The West African monsoon dynamics, Part II : The "pre-onset" and the "onset" of the summer monsoon

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ABSTRACT

The arrival of the summer monsoon over West Africa has been documented by using daily gridded rainfall data and NCEP–NCAR reanalyses during the period 1968–90, and OLR data over the period 1979–90. Two steps have been characterized through a composite approach: the preonset and the onset of the summer monsoon.

The preonset stage corresponds to the arrival in the intertropical front (ITF) at 15°N, that is, the confluence line between moist southwesterly monsoon winds and dry northeasterly Harmattan, bringing sufficient moisture for isolated convective systems to develop in the Sudano–Sahelian zone while the intertropical convergence zone (ITCZ) is centered at 5°N. The mean date for the preonset occurrence is 14 May and its standard deviation is 9.5 days during the period 1968–90. This leads to a first clear increase of the positive rainfall slope corresponding to the beginning of the rainy season over this Sudano–Sahelian area.

The onset stage of the summer monsoon over West Africa is linked to an abrupt latitudinal shift of the ITCZ from a quasi-stationary location at 5°N in May–June to another quasi-stationary location at 10°N in July–August. The mean date for the onset occurrence is 24 June and its standard deviation is 8 days during the period 1968–90. This leads to a second increase of the positive rainfall slope over the Sudano–Sahelian zone signing the northernmost location of the ITCZ and the beginning of the monsoon season. This abrupt shift occurs mostly between 10°W and 5°E, where a meridional land–sea contrast exists, and it is characterized by a temporary rainfall and convection decrease over West Africa. Preonset dates, onset dates, and summer rainfall amount over the Sahel are uncorrelated during the period 1968–90.

The atmospheric dynamics associated with the abrupt ITCZ shift has been investigated. Between the preonset and the onset stages, the heat low dynamics associated with the ITF controls the circulation in the low and midlevels. Its meridional circulation intensity is the highest at the beginning of the monsoon onset. This can lead to 1) increased convective inhibition in the ITCZ through intrusion of dry and subsiding air from the north, and 2) increased potential instability through a greater inland moisture advection and a higher monsoon depth induced by a stronger cyclonic circulation in the low levels, through higher vertical wind shear due to westerly monsoon wind and midlevel African easterly jet (AEJ) increases, through enhancement of the instability character of the AEJ, and through increased shortwave radiation received at the surface. During the monsoon onset, once the rainfall minimum occurred due to the convective inhibition, the accumulated potential instability breaks the convective inhibition, the inertial instability of the monsoon circulation is released, and the associated regional-scale circulation increases, leading to the abrupt shift of the ITCZ. Then the ITCZ moves north up to 10°N, where thermodynamical conditions are favorable.

It is suggested by the authors that the abrupt shift of the ITCZ, initiated by the amplification of the heat low dynamics, could be due to an interaction with the northern orography of the Atlas–Ahaggar Mountains. Subsidence over and north of this orography, due to both the northern branches of the heat low and of the northern Hadley-type cell, contributes to enhance the high geopotentials north of these mountains and the associated northeasterly winds. This leads to the development of a leeward trough that reinforces the heat low dynamics, maintaining an active convective ITCZ through enhanced moist air advection from the ocean, increasing the northern Hadley circulation, which reinforces the high geopotentials and the interaction with the orography through a positive feedback. The fact that an abrupt shift of the ITCZ is only observed on the western part of West Africa may result from the enhancement of moisture advection, which comes from the west and has a stronger impact west of the Greenwich meridian.

The northwest–southeast orientation of the Atlas–Ahaggar crest can induce the interaction with the heat low, first in the east where the mountains are nearer to the ITF than in the west, and second in the west. Another consequence of the possible orography-induced interaction with the atmospheric circulation is that the induced leeward trough, increasing the cyclonic vorticity in the heat low, may stimulate moisture convergence in the oceanic ITCZ near the western coast of West Africa.

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1. Introduction

Rainfall over West Africa is controlled by the advection of moisture from the Gulf of Guinea in the low levels of the atmosphere. Following the seasonal ex-

cursion of the sun, the monsoon develops over this part of the African continent during the northern spring and summer, bringing the intertropical convergence zone (ITCZ) and the associated rainfall maxima to their northernmost location in August (Hastenrath 1995). This is the time for the rainy season in the Sahel. Agriculture in the Sudano–Sahelian zone is heavily dependent on the seasonal characteristics of rainfall, that is, onset, length and termination of the rainfall season, seasonal rainfall amount, and intraseasonal rainfall distribution during the rainy season. The onset of ''useful'' rains, that is the first rains sufficient to ensure enough moisture in the soil at the time of the planting and not followed by prolonged dry spells that could prevent the survival of seedlings after sowing, is certainly the major point for agriculturists. However, there is not a unique definition for the date of the onset of the rainy season. It can be defined from traditional to semiempirical and scientific techniques, depending on local meteorological conditions, and leading for any particular year to a high dispersion of these dates for the same location (Ati et al. 2002).

The objective of this study is twofold. First, at a regional scale, we will characterize two main steps in the seasonal evolution of the monsoon over West Africa, that is, 1) what we call the ''preonset'' of the summer monsoon, defining the beginning of the rainy season over the Sudano–Sahelian zone based on the northward migration of the northern limit of the southwesterly winds of the monsoon [called the intertropical front (ITF)], 2) the real ''onset'' of the summer monsoon characterized by an abrupt northward shift of the ITCZ from 5° to 10°N (Sultan and Janicot 2000; Le Barbé et al. 2002), and leading to major changes in the atmospheric circulation over West Africa. Second, we will investigate in detail the atmospheric circulation changes associated with the summer monsoon onset and we will propose a mechanism to explain the occurrence of the abrupt shift of the ITCZ. Due to the availability of the different datasets, this study is based on the period 1968–90, a dry one over West Africa compared to the long-term mean (Hastenrath 1995).

2. Datasets

Independent datasets have been used to investigate the preonset and onset of the West African monsoon. Two of them describe rainfall and convection variability, two others are the U.S. and European reanalyses. They document the period 1968–90 or the subperiod 1979–90. We will show that all the results obtained with these different datasets are strongly consistent. As they come from independent sources (rain gauge amounts, satellite measurements, atmospheric variables from radiosoundings, pibals, etc.), it gives a high level of confidence to our results.

a. The IRD daily rainfall

Daily rainfall amount at stations located on the West African domain 3°–20°N, 18°W–25°E have been compiled by the Institut de Recherche pour le Developpement (IRD), the Agence pour la Securite de la Navigation Aerienne en Afrique et a Madagascar (ASECNA, and the Comite Interafricain d'Etudes Hydrauliques (CIEH). These data are available for the period 1968–90, including more than 1300 stations from 1968 to 1980, and between 700 and 860 for the period 1981–90. These daily values were interpolated on the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) 2.5° × 2.5° grid, by assigning each station daily value to the nearest grid point and averaging all the values related to each grid point. They were also interpolated in time, related to NCEP–NCAR daily wind fields since daily rainfall amounts were measured between 0600 LST of the day and 0600 LST of the following day. We applied a time lag of 12 h between the average time of the NCEP–NCAR daily values (0900 UTC) and an approximated average time of ''daily'' precipitation over the West African continent [2100 LST; Duvel (1989) indicates a maximum of high cloud coverage over land between 1800 and 0000 LST. Sow (1997) points out a maximum of half-hourly precipitation over the Senegal between 1700 LST and the end of the night, depending on the stations]. The greatest density of stations is located between the latitudes 5°–15°N. Data on latitude 17.5°N can also be taken into account since 30 to 45 stations are available.

b. The NOAA/OLR dataset

Since 1974, launching polar orbital National Oceanic and Atmospheric Administration (NOAA) Television Infrared Observation Satellite (TIROS) satellites has made it possible to establish a quasi-complete series of twice-daily measures of outgoing longwave radiation (OLR), at the top of the atmosphere and at a resolution of 2.5° latitude–longitude (Gruber and Krueger 1974). The interpolated OLR dataset (Liebmann and Smith 1996) provided by the Climate Diagnostics Center has been used here. In tropical areas, deep convection and rainfall can be estimated through low OLR values. Local hours of the measures varied during the period 1979–90 between 0230 and 0730 in the morning and between 1430 and 1930 in the afternoon. Since the deep convection over West Africa has a strong diurnal cycle, the sample of daily OLR based on two values separated by 12 h is enough to get a daily average. Moreover this dataset has already been widely used for tropical studies. This dataset has been used over the period 1979–90 to validate and extend the results obtained with the IRD rainfall dataset, but also during the period 1979–2002 to address a particular issue (see Fig. 5).

c. The NCEP–NCAR reanalyses

NCEP–NCAR have completed the first version of a reanalysis project with a current version of the Medium-Range Forecast (MRF) model (Kalnay et al. 1996). This dataset consists in a reanalysis of the global observational network of meteorological variables (wind, temperature, geopotential height, humidity on pressure levels, surface variables, and flux variables like precipitation rate) with a "frozen" state-of-the-art analysis and forecast system at a triangular spectral truncation of T62 to perform data assimilation throughout the period 1948 to present. This enables us to circumvent the problems of previous numerical weather prediction analyses due to changes in techniques, models, and data assimilation. Data are reported on a $2.5° \times 2.5°$ grid every 6 h (0000, 0600, 1200, and 1800 UTC), on 17 pressure levels from 1000 to 10 hPa, which are good resolutions for studying synoptic weather systems. We used data covering the period 1 June–30 September, from 1968 to 1990, with one value per day by averaging the four outputs of each day. An improved version of this reanalysis project is now available (Kanamitsu et al. 2002) but it was not used here because we need to work on the longest available dataset and this one only begins in 1979.

d. The ECMWF Reanalyses (ERA-15)

The European Centre for Medium-Range Weather Forecasts (ECMWF) completed a first reanalysis project (ERA-15), which used a frozen version of their analysis-forecast system, at a triangular spectral truncation of T106 with 31 levels in the vertical, to perform data assimilation using data from 1979 to 1993 (Gibson et al. 1997). Compared to NCEP–NCAR reanalyses, there are 17 pressure levels from 1000 to 10 hPa, with an additional level at 775 hPa and a missing one at 20 hPa. According to our objectives, daily data have been interpolated on the $2.5° \times 2.5°$ NCEP–NCAR grid. The ERA-15 dataset has been used in comparison to the NCEP–NCAR dataset in order to evaluate the uncertainty of the wind field produced by the U.S. reanalyses. Previous studies using these two reanalysis datasets already demonstrated their accuracy over West Africa for describing both interannual and synoptic timescale variability, if we consider the period after 1968 (Diedhiou et al. 1999; Poccard et al. 2000; Janicot et al. 2001). The comparisons done here also showed high consistency between these two reanalysis datasets, so only results obtained from the NCEP–NCAR reanalysis, which covers the longest period 1968–90, will be shown in this paper.

3. The detection of the preonset and the onset of the summer monsoon

a. The mean rainfall fields

Figure 1 sums up the main features of the mean meridional excursion of the axis of maximum rainfall associated with the ITCZ, with the corresponding 925-hPa wind field superimposed. The black line over West Africa represents the zero isoline of the zonal component so as to delineate the domain of the monsoon winds (i.e., westerly component) and detect the location of the ITF in the north. During April and the first half of May (Fig. 1a), the ITCZ is centered between the equator and the southern coast of West Africa, extending over both the Guinea gulf and the Guinean coast region south of 10°N. The rainfall field associated with the northern part of the ITCZ is captured by the IRD rain gauge data measurements over the land and corresponds to the first part of the first rainy season over the Guinea coast region. The isoline of 4 mm day^{-1} is located around 8°N. The corresponding wind field at 925 hPa is characterized over West Africa by the confluence along the ITF of the southwesterly monsoon winds and the northeasterly dry winds called Harmattan. The monsoon winds are controlled by the pressure gradient between the low pressures of the heat low centered along the ITF and the oceanic high pressures of the Santa Helena anticyclone. Harmattan winds are controlled by the pressure gradient between the heat low and the Libyan and Azores anticyclones. The area delineated by the zero isoline of the zonal wind component, chosen to characterize the monsoon winds, is limited at this stage of the seasonal cycle, south of 15°N (the latitude of the ITF). During the second half of May and in June (Fig. 1b), the ITCZ remains at a quasi-stable location around 5°N. Convection increases over land with rainfall maxima centered around the main orographic highs of West and central Africa and the isoline 4 mm day^{-1} located between 10° and 12°N. This is the second part of the first rainy season over the Guinea coast region. The ITF is now positioned between 15° and 20°N with its northernmost latitude between 0° and 5°E. During the first half of July (Fig. 1c), the ITCZ has already shifted to the north, reaching a second quasi-stable state around 10°N, while rainfall maxima are still present at 5°N along the Guinean coast. This transition stage is then marked by rainfall minima between 6° and 8°N, in particular over the Ivory Coast along 5°W characterizing a climatological feature of the rainfall field on this area called the "V Baoule."[1] At the same time, the area of westerly winds go on its extension over land and now over the tropical Atlantic between 5° and 18°N, the ITF reaching 20°N at its northernmost latitude. Finally, during the second half of July and into August (Fig. 1d), the ITCZ remains along 10°N with enhanced precipitation and the isoline of 4 mm day^{-1} reaching 15°N. A rainfall minimum is now evident along the Guinean coast region, associated to the so-called little dry season. The monsoon season is fully developed, due to the highest pressures in the southern tropical Atlantic and the

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[1] This comes from a "V" shape of the annual rainfall isolines due to a southward extension of low values along 5°W.

FIG. 2. Mean 1968–90 rainfall time series (mm day^{-1}) computed on the grid points located between 10°W and 10°E, along 15°N. The line represents the time series filtered to remove variability lower than 10 days. The two arrows localize the breaks in the positive rainfall slope.

northernmost location of the heat low over land. This is when the westerly wind area has its largest extension.

Figure 2 depicts the corresponding daily rainfall time series over the Sudano–Sahelian region, by averaging the rainfall values for the grid points located between 10°W and 10°E along 15°N, that is over the longitude band where the meridional land–sea contrast exists. The first-order zonal symmetry of the mean rainfall fields and of the African monsoon system enables us to work on longitude-averaged variables. Figure 2 clearly shows the progressive rainfall increase between April and August. It is however possible to detect two different steps in this mean seasonal cycle, a first one around mid-May with a first increase of the positive rainfall slope, and a second one around late June with a second acceleration of the seasonal cycle leading to the rainfall maxima of August. In the following section we will show that the first step is connected to the ITF crossing 15°N, and to the arrival of the monsoon winds advecting moist air onto the Sahelian latitudes. We will then consider this step as the preonset of the summer monsoon because it corresponds to the beginning of the rainy season on the Sudano–Sahelian region, a rainfall regime with a significant amount at the regional scale while the ITCZ is still located at 5°N. The rainfall increase of the Sudano–Sahelian zone then results from the intensification of the first rainy season over the Guinean coast region and the northward extension of the rainbelt associated with the ITCZ. We will also demonstrate that the second step, around late June, can be clearly associated with the abrupt northward shift of the ITCZ from 5° to 10°N and with major changes in the atmospheric circulation over West Africa signing at this time the development of the meteorological summer monsoon system over West Africa. This is the reason why we will call this second step the summer monsoon onset.

FIG. 1. Mean 1968–90 daily rainfall fields (mm day^{-1}) averaged respectively (a) from 1 Apr to 15 May, (b) from 15 May to 30 Jun, (c) from 1 to 15 Jul, and (d) from 16 Jul to 31 Aug. Rainfall values are from individual rain gauges gridded on a 2.5° mesh. Only rainfall values greater than 4 mm day^{-1} are colored. White areas are due to missing data. The NCEP–NCAR 925-hPa wind field is expressed in vectors and its scale (m s^{-1}) is displayed below. The black line rep-

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resents the zero isoline of the zonal wind component to delineate the domain of the monsoon wind. Thin curves display mean sea level pressure isolines (expressed in hPa minus 1000).

In the following two sections, we have applied a composite method to highlight these two steps. The idea for a better detection of the monsoon preonset (as well as the onset) is that, as this step occurrence can have a high dispersion in time, a classic average of any meteorological signal (as in Fig. 2 for instance) could be too strongly smoothed due to this dispersion. So our approach has been to define the step occurrence for each year Y, called $t_0(Y)$, and then to compute the average of any variable by taking the different $t_0(Y)$ as the same reference date t_0 before performing the average. We hope to enhance the rainfall and convection signals, as well as any meteorological signal that could be tightly linked to this step.

b. The composite signal of the summer monsoon preonset

The meteorological signals associated to the preonset stage of the summer monsoon are rather weak at a regional scale because the start of the rains over the Sudano–Sahelian zone is seldomly abrupt and is usually preceded by a succession of isolated showers of uncertain intensity with intermittent dry periods of varying duration (Ati et al. 2002). We based our approach on the idea that local convection will begin to be initiated more or less regularly when a sufficient amount of moisture advected by the southwesterly monsoon winds will be present over the Sahel. At the regional scale considered here, the latitude of the ITF, represented by the northern boundary of the 925-hPa zonal wind zero isoline, could be a good indicator of that. For each year Y between 1968 and 1990 we then define the date $t_0(Y)$ when the 925-hPa zonal wind component averaged over 10°W–10°E equals zero at 15°N, going from negative to positive values. The zonal wind time series have been filtered to remove high-frequency fluctuations and we only consider the evolution of a smooth seasonal cycle. This prevents us from defining too early a date due to a short wind fluctuation giving positive values and coming back rapidly to negative ones, meaning that the ITF is not yet maintaining north of 15°N. We then computed the averaged time series for the period 1968–90 by using each $t_0(Y)$ as the same reference date. We performed the same averaging procedure for the wind modulus at 15° and the rainfall at 15°N.

Figure 3 shows the corresponding composite time series computed between $t_0 - 60$ days to $t_0 + 170$ days. The average date t_0 is the 14 May and the standard deviation on the period 1968–90 is 9.5 days. Due to the definition of the method, we see in Fig. 3b that the date t_0 corresponds to the reversal of sign of the zonal wind component at 925 hPa, going from negative values due to the Harmattan occurrence to positive values due to the monsoon winds occurrence, in association with the northward progression of the ITF. It also corresponds to a minimum of the wind modulus (Fig. 3b), and to a maximum of both the relative vorticity and the wind

FIG. 3. (a) Mean composite time series of daily rainfall (mm day^{-1}) averaged over 10°W–10°E along 15°N, averaged over the period 1968–90 by using as the reference date the date for each year when the (10°W–10°E) averaged zonal wind component at 15°N equals zero going from negative to positive values. Bars represent the raw values and the curve the low-filtered values to remove rainfall variability lower than 10 days. Values are presented from $t_0 - 60$ days to t_0 plus 170 days. (b) Same as (a) but for the unfiltered zonal wind component (black line) and the unfiltered wind modulus (dotted line), expressed in m s^{-1}. In both (a) and (b) the vertical line localizes the date t_0 (the mean date over the period 1968–90 is 14 May).

convergence (not shown). The corresponding rainfall time series (Fig. 3a) now depicts a clear break at t_0 with a strong increase of the positive rainfall slope, meaning that the arrival of the ITF at 15°N can be considered a meaningful signal for the beginning of the rainy season over the Sudano–Sahelian zone.

c. The composite signal of the summer monsoon onset

A similar approach has been used to highlight the meteorological signal associated with the summer monsoon onset corresponding to the abrupt northward transition of the ITCZ. This transition is highlighted when computing time–latitude diagrams of daily rainfall values averaged over the longitudes 10°W–10°E. The zonal distribution of rainfall over West Africa (Fig. 1) enables us to use such diagrams without any significant loss of information. Figure 4a depicts such an example for the year 1978. This rapid shift of the ITCZ is clearly identifiable on this type of diagram. It leads to a sharp reversal of the meridional rainfall gradient over West Af-

FIG. 4. (a) Time–lat diagram from 1 Mar to 30 Nov 1978 of daily rainfall (mm day^{-1}), averaged over 10°W–10°E and filtered to remove variability lower than 10 days. Values greater than 5 mm day^{-1} are colored. (b) Time sections of diagram (a) at 5° (green curve), 10° (black curve), and 15°N (red curve). In (a) and (b), the vertical line localizes the date selected for the ITCZ shift (17 Jun). (c) Composite time–latitude diagram of daily rainfall (mm day^{-1}) averaged over 10°W–10°E, filtered to remove rainfall variability lower than 10 days, and averaged over the period 1968–90 by using as the reference date the shift date of the ITCZ for each year. Values are presented from t_0 (the shift date) −90 days to t_0 + 140 days. Values greater than 5 mm day^{-1} are colored. (d) Time sections of diagram (c) at 5° (green curve), 10° (black curve), and 15°N (red curve). In (c) and (d), the vertical line localizes the date of the ITCZ shift at t_0 (the mean date over the period 1968–90 is 24 Jun).

rica, which is also evident in Fig. 4b by the crossing of the rainfall time series at 5° and 10°N.

As for the detection of the monsoon preonset, a quasi-objective method can be built up to define a date for the ITCZ latitudinal shift for each year between 1968 and 1990. An empirical orthogonal function (EOF) analysis (Richman 1986) has been performed on time–latitude diagrams of daily rainfall values averaged over 10°W–10°E, for each year from 1 March to 30 November. Most of the rainfall variance decomposed by the EOF analysis is explained by the two first components. The first one (about 91% of the variance in 1968–90) is highly correlated with the rainfall time series at 10°N (correlation of 0.9 in 1968–90) and the second one (about 9% of the variance in 1968–90) is highly correlated with the rainfall time series at 5°N (correlation of 0.75 in 1968–90). The rainfall indexes (i.e., the rainfall time series) at 10° and at 5°N can then be used to sum up rainfall variability over West Africa and to define a date for the ITCZ shift. For few years, especially the dry ones like 1983 or 1984, we must have considered the rainfall index at 7.5° instead of 10°N because the axis of maximum rainfall is located at a more southern location in summer. The time series of the rainfall in-

dexes at 5° and 10°N for 1978 are shown in Fig. 4b. A rainfall maximum occurs during May–June when the ITCZ is located at 5°N. The abrupt shift of the ITCZ from 5° to 10°N can be defined by simultaneously, a decrease of the 5°N rainfall index and an increase of the positive slope of the 10°N rainfall index. Because of the lag between these two moments, an uncertainty of a few days remains. So we look for an increase of the positive slope of a similar rainfall index at 15°N during this time to specify an only date. For 1978 the date of 17 June has been selected (see vertical line in Fig. 4b).

This method has been used to define a date of the ITCZ shift for each year from 1968 to 1990. The mean date t_0 found for this shift over the period 1968–90 is 24 June and the standard deviation is 8.0 days. Correlations computed between the preonset dates and the onset dates ($r = 0.01$), as well as with the summer rainfall amount over the Sahel ($r = -0.16$ for the preonset dates, and $r = -0.21$ for the onset dates) are not significant over the period 1968–90, indicating that independent mechanisms control these different variables. To provide a more precise statistical distribution of these ITCZ shift dates, we have extended this analysis over
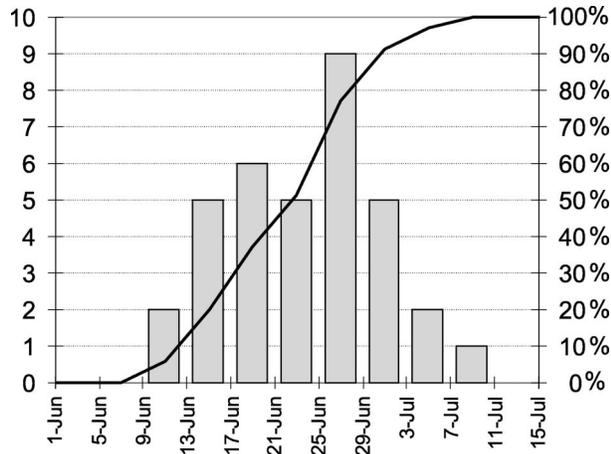
FIG. 5. Histogram of the ITCZ shift dates over the period 1968–2002 (bars; left scale) defined from the NOAA/OLR dataset, and the cumulative distribution function (line; right scale).

the period 1968–2002 by using the NOAA/OLR data. We first made sure that we can apply a similar approach to define these dates from OLR indexes at 5° and 10°N by verifying that we find dates similar or very close to those found with the IRD rainfall indexes during the common period 1979–90, then we defined the ITCZ shift dates from the OLR values over the period 1991–2002. Figure 5 shows the histogram of the ITCZ shift dates over the period 1968–2002 as well as the corresponding cumulative distribution function. Over this period, the averaged date is the same (24 June) and the standard deviation is very similar (7.0 days). This distribution can be used as the first step of a probabilistic forecast of the summer monsoon onset date over West Africa, based on a 35-yr dataset specific to a dry period compared to the long-term mean (Hastenrath 1995). Le Barbé et al. (2002) showed that the mean date of the ITCZ shift is not significantly different from the previous wet period 1950–70. So our probability distribution function may also be valuable for a different climate state, but as the approach of Le Barbé et al. is rather different from our approach, this point should be investigated more precisely; however, this is out of the scope of this paper. Hereafter, all the computations have been based on the period 1968–90 where the ITCZ shift dates have been defined from the IRD rainfall indexes, because we get a more precise signal from daily rainfall amounts measured at the station network than from twice-daily OLR measurements. We checked however that the conclusions infered from the results presented below over the period 1968–90 are confirmed over the period 1968–2002.

Figure 4c shows the composite of the mean 10°W–10°E daily rainfall values averaged over the period 1968–90 by using the ITCZ shift date for each year as the respective reference date. This figure points out latitude–time rainfall variations between t_0 (the shift date) minus 90 days and t_0 plus 140 days. Figure 4d shows the cor-

responding rainfall indexes at 5°, 10°, and 15°N. The rainfall maximum at 5°N, also evident at 10° and 15°N, occurs about 10 days before t_0. Then rainfall decreases slightly, but over West Africa it only decreases to a relative minimum value at the beginning of the ITCZ shift. At t_0, the shift is detected by a new positive slope of the rainfall index at 10° and at 15°N. The ITCZ reaches the latitude 10°N about 10–20 days after t_0 where rainfall increases until mid-August. The withdrawal of the ITCZ is quite sharp too, except that the June rainfall maximum at 5°N has no counterpart in late summer when the monsoon retreats equatorward and we do not observe any rainfall minimum between the summer monsoon season and the second rainy season along the Guinea coast. This also offers a different view from the progression of the northern limit of the ITCZ (see the 1–4 mm day^{-1} isolines), which has a more gradual latitudinal variation during the onset than during the withdrawal.

Figure 6 shows similar composite time–latitude diagrams but for OLR values and for meridional cross sections at particular longitudes, from 20°W to 10°E. Using OLR values enables us to test the previous results on an independent dataset and to consider a wider region without any spatial gap. At 0° (Fig. 6c) and at 10°W (Fig. 6b), the OLR patterns are very consistent with the rainfall (Fig. 4c). Again we see the shift of the ITCZ around t_0, from the latitude 5°N at the time of the first rainy season over the Guinea coast region to the latitude 10°N at the time of the summer monsoon season over the Sudano–Sahelian region. This shift, as seen through the OLR values, is again concomitant with a temporary decrease of convection over all the latitudes of West Africa at this longitude. Similar results have been obtained at 5°E (not shown). At 10°E (Fig. 6d), the OLR pattern is a bit different. At this longitude as well as for more eastward longitudes crossing central Africa, the convection decrease at t_0 is still evident but the ITCZ shift does not occur anymore. The ITCZ progresses regularly to the north before t_0 and stays centered around 6°N after t_0 while its northern boundary extends northward during the first half of summer. Finally at 20°W over the ocean, the northward excursion of the ITCZ is even more regular with no latitudinal shift and no clear convection decrease at t_0. These results suggest that the ITCZ shift leading to the abrupt summer monsoon onset is working well on the longitudes where a land–sea contrast exists between 10°W and 5°E, that is where the concept of the monsoon system, as a large-scale cross-equatorial atmospheric circulation from an oceanic basin on one side and a land area on the other side, is well established. Outside this longitudinal band, the ITCZ loses this particular behavior to show a rather progressive meridional excursion.

Although observed rainfall data are used in this work, the consistency of the seasonal cycle of precipitations in the NCEP–NCAR and ECMWF reanalyses has been examined to estimate the accuracy of the associated temporal fluctuations of the reanalysis winds (Sultan 2002;

FIG. 6. Composite time–lat diagrams of daily OLR (W m^{-2}), filtered to remove variability lower than 10 days, and averaged over the period 1979–90 by using as the reference date the ITCZ shift date for each year. Values are presented from t_0 (the shift date) 90 days to t_0 + 140 days. Values lower than 240 W m^{-2} are shaded. (a) 20°W; (b) 10°W; (c) 0°; (d) 10°E. In (a)–(d) the vertical line localizes the date of the ITCZ shift at t_0 (the mean date over the period 1968–90 is 24 Jun).

not shown). It appears that both reanalyses produce a very good timing of the ITCZ shift, beginning at the date t_0 defined from the IRD rainfall indexes; however, if the ERA rainfall field exhibits a very clear abrupt shift, it is less marked in the NCEP–NCAR rainfall field because after the monsoon onset, precipitation increases over the Sahel band but it remains rather high along the Guinea coast and does not very clearly show the little dry season occurrence in this area (not shown). We know that precipitation is derived solely from the model fields forced by the data assimilation (classified as the C-type variable) whereas the winds are strongly influenced by observed data (classified as the A-type variable, the most reliable class; Kalnay et al. 1996). So despite the slight discrepancy of the NCEP–NCAR rainfall fields, and also because we checked the consistency between the NCEP–NCAR and ERA data, the following results based on the wind fields can be considered to be reliable.

4. The atmospheric dynamics of the monsoon onset

In the previous section we have shown that it is necessary and possible to significantly differentiate what we call the summer monsoon preonset (i.e., the beginning

of the rainy season over the Sudano–Sahelian zone characterized by the arrival of the ITF at 15°N while the ITCZ is still centered at 5°N) and what we call the summer monsoon onset (i.e., the ITCZ shift from 5° to 10°N). The ITCZ shift appears as a main scientific issue for the understanding of the West African monsoon dynamics but it is also a key point for agriculturists. Ati et al. (2002), for instance, compared five methods to determine the onset of the growing season in northern Nigeria. They determined the best method to be a "hybrid" method defined to minimize the risk of false starts without shortening the growing season beyond the necessary length for the crop varieties selected. For an ensemble of eight stations located between 11.5° and 13°N and using data collected between 1961 and 1991 depending on the stations, this optimal method provides 22 June as the mean date of the growing season, that is very close to our mean date for the monsoon onset (24 June). So it is necessary to document this onset and to investigate some more of the possibly involved mechanisms.

a. A meridional view

We have seen that it is possible to consider at a first order the West African monsoon as a zonally symmetric

atmospheric system. So in this section, to complement the results shown on rainfall and OLR (Figs. 4 and 6), we describe some of the main features of the monsoon seasonal cycle and its onset by using latitude cross sections averaged over the longitude band 10°W–10°E. Figure 7 depicts time–latitude diagrams for the relative vorticity at 925 hPa (Fig. 7a), the vertical velocity at 850 hPa (Fig. 7b), the vertical velocity averaged on 700–500 hPa (Fig. 7c), the meridional wind component at 600 hPa (Fig. 7d) and at 925 hPa (Fig. 7e), the zonal wind component at 600 hPa (Fig. 7f) and at 925 hPa (Fig. 7g), the vertical shear of the zonal wind between 925 and 600 hPa (Fig. 7h), the top of the monsoon layer (Fig. 7i), and the precipitable water height (Fig. 7j) (all of these diagrams are expressed in the referential $t_0 - 90/t_0 + 140$, where t_0 represents in average the 24 June for the monsoon onset).

The 925-hPa relative vorticity diagram (Fig. 7a) enables us to follow the meridional excursion of the ITF and of the heat low, expressed by positive maxima. From $t_0 - 90$ to $t_0 - 10$, the ITF moves gradually to the north, passing at 15°N around $t_0 - 40$, that is at the time of the monsoon preonset as we previously saw ($t_0 - 40$ is about the mean date defined for the preonset— 14 May). From $t_0 - 10$, the northward movement of the ITF accelerates, passing at 17.5°N at t_0 and going northward until about $t_0 + 20$ when it reaches about 21°N and stays during most of the summer with its greatest values before retreating southward. South of 10°N, relative vorticity is negative with minima located between the equator and 3°N, which signs the anticyclonic curvature of the winds crossing the equator. Tomas and Webster (1997) showed that negative absolute vorticity during northern summer just north of the equator over the Guinea Gulf is a sign of inertial instability of the low-level atmospheric circulation. Both the highest negative values of relative and absolute vorticity occur at t_0, suggesting that the summer monsoon onset is characterized by a maximum of inertial instability over the Guinea gulf, which is released when the monsoon begins to intensify.

Figure 7b depicts similar time–latitude diagram for the 850-hPa vertical velocity. It again enables us to follow the meridional excursion of the ITF and the associated heat low, represented by the axis of dry convection maximum in the low levels. Its seasonal migration is very similar, upward motion being associated with cyclonic vorticity. The greatest values are present around $t_0 - 80$, that is approximately at the beginning of April, then decrease regularly as the ITF moves to the north. However a temporary enhancement of dry convection occurs between $t_0 - 10$ and t_0 (it is the greatest at t_0) when the ITF is located at 17.5°N and when deep convection in the ITCZ, still located at 5°N, decreases (Figs. 4 and 6). Figure 7c (vertical velocity averaged on 700–500 hPa) and Fig. 7d (meridional wind component at 600 hPa) describe the upper part of the transverse circulation associated to the heat low. South

of the maximum of the upward 850-hPa vertical velocity, which is located at 17.5°N at t_0, both the 600-hPa northerly wind and 700–500-hPa downward velocity have the greatest values around 12.5°N at t_0, following a seasonal cycle similar to the one of 850-hPa vertical velocity. In the lower layers, the 925-hPa meridional wind (Fig. 7e) also has its greatest values at t_0 between 5° and 10°N. Then the whole meridional transverse circulation associated to the heat low has its greatest intensity at t_0. Figure 7c shows that the subsiding branch of the low-level circulation, located at 12.5°N at t_0, clearly separates the dry convection area in the north from the lower part of the deep convection associated to the ITCZ in the south.

The seasonal march of the ITF is also closely linked with the evolution of the African easterly jet (AEJ) described by the zonal wind component at 600 hPa, the pressure level of the jet core (Fig. 7f). This jet is located south of the ITF, between the dry convection area of the heat low and the deep convection of the ITCZ. The highest speed of the jet is observed close to t_0 at 10°N, concomitant with the enhancement of the low-level meridional circulation associated with the heat low and the northward acceleration of the relative vorticity maximum axis. This is consistent with the work of Thorncroft and Blackburn (1997) who showed that the AEJ is mostly controlled by the direct meridional circulation of the heat low, the northerly returning flow in the midlevels inducing an easterly acceleration due to the planetary vorticity advection. At 925 hPa (Fig. 7g), the westerly zonal wind is increasing in the latitude band 5°–15°N from about $t_0 - 30$, with a stronger acceleration at 10°N from $t_0 - 10$ (shown by the time section at this latitude), characterizing the monsoon enhancement. This leads to an increase of the vertical shear of the zonal wind between 925 and 600 hPa, which is the greatest at t_0 along 10°N (Fig. 7h).

This monsoon enhancement can be well characterized through the estimation of the top of the monsoon layer (Fig. 7i). This estimation has been defined following Lamb (1983) by the computation of the pressure levels where the zonal wind component equals zero, going from westerly downward to easterly upward. The areas shaded in Fig. 7i show the lowest pressure values, that is, the greatest depth of the monsoon layer. We must not consider the shaded areas in the upper corners, which correspond to the northern part of the anticyclonic cell controlling westerly winds at these latitudes and at these periods of the year. From $t_0 - 90$ to $t_0 - 10$, the top of the monsoon layer is located at a pressure level around 870 hPa and moves northward from 5° to 10°N. Then, the monsoon depth grows drastically and extends both northward and southward, during summer reaching 840 hPa north of 15°N and 760 hPa at 5°N. The monsoon onset is then characterized by an abrupt and large meridional and vertical development of the westerly wind layer as the ITCZ moves from 5° to 10°N. This evolution can be seen also in the precipitable water height for

FIG. 7. (a) Composite time–lat diagrams of NCEP–NCAR daily relative vorticity (10^{-6} s^{-1}), averaged over 10°W–10°E, filtered to remove variability lower than 10 days, and averaged over the period 1968–90 by using the ITCZ shift date as the reference date for each year.

which the northward progression abruptly accelerates at t_0, bringing, for instance, the isoline 40 kg m^{-2} previously stationary around 9°N since $t_0 - 60$, to 15°N at $t_0 + 30$ (Fig. 7j).

The development of the monsoon over West Africa is also seen in other features of the atmospheric circulation: 1) the setup of the tropical easterly jet (TEJ) at 200 hPa, which attains its greatest speed in August when convection and rainfall are the highest, leading to an enhancement of the anticyclonic circulation in the upper levels over West Africa (not shown); 2) the abrupt shift of the zonal winds at 600 hPa between 20° and 30°N (see Fig. 6f), which signs a similar meridional shift of the axis of the high geopotentials at this pressure level (see discussion on Fig. 14c further in the text).

A synthesis of these results is provided through the computation of the mean composite pressure–latitude cross sections of the 3D wind averaged on 10°W–10°E at $t_0 - 10$, t_0, and $t_0 + 20$ (Fig. 8). They depict the meridional and zonal atmospheric circulation associated to the West African monsoon system on the longitude band where land–sea contrast exists and where the ITCZ abrupt shift occurs. The three steps, $t_0 - 10$, t_0, and $t_0 + 20$, represent, respectively, the beginning of the temporary convection decrease, the convection minimum associated with the beginning of the ITCZ abrupt shift (summer monsoon onset), and the time when the summer monsoon regime first establishes.

Figure 8a depicts the meridional cross section of the atmospheric circulation at $t_0 - 10$ before the summer monsoon onset. We do not get a traditional view of the meridional Hadley-type circulations. They are in fact largely disturbed by the predominancy of the heat low transverse circulation between the surface and 500 hPa. This circulation is well developed with a dry convection area between 15° and 20°N, with one subsidence area around 25°N and another one in the midtroposphere between 10° and 15°N that merges with the northern subsiding branch of a quite reduced northern Hadley-type cell rejected in the upper part of the troposphere by the ascendance area of the heat low. The northern Hadley-type subsiding branch is in fact split into two different branches in the midtroposphere apparently due to the corresponding diffluent upper branches of the heat low. The subsiding area located between 10° and 15°N in the midtroposphere appears to interact with the deep convection of the ITCZ around 5°N, which constitutes the common upward branch of the two Hadley-type

cells, by advecting air masses downward and then weakening the upward vertical velocities in the midtroposphere. The heat low transverse circulation not only disturbs the northern Hadley-type circulation but also distorts the meridional circulation of the southern Hadley-type cell by controlling the northern extension of the low-level southerly monsoon winds up to 17.5°N. South of the equator, the latitudinaly extended subsiding area in the whole troposphere signs the southern branch of the southern Hadley-type circulation. The heat low circulation also controls the zonal circulation of the lower half of the troposphere (see isolines in Fig. 8). The dry convection at the latitude of the ITF is associated with the cyclonic circulation of the low pressure center in the lower levels, controlling the westerly component of the monsoon winds, and with the anticyclonic circulation around 600 hPa at the top of the heat low controlling the speed of the AEJ. So the heat low appears to be a key component of the monsoon system over West Africa before its onset.

The step t_0 is the most important one because it signs the summer monsoon onset as the beginning of a drastic amplification of the monsoon system that will be settled about 20 days later. One of the most striking points at step t_0 is the occurrence of a minimum in the convective activity over West Africa, just before the abrupt northward shift of the ITCZ (Fig. 8b). Between $t_0 - 10$ and t_0, the monsoon system has not moved northward but some part of it has intensified: the transverse circulation of the heat low is the strongest at t_0, associated both with an increase of the dry convection at 17.5°N and of the subsiding branch between 10° and 15°N. This intensification leads also to a stronger zonal circulation with higher westerly wind speed in the monsoon layer and a bit higher AEJ speed in the midlayer, resulting to a maximum of the vertical zonal wind shear between 925 and 600 hPa (Fig. 7h). We also observe a slight weakening of the deep convection in the ITCZ still located at 5°N, which is consistent with the minimum of rainfall values and maximum of OLR values. The northern Hadley-type cell is still reduced with its northern subsiding branch split into two parts because of its location above the heat low. The heat low enhancement at t_0 may be an explanation for the temporary convection minimum observed over West Africa at this step: the associated transverse circulation could stimulate intrusion of dry air from the northern mid- and upper layers of the troposphere into the moist deep convection lead-

FIG. 8. Composite pressure–lat cross sections of the NCEP–NCAR 3D atmospheric circulation averaged over 10°W–10°E, filtered to remove variability lower than 10 days, and averaged over the period 1968–90 by using as the reference date the ITCZ shift date for each year: (a) at $t_0 - 10$; (b) at t_0; (c) at $t_0 + 20$. The meridional-vertical circulation is represented by streamlines, shaded areas depict vertical velocity (unit in 100 Pa s^{-1}; negative values for upward motion), contours indicate zonal wind velocity (m s^{-1}; negative values for easterly wind).

ing to some convective inhibition in the ITCZ; as at the same time this transverse circulation helps to enhance westerly moisture advection in the monsoon layer and to increase the vertical zonal wind shear, both factors being favorable to convection, the convective inhibition could not last a long time, which could contribute to the abrupt summer monsoon onset.

Once the summer monsoon onset occurs, the atmospheric system shows large modifications. We previously saw that t_0 corresponds to the release of the inertial instability, and to a huge increase of the monsoon layer

and of the precipitable water height as the ITF goes on rapidly to the north and the ITCZ "jumps" to 10°N. Figure 8c shows the state of the atmospheric system at $t_0 + 20$ when the ITCZ has arrived at 10°N. We clearly see that the whole system has moved northward, the heat low as well as the ITF being located at 20°N, the ITCZ at 10°N, the AEJ 5° northward, and the monsoon zonal winds having increased. The intensity of the heat low associated transverse circulation is weaker than at t_0, as well as the AEJ wind speed. Another striking point is the large development of the northern Hadley-type

cell since its northern subsiding branch is not more perturbed by the heat low circulation because of its more northern latitudinal location, enabling its extension in the whole troposphere depth and leading it to merge with the northern subsiding branch of the heat low. This disorganizes the northern part of the heat low transverse circulation, by the reversal of the meridional winds from southerly to northerly in the whole troposphere and especially in the midlevels. This development of the northern Hadley-type cell can be explained by the intensification of convection in the ITCZ and its northern progression, as well as we observe an increase of the TEJ speed by more than 2 m s^{-1}. So once the summer monsoon is settled, the associated atmospheric circulation seems to be more controlled by the deep convection and the Hadley-type meridional circulation than before. However the heat low appears to still be able to modulate the atmospheric circulation in the low and midlevels. It may have some impact on the moist air advection from the southwest but also on intrusion of dry air coming from high levels of the northern midlatitudes, modulating convection and rainfall at intraseasonal scales (Sultan et al. 2003).

b. A zonal view

We said that we can consider at the first order the West African monsoon as a zonally symmetric atmospheric system. It enables us to highlight the heat low dynamics and suggests that it could have a significant impact on the summer monsoon onset specific to West Africa. On the other hand we have learned a few things about the heat low dynamics itself, and, in particular, we cannot explain why it enhances at t_0. Moreover, it seems that the onset is also characterized by an enhancement of the zonal wind component of the monsoon while the meridional component decreases from t_0 south of 10°N (Fig. 7e). So we further investigate the mechanisms of the monsoon onset by looking at it along a zonal axis.

In Fig. 4d the composite rainfall time series at 10°N can be considered as being constituted by a smooth seasonal cycle and superimposed short-term fluctuations clearly evident between $t_0 - 20$ and $t_0 + 20$. To better analyze these short-term fluctuations, which belong to the intraseasonal timescale, and to separate them for a smoother seasonal signal, we have investigated the composite fields of the atmospheric signals filtered between 10 and 60 days around t_0. Figure 9 shows the time sequence of the mean composite filtered 925-hPa wind field from $t_0 - 12$ to $t_0 + 21$, and the superimposed composite unfiltered rainfall field. The following comments take into account the superimposition of these filtered wind fields on the mean wind field. At $t_0 - 9$, two weak circulation centers are evident along 20°N in the filtered wind field, an anticyclonic one centered at 10°W and a cyclonic one at 12.5°E. These centers induce northerly wind anomalies over West Africa between

FIG. 9. Composite time sequence of the 10–60-day filtered NCEP–NCAR 925-hPa wind field averaged over the period 1968–90 by using as the reference date the ITCZ shift date for each year. The composite unfiltered rainfall field is superimposed (colored; mm day^{-1}). The scale of the vectors is indicated below (m s^{-1}).

10°W and 10°E while rainfall has begun to decrease since $t_0 - 10$ (Fig. 4d). These two centers develop and move westward in the following days. At t_0, the anticyclonic center has moved over the northern tropical Atlantic basin, and the cyclonic center is now located at 0° and 17.5°N. It is associated to the greatest dry convection in the heat low located at this latitude (Fig. 8) and the weakest rainfall values between 10°W and 10°E (Fig. 4d), whereas this center now induces southwesterly wind anomalies over a part of West Africa. At $t_0 + 3$, this cyclonic circulation extends greatly westward as well as the area of southwesterly wind anomalies. However, it is the time where rainfall over the central part of West Africa is the weakest, except west of 5°W and east of 7.5°E where rainfall begins to enhance, the first sign of the northern extension of the ITCZ. This is consistent with our interpretation of Fig. 8 suggesting that the heat low dynamics can simultaneously increase the convective inhibition by dry subsident air intrusion in the midlevels and increase the convective potential below by sustaining more humid air advection from the southwest. At $t_0 + 6$, the ITCZ begins its northward move by its northern boundary (isoline 4 mm day^{-1}) while the cyclonic center, now located at about 10°W, induces southwesterly wind anomalies over the large part of West Africa, moisture being imported not only from the Guinea gulf but also from the northern tropical Atlantic. At $t_0 + 9$, the heart of the ITCZ has clearly moved northward but has not reached 10°N yet. The moisture advection enhancement is now mostly controlled through westerly winds while the cyclonic center has moved to the north, being located

FIG. 10. Composite time–lon diagram of the 10–60-day filtered NCEP–NCAR 925-hPa relative vorticity (10^{-6} s^{-1}) at 17.5°N, averaged over the period 1968–90 by using the ITCZ shift date as the reference date for each year. Values are presented from t_0 (the shift date) minus 20 days to t_0 + 20 days. (b) Same as (a) but for the 850-hPa vertical velocity at 17.5°N (unit in 100 Pa s^{-1}). Negative values represent upward velocity. (c) Same as (b) but for 400-hPa vertical velocity at 10°N.

along the western coast of West Africa between 20° and 22.5°N. At t_0 + 12 and t_0 + 15, the westerly wind advection is maintained by a rather weak cyclonic circulation centered along 10°N while the northern cyclonic center is now visible over the ocean. The ITCZ get its final location at 10°N around t_0 + 21 with the appearance of rainfall minima along the Guinean coast. So the examination of the time series of the filtered wind field patterns enables to suggest that the summer monsoon onset is not only due to a meridional dynamics associated to the heat low but that it is also associated with a westward-propagating signal at an intraseasonal timescale, which induces the enhancement of the heat low rotational and transverse circulation.

Figure 10 helps to quantify these observations by depicting composite longitude–time diagrams of the 10–60-day filtered 925-hPa relative vorticity at 17.5°N (Fig. 10a), 850-hPa vertical velocity at 17.5°N (Fig. 10b), and 400-hPa vertical velocity at 10°N (Fig. 10c), computed between t_0 − 20 and t_0 + 20. The diagrams show between t_0 − 10 and t_0 + 10 the westward propagation of positive values of relative vorticity at 17.5°N, at an approximate mean phase speed of 4° longitude per day, concomitant with the enhancement of dry convection in the heat low at the same latitude and with a decrease of the 400-hPa upward velocity at 10°N. These signals are maintained mainly during the phase of decreasing then increasing rainfall at 10°N characterizing the specific summer monsoon onset over West Africa.

Figure 11a shows the composite unfiltered NCEP–NCAR 925-hPa wind field at t_0 − 15. Colored areas depict relative vorticity values and the solid line is the zero isoline of the zonal wind component depicting the

ITF over the Sahel. The different positive maxima of vorticity along the ITF localize the semipermanent cyclonic vorticity centers. Four maxima are identified over land at 30°, 15°E, 0°, and 15°W. Fifteen days after, at t_0 (Fig. 11b), the ITF has moved a bit northward. Figure 11d shows the composite wind field of the difference between t_0 and t_0 − 15. Colored areas depict OLR differences and isolines relative vorticity differences. The northward ITF location at t_0 is depicted in Fig. 11d by positive relative vorticity anomalies between 15° and 25°N and southwesterly wind anomalies on the southern side. However, the positive anomalies of vorticity are greater east of 0°. Associated with this vorticity field, negative OLR anomalies, meaning greater convection, are located mostly east of 5°E. They are located along 12.5°N, on the northern boundary of the ITCZ still located at 5°N. Small negative OLR anomalies also occur over the ocean north of the ITCZ. Positive OLR anomalies along 2.5°N sign the northward progression of the ITCZ southern boundary. Between approximately 10°W and 5°E, only positive OLR anomalies are observable, signaling the convection decrease at these longitudes just before the monsoon onset (Fig. 6).

Figure 11e shows the difference fields between t_0 − 10 and t_0 + 10. The OLR pattern is similar to the previous one but amplified. However, the relative vorticity field is different. The greatest anomalies are now located on the western part of northern Africa, and especially west of 5°E. This leads to developed southwesterly anomaly winds over West Africa and over the northern tropical Atlantic. At the same time, we can observe an amplification of an anticyclonic circulation along 30°N north of the Atlas Mountains (see Fig. 12), and of as-

FIG. 11. (a) Composite unfiltered NCEP–NCAR 925-hPa wind field at $t_0 - 15$ averaged over the period 1968–90 by using as the reference date the ITCZ shift date for each year. Colored areas depict relative vorticity values (10^{-6} s^{-1}). The solid line is the zero isoline of the zonal wind component. (b) Same as (a) but for t_0. (c) Same as (a) but for $t_0 + 20$. (d) Composite unfiltered NCEP–NCAR 925-hPa wind field [vector scale (m s^{-1}) is displayed below] for the difference between t_0 and $t_0 - 15$, averaged over the period 1968–90 by using the ITCZ shift date as the reference date for each year. Shaded areas depict OLR differences (W m^{-2}). Isolines represent relative vorticity differences (10^{-6} s^{-1}). (e) Same as (d) but for the difference between $t_0 + 10$ and $t_0 - 10$. (f) Same as (d) but for the difference between $t_0 + 20$ and t_0.

sociated northeasterly winds. Figure 11c shows the wind field pattern at $t_0 + 20$, and Fig. 11f the difference fields between $t_0 + 20$ and t_0. The positive vorticity anomalies are the greatest west of 5°E (Fig. 11f) and signal the most northward movement and amplification of the ITF at these longitudes. This induces southwesterly anomaly winds over the northern tropical Atlantic and over the large part of West Africa north of 10°N. Negative vorticity differences are still present north of the Atlas Mountains. The negative OLR differences are now covering all of West Africa signaling, in particular, the abrupt displacement of the ITCZ from 5° to 10°N on the longitudes 10°W–5°E. So, between $t_0 - 15$ and t_0

$+ 20$ the cyclonic activity of the ITF amplifies, first east of 5°E over land, and then west of 5°E over land and over the ocean, concomitant with the abrupt shift of the ITCZ around the Greenwich meridian area. The heat low circulation over land may also have an impact on the atmospheric circulation over the northern tropical Atlantic and on convection in the oceanic ITCZ.

5. A possible mechanism for the monsoon onset

In the previous section we have suggested that the abrupt shift of the ITCZ at the time of the summer monsoon onset could be due to an interaction between

FIG. 12. Orography of northern Africa (m; areas higher than 400 m are shaded by step of 400 m).

the deep convection and the meridional transverse circulation associated with the heat low. When the heat low significantly increases in an area located west of 5°E and south of the Atlas Mountains, this enhancement is accompanied by a similar increase of the anticyclonic vorticity north of the Atlas Mountains. We now discuss the hypothesis that the orography in North Africa could be responsible for this acceleration of the seasonal cycle of the monsoon over West Africa.

This hypothesis has been deduced from a work of Semazzi and Sun (1997). Although most of West Africa south of the Sahara is generally free of steep terrain, nearby mountains ranges, such as the Atlas Mountains, the Ahaggar Plateau, and the Tibesti highs, could exert significant control of the atmospheric circulation, and then of the convection in the ITCZ. Figure 12 shows that this orography is oriented along a northwest–southeast axis, so being nearer to the ITF trough in the east than in the west. In a previous work, Viltard et al. (1990) showed that the ITF over West Africa during northern summer is characterized by three semipermanent troughs (see our Fig. 11c for instance) and that some of them could be induced by orography. Semazzi and Sun (1997) examined the impact of orography on the West African summer climate in the context of different sea surface temperature anomalies through GCM simulations with and without orography. They found that these orographic effects are composed of (i) a quasi-stationary orographic ridge-trough dipole generated by the passage of the low-level prevailing easterlies over the Atlas–Ahaggar Mountain complex over northern Africa, and (ii) an elongated zonal windward orographic ridge generated as the cross-equatorial summer monsoons from the South Atlantic Ocean basin ascends over

the elevated landmass of West Africa. This results in wetter conditions over the Sahel and drier conditions farther south along the Guinea coast region, which would have been the case in the absence of orography, leading to a permanent orographic-induced rainfall dipole pattern over West Africa that we can associate to a more northward location of the ITCZ.

We can think about applying this hypothesis in the context of the seasonal cycle of the ITCZ. In the experiment of Semazzi and Sun, the orography-induced forcing leads to a northward location of the ITCZ during the summer due to the monsoon circulation enhancement because of the leeward effect. In our investigation, the onset is also characterized by a northward shift of the ITCZ in combination with the enhancement of the heat low circulation. This is the main reason why Semazzi and Sun's results and ours can be associated. If at a certain time during the northward course of the ITCZ, the interaction between orography and the low-level atmospheric circulation amplifies, the orographic-induced more northward location of the ITCZ could explain its abrupt shift concomitant with the onset of the summer monsoon. Figure 13 is a reproduction of Semazzi and Sun's Figs. 7 and 8, that is the difference between a control run with orography and a control run without orography of the July–September geopotential height and wind field patterns at 850 hPa. We see a distinct windward high pressure and leeward low pressure distribution across the Atlas–Ahaggar Mountain range. The low-level cyclonic circulation associated with this trough is the largest over the western part of Africa but extends far eastward over the land and contributes to enhance westerly advection of moisture from the northern tropical Atlantic to the Sahel. We can ob-

(a) Geopotential height at 850hPa

(b) Horizontal wind at 850hPa

FIG. 13. (a) Difference between a control run with orography and a control run without orography of the 850-hPa geopotential height (contour interval 2 m) and (b) 850-hPa streamline wind patterns in the experiment of Semazzi and Sun (1997).

(a) Horizontal wind and Geopotential height at 925hPa

(b) Geopotential height gradients and vertical velocity

(c) Geopotential height at 600hPa

Time (t0−90d. − − −> t0+140d.)

FIG. 14. (a) Composite difference between $t_0 + 12$ and $t_0 - 5$ of the 925-hPa geopotential height (mgp) and streamline wind patterns averaged over the period 1968–90 by using the ITCZ shift date as the reference date for each year. The location of the three centers used to compute time series are displayed. The dashed line represents the crest line of the orography. (b) Composite time series of the 925-hPa geopotential height difference between values at 30°N, 7.5°E (point 2) and 22.5°N, 10°W (point 1; thick curve; mgp), of the 925-hPa geopotential height difference between values at 30°N, 7.5°E (point 2) and 17.5°N, 15°E (point 3; thin curve; mgp), of the averaged 10°W–10°E 600-hPa vertical velocity at 27.5°N (dashed curve; unit in 100 Pa s^{-1}). Values are presented from $t_0 - 90$ days to $t_0 + 140$ days, and filtered to remove variability lower than 10 days. The two scales in the left ordinate correspond, respectively, from right to left to the thick and thin curves; the scale of the right ordinate corresponds to the dashed curve. (c) Composite time–latitude diagrams of the NCEP–NCAR daily 600-hPa geopotential height (mgp), averaged over 10°W–10°E, filtered to remove variability lower than 10 days, and averaged over the period 1968–90 by using the ITCZ shift date as the reference date for each year. Values are presented from t_0 minus 90 days to $t_0 + 140$ days.

serve also that these patterns are oriented as the orography on a northwest–southeast axis and that they extend farther to the west over the northern tropical Atlantic, inducing a rainfall dipole over this part of the ocean similar to the one over West Africa (Fig. 9a of Semazzi and Sun). So the orography of northern Africa can have also a significant impact of the oceanic ITCZ.

Figure 14a shows the composite difference between $t_0 + 12$ and $t_0 - 5$ of the 925-hPa geopotential height and streamline wind patterns. It shows a very high similarity with the results of Semazzi and Sun on the two sides of the Atlas–Ahaggar Mountain axis (similar fields have been obtained at 850 hPa, the level chosen by Semazzi and Sun; not shown). We can observe the positive geopotential height anomalies centered at 7.5°E on the windward side of the mountains (point labeled 2) and the two negative anomalies centered at 15°E and 10°W on the leeward side (labeled 3 and 1, respectively). These geopotential anomalies are associated respectively with anticyclonic circulation on the north leading to enhanced northerly winds from the Mediterranean Sea, and with cyclonic circulation on the south inducing enhanced monsoon winds over land and also over the

northern tropical Atlantic basin east of 30°W (see Fig. 11f). The orography-induced low-level circulation over West Africa could then favor the enhancement of the heat low both east and west of the Greenwich meridian and play a role in the ITCZ abrupt shift, as well as in the northward progression of the oceanic ITCZ.

Figure 14b shows the composite time series from $t_0 - 90$ to $t_0 + 140$ of the 925-hPa geopotential height difference between values at 30°N, 7.5°E (point labeled 2 in Fig. 14a) and 22.5°N, 10°W (point 1; thick curve), of the 925-hPa geopotential height difference between values at 30°N, 7.5°E (point 2) and 17.5°N, 15°E (point 3; thin curve), and of the averaged 10°W–10°E 600-hPa vertical velocity at 27.5°N (dashed curve). The thick and the thin curves again highlight the abrupt change of the atmospheric circulation with their steep slope occurring from $t_0 - 10$ to $t_0 + 20$, following a rather stable state before $t_0 - 10$. The time series of the geopotential heights associated to the anticyclonic center and to the cyclonic centers separately show an abrupt break around $t_0 - 10$ (not shown). The dashed curve in Fig. 14b shows the downward vertical velocity at 27.5°N, that is the location of both the northern subsiding branch of the heat low and the northern Hadley-type cell (Fig. 8). The subsidence in this area begins to increase around $t_0 - 35$, then more rapidly from $t_0 - 20$ to $t_0 + 40$. This enhancement is due first to the northern extension of the subsiding part of the transverse cell associated to the heat low and to its intensification around t_0, second to the large development of the northern Hadley-type cell after t_0, in particular of its northern subsiding part. So the windward subsidence and high geopotentials begin to increase first, increasing then the associated northeasterly winds, leading finally to the leeward cyclonic circulation enhancement.

Figure 14c shows the latitude–time cross section of the 600-hPa geopotential heights averaged over 10°W–10°E. Figure 7f has highlighted the meridional abrupt shift of the zonal winds at 600 hPa between 20° and 30°N signing a similar shift of the axis of the high geopotentials at this pressure level. We see in Fig. 14c that it is concomitant with a strong increase of the geopotential heights to the highest values reached during the summer. This anticyclonic center is located after the monsoon onset above and northward of the North Africa highlands, and it is associated with the enhancement of the subsiding motions of the northern Hadley-type circulation. This signal of the monsoon onset can be put in parallel with the Asian monsoon onset where a similar northward abrupt shift of the subtropical westerly jet and the associated subtropical high pressures have been identified in the high troposphere above the Tibet highlands (Hahn and Manabe 1975).

Figure 14a shows a background signal specific to the West African summer monsoon onset over the period 1968–90. We have investigated the interannual variability associated with this pattern by performing an automatic classification (Janicot 1992) of these geopotential fields (not shown). We obtained four classes with similar spatial patterns, retaining the three centers (labeled 1, 2, and 3 in Fig. 14a) at about the same locations, with similar geopotential height gradient signs, but with different height amplitudes associated with these centers. However we could not define any

significant characterization of the ITCZ shifts related to these different classes (the four averaged rainfall fields depict a similar abrupt shift), nor any correlation with the summer rainfall amount over the Sahel. This suggests 1) the independency between the mechanisms associated to the interannual rainfall variability of the West African monsoon and the mechanisms induced in the monsoon onset, and 2) the hypothesis that the orography-induced mechanisms shown by Semazzi and Sun on the mean strength of the West African monsoon can work in the monsoon onset of individual years.

We suggest that the northern progression of the heat low and its transverse circulation leads to the enhancement of the anticyclonic circulation on the windward side of the Atlas–Ahaggar Mountain crest, inducing enhanced northeasterly winds that interact with the orography and lead to the increase of the leeward trough and associated westerly advection of moisture inland, contributing to the enhancement of the heat low intensity at t_0. Then the ITCZ moves to the north and intensifies after breaking the convective inhibition induced by the heat low. This leads to the enhancement of the northern Hadley-type cell and to the development of its northern subsiding branch, which also contributes to increase the high geopotentials north of the mountains and then the orography-induced lee-trough southward. The abrupt shift of the ITCZ could then be explained by a positive feedback between the atmospheric circulation of the heat low, the convection in the ITCZ, and the Hadley-type circulation, through the control by the orography. The apparent westward signal seen in the filtered wind fields could be due to the northwest–southeast orientation of the orography axis, which induces the interaction with the heat low first in the east where the mountains are nearer to the ITF than in the west, and second in the west. The fact that we observe an abrupt shift of the ITCZ only in the western part of West Africa might result from the enhancement of moisture advection that comes from the west and has a stronger impact west of the Greenwich meridian.

6. Synthesis and conclusions

As the seasonal cycle advances in the first half of the year, a land–sea temperature gradient grows between West Africa and the equatorial and southern Atlantic basin, and the resulting low-level meridional pressure gradient moves northward. Over West Africa, the high northward temperature gradient and the high southward moisture gradient between the Sahara and the Guinean coast induce a meridional lag between the surface dry static energy maximum (located more northward) and the moist static energy (located more southward). This shows that the deep convection in the ITCZ controlling the two Hadley-type cells is located to the south of the dry convection of the heat low, limited in the low and midlevels of the troposphere, and capped by the subsiding branch of the northern Hadley-type cell. The

ground location of the heat low is also characterized by a positive relative vorticity maximum and by the confluence of the moist southwesterly winds of the monsoon and the dry northeasterly Harmattan, that is the ITF. At the preonset of the summer monsoon, that is at approximately $t_0 - 40$ days (where t_0 means the date of the summer monsoon onset), the ITF is located at 15°N and the ITCZ at 5°N. Between $t_0 - 40$ and $t_0 - 10$, the heat low dynamics dominates north of the ITCZ and a part of its northern subsiding area is located above and north of the Atlas–Ahaggar crest. During that time, isolated convective systems begins to occur south of the ITF, contributing to the northern progression of the ITCZ northern boundary, and may be to a slight increase of the northern Hadley-type cell dynamics and its subsiding motions, reinforcing the subsidence over the Atlas–Ahaggar orography. These two factors contribute to enhance the high geopotentials north of these mountains and the associated northeasterly winds.

Between $t_0 - 10$ and t_0, the interaction between orography and northeasterly winds on the windward side of these mountains enhances and lead to the development of a leeward trough that reinforces the heat low dynamics. This has opposite consequences. On one hand, the midlevel subsidence of the southern branch of the heat low transverse circulation increases and may be accompanied by intrusion of dry air from the north. This can increase convective inhibition in the ITCZ and explain the observed temporary convection and rainfall decrease. On the other hand, cyclonic vorticity increases at low levels contributing to a greater inland moisture advection by stronger southwesterly winds and a deeper monsoon layer. In the midlevels, the anticyclonic circulation increases leading to the enhancement of the AEJ. Then the vertical wind shear between the low and midlevels becomes higher. Deeper monsoon and higher vertical wind shear enhance the local potential instability. The observed weaker convection also leads to higher downward solar radiation received at the surface (seen in the NCEP reanalysis; not shown), a favorable condition for potential instability increase. Finally as the heat low transverse circulation also maintains the instability of the AEJ by imposing a reversal of the meridional gradient of potential vorticity (Thorncroft and Blackburn 1997), the enhanced heat low dynamics increases the instability character of the AEJ (negative meridional gradient of potential vorticity is enhanced at 10°N between $t_0 - 10$ and t_0; not shown), which may favor easterly wave development and then convection.

From t_0 to $t_0 + 20$, the accumulated potential instability breaks the convective inhibition, the inertial instability of the monsoon circulation is released and the associated regional scale circulation increases, leading to the abrupt shift of the ITCZ. Then the ITCZ moves to the north, where thermodynamical conditions are favorable, up to 10°N. The northern Hadley-type circulation drastically enhances, which leads to the development of subsidence in the whole troposphere between

25° and 30°N. Then geopotential highs increases north and above the Atlas–Ahaggar Mountains, which enhances the orography-induced windward high and leeward trough again, maintaining an active convective ITCZ during the summer monsoon. The abrupt shift of the ITCZ could then be explained by a positive feedback between the atmospheric circulation of the heat low, the convection in the ITCZ, and the northern Hadley-type circulation, through an interaction with the orography.

The northwest–southeast orientation of the Atlas–Ahaggar crest can induce the interaction with the heat low first in the east where the mountains are nearer to the ITF than in the west, and second in the west. The fact that we observe an abrupt shift of the ITCZ only in the western part of West Africa may result from the enhancement of moisture advection that comes from the west and has a stronger impact west of the Greenwich meridian.

Another consequence of the orography-induced interaction with the atmospheric circulation is that the induced leeward through, increasing the cyclonic vorticity in the heat low, may stimulate moisture convergence in the oceanic ITCZ near the western coast of West Africa. However, a similar ITCZ abrupt shift is not observed over this part of the northern tropical Atlantic.

Finally, once the monsoon onset occurs, the developed northern Hadley-type circulation can favor intrusion of dry air from the north and the upper levels into the ITCZ. These intrusions can be also controlled by the northerly winds of the heat low transverse circulation. This question has been addressed in Sultan et al. (2003) through the study of the intraseasonal timescale modulation of convection observed during the West African summer monsoon. This mechanism should not work efficiently before the summer monsoon onset since at this time the northern Hadley-type cell is not developed enough to the north to allow northerly advections far from the ITCZ.

The hypothesis of the abrupt shift of the ITCZ has been deduced from the observations. It must be confirmed through other ways of investigation. In particular atmospheric GCM experiments similar to that of Semazzi and Sun, but in the context of the seasonal cycle, would be a good way to test this hypothesis. However, these types of models have to first accurately reproduce the dynamics of the West African monsoon described here to be validated (Vernekar and Ji 1999). On the other hand, the hypothesis presented here is not necessarily the only explanation for the ITCZ abrupt shift. For instance, interactions between land surface processes and deep convection, especially through the impact of soil moisture and of the sensible heat input in the boundary layer, have been previously involved to explain northward-propagating rain bands in the Indian monsoon (Webster 1983; Gueremy 1990; Ferranti et al. 1999). Similar mechanisms could be involved in the ITCZ latitudinal shift over West Africa. Xie and Saiki (1999)

have proposed another mechanism to explain delayed abrupt onsets observed in the Asian summer monsoons by the concept of "geostrophic monsoon": in a rotating atmosphere, the land–sea temperature difference does not necessarily lead to a direct overturning cell, but instead can be balanced by adjusting its vertical shear so as to be in a thermal wind balance with the meridional temperature gradient; so the onset is delayed and can be initiated by the explosive growth of a traveling disturbance. Plumb and Hou (1992) also showed that such a thermal wind adjustment is possible for an off-equatorial heating, and that an intensive meridional circulation occurs when the subtropical heating rate exceeds a certain threshold, another source for an abrupt change of the monsoon circulation and its onset. These types of mechanisms must be investigated further through ocean–atmosphere coupled GCMs since the monsoon onset could have a significant impact on the oceanic ITCZ near the western coast of West Africa.

In this work, by using gridded rainfall and NCEP–NCAR reanalysis data during the period 1968–90 (OLR was used since 1979), we have described the arrival of the summer monsoon over West Africa and characterized two steps: its preonset and its onset. We have shown that these steps are not correlated in time nor correlated with the summer rainfall amount over the Sahel. The results on the definition of these steps have been obtained through the computation of composite means based on the preonset or the onset date for each year. This method assumes that the rate of development before and after each of these steps is similar each year. We have characterized the preonset stage of the summer monsoon by the arrival of the ITF at 15°N around mid-May while the ITCZ is still centered at 5°N. The onset has been pointed out by an abrupt latitudinal shift of the ITCZ in late June from 5° to 10°N. This composite approach enables to highlight the atmospheric processes associated with these steps of the West African monsoon seasonal cycle. It could be thought that the composite procedure introduces artificial signals since, for the monsoon onset for instance, as we select t_0 so as to detect a shift, it is logical to find such a shift. However Le Barbé et al (2002) found a similar meridional shift of the ITCZ by computing a classic average of the rainfall seasonal cycle. So our composite approach, not only does not create the ITCZ shift, but enables us to focus more precisely on the associated dynamical mechanisms.

This is not only an interesting meteorological and climatological issue, it is also a major point for agriculturists. A better knowledge of the mechanisms of the arrival of the summer monsoon should help providing efficient forecast of the beginning of the growing season. This point has not been considered here since we have used a smoothing procedure to remove high-frequency transient signals from the composite rainfall and dynamical patterns associated to the summer monsoon onset. Forecasting this onset in an operational context need to take these transients into account, in addition to the composite scenario proposed here.

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