these large, organized disturbances are approximately
limited to the latitudes between the AEJ-N and the AEJ-S during the seasons when both jets are present (Fig. 16). Moreover, the MCSs appear to be quite rare in East Africa, in the longitudes where the AEJ-S appears to disappear. Consequently, this connection may be worthy of further exploration.

It has generally been accepted (e.g., Krishnamurti and Pasch 1982; Thorncroft and Blackburn 1999) that the origin of the AEJ-N is basically the temperature gradient induced by the contrast between the Sahara and the humid Guinea coast region to the south. Our results further confirm this and also demonstrate that the origin of the AEJ-S is also related to the surface temperature gradient. In this case, the temperature contrast is produced by semiarid regions of the Southern Hemisphere, compared to the subhumid, perennially vegetated lands to the north.

As suggested in section 5, this implies a feedback between the rainfall regime and the jets. In the case of the AEJ-N, the maximum temperature gradient (and jet core) lies in proximity to the boundary between the dry region north of the jet and the area near and south of the core where rainfall is occurring. The effect of the rainfall is to reduce surface temperatures (as a consequence of soil moisture and rapidly growing vegetation cover). This, together with the northward progression of the solar maximum, displaces the maximum temperature gradient (and the jet core) farther northward. Analogous arguments were used by Webster (1983) to explain the northward advance of the Indian monsoon. In the case of the AEJ-S, there is greater rainfall on the poleward side of the jet. Consequently, the poleward surface temperature cools more and the feedback appears to work in the opposite sense.

There is considerable asymmetry in the northward and southward excursion of the AEJ-N (and rainbelt). This may be partially a consequence of the asymmetry in the surface heating over West Africa. During the northward excursion, which is prior to the Sahel/Soudan rainy season, the region is dry and, hence, comparatively warm. During the rainy season, the juxtaposition of the summer heating of the Sahara and the relatively moist and vegetated surface over the Sahel maintains the temperature maximum at high latitudes. In September, the Sahara rapidly cools as the solar maximum moves to the equator and the rainy season ends, rapidly drying the Sahelian land surface. Consequently, the latitude of maximum temperature (and maximum temperature gradient) shifts abruptly southward after September.

Fig. 14. Cross section of mean rainfall (mm month$^{-1}$) as a function of month and latitude, averaged for (a) western sector, 10°W–10°E (183 stations are used to produce this cross section), and (b) 10°–30°E (420 stations are used to produce this cross section). The location of the AEJ-N and AEJ-S is indicated by the black solid lines. Because of the lack of stations (Fig. 2b) the contours are not plotted south of the equator in the western sector.
c. Tropical easterly jet

We further examine the relationship between the seasonal cycles of rainfall and the TEJ in Fig. 17. The figure shows the precipitation and the zonal wind at 200 mb for August and February 1989. The zero contour of the $du/dy$ field near the easterly flow is included to show the axis of the TEJ. The distinctive Asian and West African–Atlantic branches of the TEJ are evident. Although this is just an example from one year, the features of relevance are common to most years. In fact, the August plot is very similar to Fig. 8 in Koteswaram (1958). We identify three distinct regions (A, B, and C). Region A, central Africa, is in the left exit region of the main Asian branch of the TEJ. It is therefore likely to be a region of enhanced upper-level divergence that could promote convective activity, as suggested by Reiter (1969). This suggests that the variability of the TEJ may influence rainfall variability in equatorial Africa.

This is not the situation in region B, West Africa (west of 10°E), where the main area of convective activity is to the north of the jet axis. The location of convective activity relative to the TEJ is consistent with the work of Thornicroft and Blackburn (1999). They attribute the formation of the TEJ to the easterly deflection of the southward outflow of convection due to the Coriolis force. The implication is that in region B, variability in the TEJ is a response to, and not a cause of, rainfall variability. This is supported by differences maps of wind field between West African wet and dry years. These show that the largest differences in the TEJ are west of 0°E. For the TEJ to promote convective activity in this region, it would have to be through enhancing convection as the right entrance region of the West African–Atlantic branch of the TEJ, as suggested by Druyan (1998).

Region C, in the Southern Hemisphere, is not usually considered part of the TEJ. However, it appears to be the Southern Hemisphere version of the process Thornicroft and Blackburn (1999) described. Convection located around the southeastern coast of Africa creates an upper-level outflow. The northward branch of the outflow is deflected westward by the Coriolis force. This explains the seasonal migration of the 200-mb easterly wind maximum (Fig. 7) and why it reaches a minimum as it crosses the equator.

Summing up the TEJ discussion, it is noted that the strong zonal flows near regions B and C are a response to convective activity. The convective activity in region A, however, may be enhanced or promoted by the upper-level divergence associated with the TEJ (Asian branch) exit region. It is also noted that the features of the 200-mb zonal wind described here and the mech-
Fig. 17. Contours of 200-mb $u$ for 1989; the solid line is the zero contour of the $du/dy$ field (indicating the TEJ axis). Shading represents precipitation (mm day$^{-1}$). Wind data are from NCEP and rainfall data are from first authors archive: (a) Aug, and (b) Feb. Significance of letters is explained in the text.
anisms attributed to them are not covered in classic descriptions of the TEJ (e.g., Koteswaram 1958; Hastenrath 1988).

9. Summary and conclusions

The annual cycle of the basic state over West Africa and central Africa has been described in some detail. The NCEP reanalysis of zonal wind compared well with available radiosonde observations, adding confidence to the climatologies produced. The increased resolution of the dataset provided clear depiction of such features as the AEJ-N, AEJ-S, ITCZ, TEJ, and the monsoonal jet. The seasonal migration of these features is asymmetric in that the northward excursion occurs much more slowly than the southward. This creates a similar asymmetry in the seasonal cycle of rainfall, with the rainy season taking on a very abrupt end. There is also an asymmetry in the intensity of the rainfall belt, which is substantially more intense during the months when both the AEJ-N and the AEJ-S are present (i.e., August–November). The belt of intense rains is roughly bound by the cores or the two jets during this period. The cores also roughly define the zones where mesoscale convective systems occur over equatorial Africa. This suggests that the AEJs play an important role in the development of the rainy season.

Observations and calculations support the claim that the AEJ-N is formed from the local reversal of the surface temperature gradient. This study also shows that, although weaker, the AEJ-S is formed in a similar manner. A comparison of the MTGs in the Northern and Southern Hemispheres revealed that precipitation appears to moderate the MTG. It thus, in turn, moderates the annual cycle of the jets. In the Northern Hemisphere, rainfall modulates the surface temperature at the latitude and south of the MTG but not over the practically rainless Sahara. The net effect is that the jet strengthens as it is displaced north after the onset of the rains. In the Southern Hemisphere the rainfall is more abundant poleward of the MTG. Consequently, shortly after the onset of the rains and as the jet begins to move poleward, it weakens to the point where it is not a very distinct feature.

From the above discussion it is clear that there is considerable feedback between the two African easterly jets and the tropical rainbelt. Not only does the rainy season influence the jets via the MTGs, but we suggest that the jets also modulate the development of the rainy season. The clearest example is that the AEJ-N controls the development and organization of rain-bearing systems via the generation of waves. This may also be the case with the AEJ-S, but that has not yet been established.

Our results similarly suggest that the tropical rainbelt also has some influence on the tropical easterly jet at 200 mb. The Asian branch of the TEJ is forced remotely by the Tibetan high. This may act as a mechanism of rainfall variability over central Africa, which is in the jet’s left exit region. Over West Africa, the TEJ seems at least partly a response to the equatorward outflow of convection just to its north. This process would appear to explain the TEJs distinct semiannual cycle, its weakening over the equator, and its appearance in the Southern Hemisphere in January and February.

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