Climatology of coastal wind regimes in Benin

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Abstract

Using surface and lower-tropospheric wind observations at Cotonou (Benin), a comprehensive ten-year (2006–2015) climatology of coastal wind regimes at the West African Guinea Coast is presented for the first time. Four wind regimes are objectively identified: the monsoon and Harmattan winds, the nocturnal low-level jet (NLLJ) and the land-sea breeze (LSB) system. Amongst other things, these regimes are important for the occurrence of coastal squall line systems and for the redistribution of pollutants emitted from coastal cities. The seasonal cycle is split into the long (December–February) and short (July–September) dry seasons, the long (May–June) and short rainy (October–November) seasons, and the beginning of the long rainy season (March–April). In terms of the LSB, the lack of a diurnal reversal in wind direction obscures sea breeze activity at the height of the southwesterly monsoon flow in boreal summer and during a period of stronger winds in March-April related to large land-sea thermal contrasts. During these periods, the sea breeze manifests itself in a relative increase of the monsoon background flow during daytime, while early morning land breezes are infrequent. The latter expectedly, peak during the long dry season in December with a secondary peak during the long rainy season in May–June. The large–scale monsoon and Harmattan flows, mutually exclusive by definition, occur respectively, 88.2–98.4 % and 1.6–11.8 % per year with Harmattan days restricted to the long dry season and deepest monsoon flows of about 2.5 km confined to the long rainy season. The NLLJ is most prominent during the little dry season with, however, a similar and hitherto less documented peak in occurrence frequency during the beginning of the long rainy season. At this time of the year, the African Easterly Jet passes northward over Cotonou, causing the largest 3000–600 m vertical wind shear.

Keywords: monsoon, Harmattan, nocturnal low-level jet, land-sea breeze

1 Introduction

The West African Guinea Coast stretches from Cape Palmas at the border between Liberia and Ivory Coast to Cape Lopez in Gabon. The seasonal cycle of its weather and climate is influenced by the West African Monsoon (WAM, Fink et al., 2017). The area is currently experiencing considerable economic growth associated with a major population increase and strong urbanization (Knippertz et al., 2015). Due to rain-fed subsistence farming and several large hydropower installations, rainfall variability has immediate socio-economic ramifications. Going back to 1886, Nicholson et al. (2018) analyze the longest rain gauge series for the Guinea Coast ever and conclude that the dry conditions prevailing since about 1968 are unprecedented within this period. On the other hand, the frequency of extreme rainfall events appears to have increased in the region (Sanogo et al., 2015) and flashflood events have recently been reported (Maranan et al., 2018). In this context, Maranan et al. (2018) have shown the importance of low-level wind shear for the occurrence of strong mesoscale convective systems along the Guinea

Coast, a relationship that has been demonstrated for the Sahel by Taylor et al. (2017). Moreover, Fink et al. (2010) note that due to land-sea breeze (LSB) related afternoon convective systems, rainfall after the long boreal winter dry season starts earlier in the hinterland than directly at the coast. Maranan et al. (2018) suggest that a mid-day peak in moderate to strong convective systems along the Guinea Coast is related to LSB convection. Apart from rainfall, wind regimes at the Guinea Coast also play an important role for the redistribution of pollutants emitted from megacities, as has recently been outlined by Knippertz et al. (2017) and Flamant et al. (2018a). Despite the importance of wind regimes for rainfall and air pollution at the Guinea Coast, very few studies have hitherto addressed this topic. Thus, the overall aim of the present study is to advance our understanding of the types and climatological characteristics of coastal wind systems.

Low-level monsoonal southwesterlies are the dominant large-scale, near year-round wind regime along the Guinea Coast. They are strongest and deepest during the peak monsoon period from July to September (JAS) with near-surface wind speeds on the order of $4 \, \mathrm{ms}^{-1}$. In this three-month period, the rainfall in the Sahel attains its seasonal peak, while large parts of the Guinea Coast experience the little dry season (e.g.

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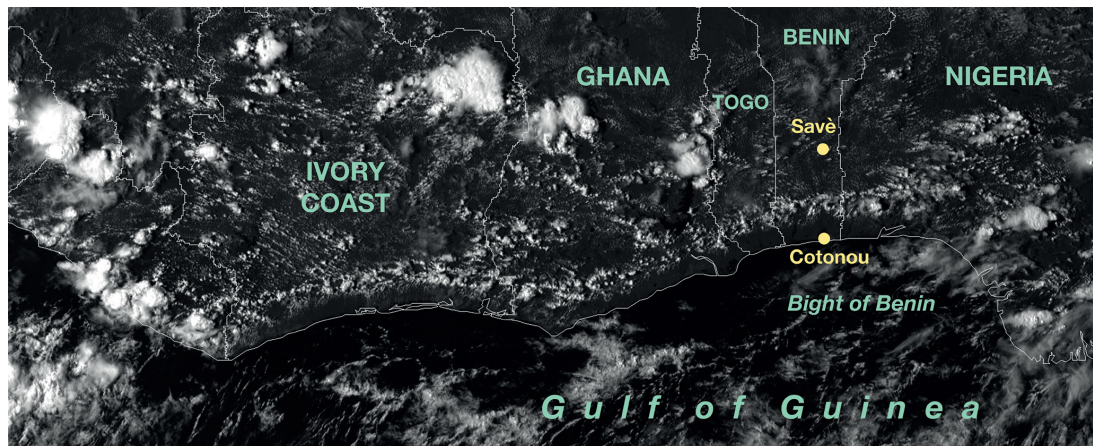
Figure 1: Visible image of the Guinea Coast from Meteosat Second Generation 2 for 18 September 2016 at 1500 UTC. Country names, ocean areas and names of surface stations used are indicated.

Fink et al., 2017). The top of the southwesterly monsoon flow, often also termed the monsoon layer, varies between 850 and 700 hPa, though deeper flows with a westerly wind component occur during certain synoptic weather situations (Knippertz et al., 2017; Flamant et al., 2018b). During this time of the year, the development of a southwesterly nocturnal low-level jet (NLLJ) is a widespread phenomenon and has been documented for coastal, Soudanian, and Sahelian stations (Lothon et al., 2008; Abdou et al., 2010; Schrage and Fink, 2012; Schuster et al., 2013; Kalthoff et al., 2018). In boreal winter, the Intertropical Discontinuity (ITD), the confluence zone between the moist and cool monsoonal southwesterlies and the dry Harmattan winds, is located just north of the Guinea Coast at 8° N (Fink et al., 2010). During periods from a few days up to two weeks, the dry and dusty Harmattan winds push southward across the Guinea Coast onto the Gulf of Guinea waters (Hamilton and Archbold, 1945; Fink et al., 2010). During these periods, coastal stations experience northerly to northeasterly 10 m winds throughout most of the day with unusually low humidities. From December to mid-February, 2 m temperatures are somewhat cooler, especially at night. However, Knippertz and Fink (2006) documented an early March Harmattan case that brought about a heat wave along the Guinea Coast. In Cotonou, this was partly related to the fact that the daytime sea breeze (SB) with cooler air from the ocean did not set in until 15 UTC (16 LT) allowing the 2 m temperature to reach 37 °C. While monsoon, Harmattan and NLLJ are usually larger-scale features, the fourth wind regime, the LSB circulation at the Guinea Coast is a local feature forced by the diurnal cycle of land-sea thermal contrasts. Despite its above-mentioned importance for coastal thunderstorms and atmospheric pollution, surprisingly few studies have investigated the climatology of the LSB regime in southern West Africa. Some studies used visible satellite imagery to document the daytime inland propagation of clear skies behind the SB front (Cautenet and Rosset, 1989; Schrage and Fink, 2012; van der Linden et al., 2015). The visi-

ble Meteosat image for 18 September 2016 at 15 UTC shows clear skies over the ocean and immediate coastal strip with a line of convective cloud likely demarcating the SB front convergence (Fig. 1). Due to the clearing of skies after the passage of the SB front in JAS, coastal stations show a much higher incoming solar radiation when compared to inland stations where irradiation is substantially reduced due to an extensive morning low-level cloud deck (Knippertz et al., 2011). This suggests an important role of the LSB for the distribution of potential solar power yield from photovoltaic installations. Bajamgnigni Gbambie and Steyn (2013) and Abayomi et al. (2007) used surface wind observations from Beninese and Nigerian stations, respectively, to infer climatologies of the LSB. The former authors found a peak in SB occurrence at Cotonou (Benin) in December and May with a minimum in August during the little dry season. However, while the coastal land-sea temperature contrast was large in December, it was smaller in May than in August. Land-sea breeze climatologies are of course dependent on their definition and Bajamgnigni Gbambie and Steyn (2013) imposed a wind direction reversal criterion. This might not be appropriate in cases of too strong background winds and/or to weak nighttime cooling (Azorin-Molina, 2007; Azorin-Molina et al., 2011), when the local LSB impact on the wind field is a morning increase and afternoon decrease of wind speed rather than a directional changes.

In this study, seasonal and diurnal climatologies of four coastal wind regimes, the monsoon and Harmattan winds, the NLLJ, and the LSB will be compiled for the station of Cotonou (Benin, 6° 21′ N, 2° 23′ E) for 2006–2015. A flexible definition will allow for the detection of LB and SB days without a wind reversal. In this context, attention will be paid to periods when a strong coastal land-sea temperature contrast (i.e. a driver of a classical LSB) or known diurnal variations in the large-scale monsoon systems drive diurnal variations in low-level coastal winds. In a first study of this kind, we take advantage of the rare availability of four-times daily, low-level wind profiles from a combination of PI-

Table 1: Percentage of missing values at Cotonou station from 10 m to 3000 m height during the period 2008–2015.

Height (m)	10	600	900	1500	2100	3000
Missing values (%)	3	31	34	42	52	62

LOT balloon and radiosonde data to extend previous climatologies from the surface to the lowest 3 km of the atmosphere. In their discussion of the LSB at the Guinea Coast, PARKER et al. (2017) note that the nighttime land breeze (LB) is shallower than the daytime SB. Thus we will study the vertical structure, but also discuss the frequency of days when certain regimes occur simultaneously.

The data and methods used in this study are described in Section 2. Section 3 presents the climatology of low-level wind regimes at Cotonou during the seasonal scale. The fourth section summarizes and concludes the study.

2 Data and Methods

2.1 Data set

The observational data set used in this work includes surface and upper-air data from Cotonou meteorological station. Surface data consist of 10-years (2006–2015) of 3-hourly parameters (10 m wind, 2 m temperature and relative humidity) and the daily rainfall amount, all provided in hard copy by the Beninese National Meteorological Agency. Hourly sea surface temperature (SST) data measured at the station of Cotonou Port by "Institut de Recherches Halieutiques et Océanologiques du Bénin" (IRHOB) are also used to analyze the LSB. Upper-air data include 8-years (2008–2015) of wind observations extracted from Cotonou PILOT and TEMP reports. These reports are provided by the OGIMET website (http://www.ogimet.com) that collects and processes freely available data, mainly from the National Oceanic and Atmospheric Administration (NOAA). Cotonou upper-air station provides four observations per day (00, 06, 12 and 18 UTC) that are all fed into the Global Telecommunication System as PILOT and TEMP reports. The latter started in September 2012 for the 12 UTC reporting time only. The reports include wind data for standard isobaric surfaces and for fixed additional levels, the altitudes of which are given in metric units. As this work focuses on low-level wind regimes, the standard isobaric surfaces used here are 850 hPa and 700 hPa, approximately 1500 m and 3000 m above sea level, respectively. In addition to the 10 m surface winds, the other levels with fixed metric heights selected from the reports are 600 m, 900 m and 2100 m, giving six levels in total. Regarding the quality of PILOT winds, it is known that optical tracking of PILOT balloons yield meaningful wind speeds though some levels for few soundings might be erroneous (e.g. RIDER and ARMENDARIZ, 1968). However, we consider the 8-year average PILOT balloon winds as accurate enough to study the details of the four wind regimes.

At the Guinea Coast, due to logistic issues, several upper-air stations do not usually provide four soundings per day. For example, PILOT reports in Lomé and Lagos stopped several years ago. Even at Cotonou station, because of issues related to the optical tracking used for the PILOT balloon soundings, bad weather conditions and cloud cover, missing values are numerous in the data set, especially at 2100 m and 3000 m heights. Wind measurements at Cotonou before 2008 are very infrequent on the OGIMET website and the frequency of observations decreases from the surface to 3000 m height (not shown). For the 2008–2015 period, missing data represent 3 % of the total number of observations for 10 m surface winds and can reach 62 % for winds at 3000 m height (Table 1). For this period, only 70 % of the PILOT and TEMP reports expected are in fact obtained.

2.2 Methods

Over the last few decades, different methods for identifying LSB episodes were tested and helped to construct databases for many parts of the world. In coastal regions of West Africa, however, few descriptions of LSB systems are found in the literature and no databases are available. Usually the first criterion to detect LSB episodes is the distinctive wind behaviour with the rapid shift in wind direction from onshore to offshore at SB onset. A review of criteria used for both automated and manual identifications of sea or lake breezes was recently given by AZORIN-MOLINA et al. (2011). The challenge is to ensure that the changes measured in meteorological variables are in fact produced by SB frontal passages (RYZNAR and TOUMA, 1981). Based on STEYN and FAULKNER (1986), BAJAMGNIGNI GBAMBIE and STEYN (2013) described the SB phenomenon at Cotonou using surface observational data over the period 2006–2010. Their first criterion, the most important one, was related to the reversal of wind direction. Because a land–sea temperature difference is necessary to drive the SB circulation (GUSTAVSSON et al., 1995), the second filter was based on the land–sea thermal gradient. The third criterion is related to the daily sunshine duration, as one expects that SB, a thermally driven local circulation, occurs only on days when the sun shines for a significant fraction of the day (STEYN and FAULKNER, 1986). For the Bay of Alicante AZORIN-MOLINA et al. (2011) utilize a steady increase in the 30 minute wind speed and maximum wind speed after sunrise until noon, followed by a gradual decrease from afternoon until sunset as a detection criterion. Nocturnal offshore winds are rare in summer months over the Bay of Alicante, probably because the cooling is not strong enough (AZORIN-MOLINA, 2007). Thus, SB passages are not associated with a reversal in wind direction, but rather with a typical in-
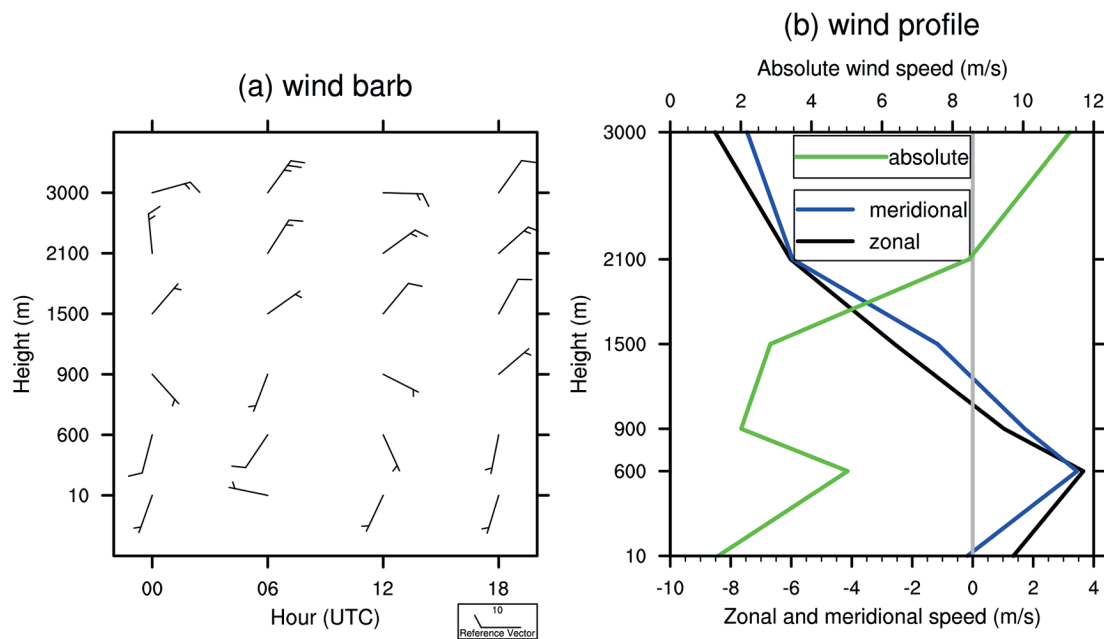
Figure 2: Example of LSB situation at Cotonou: (a) Diurnal cycle of low-level winds and (b) vertical profile of the zonal and meridional wind components at 06 UTC, both averaged for November 2015.

crease in wind speed as a function of positive temperature difference between land and sea (AZORIN-MOLINA et al., 2011).

Specifically for Cotonou, this paper will expand past work through the use of upper-air data and by placing the LSB into the context of other low-level wind regimes, i.e. the Harmattan winds, the NLLJ and the monsoon flow. We use a set of criteria to assign a given day to one (or sometimes several) wind regimes or to classify it as "unspecified". The definitions of the four wind regimes are given below.

2.2.1   LB and SB detection

In the coastal region of Benin, where the shoreline is almost zonal, slight northwesterly or westerly winds are the main characteristic of the LB system in the early morning, which is expected to be most frequent in boreal winter. This is evident from the average offshore wind at the surface at 06 UTC in Figure 2a that exemplarily shows the example of November 2015. The LB flow is rather shallow, as already at 600 m the meridional component has shifted to southerlies (blue line in Fig. 2b). In absence of higher resolution measurements it is difficult to delineate the return flow. The change in surface wind direction from westnorthwest at 06 UTC to southwest at 12 UTC (Fig. 2a) is therefore the most distinctive feature of the SB in Benin we can detect with our data.

In this study LB and SB occurrence at Cotonou station are detected separately using specific criteria. LB occurrence is detected by offshore 10 m winds by nighttime and in the early morning (i.e. 00–06 UTC). The following flexible filters, derived from AZORIN-MOLINA et al. (2011), are used to detect SB events without explicitly requesting a wind reversal:

- Filter 1: The 3-hourly wind speed at 10 m shows a steady increase from 06 UTC to 12 UTC and a gradual decrease from afternoon until 21 UTC. The threshold for the increase or decrease in speed is arbitrarily fixed here at $2\,\mathrm{ms}^{-1}$.
- Filter 2: The 3-hourly wind direction at 10 m ranges from 100° to 260°, i.e. onshore flow, after sunrise (09, 12 and 15 UTC).
- Filter 3: The daily average daytime 2 m air temperature between sunrise and sunset (06–18 UTC) over land is greater than the daily average sea surface temperature between sunrise and sunset.

In the results section we will discuss the differences to the study by BAJAMGNIGNI GBAMBIE and STEYN (2013), who require a reversal of wind direction and a minimum number of sunshine hours in addition.

2.2.2   Harmattan detection

At Cotonou station, the Harmattan is represented by low-level northeasterly winds as shown exemplarily in Figure 3a for December 2015. Except at 18 UTC, when the SB phenomenon occurs, and at 00 UTC at 600 m height, when the NLLJ starts to set in, northeasterly Harmattan winds prevail at lower levels. While the meridional component decreases from surface to 3000 m, the zonal component increases (Fig. 3b), indicating a gradual shift from the Harmattan winds to mid-level easterly winds.

To detect the infrequent Harmattan bursts at Cotonou, which mainly occur in winter months, the following two filters are employed, using surface observations:

- Filter 1: The 3-hourly winds are northerly (340° to 80° clockwise) between 09 and 15 UTC.
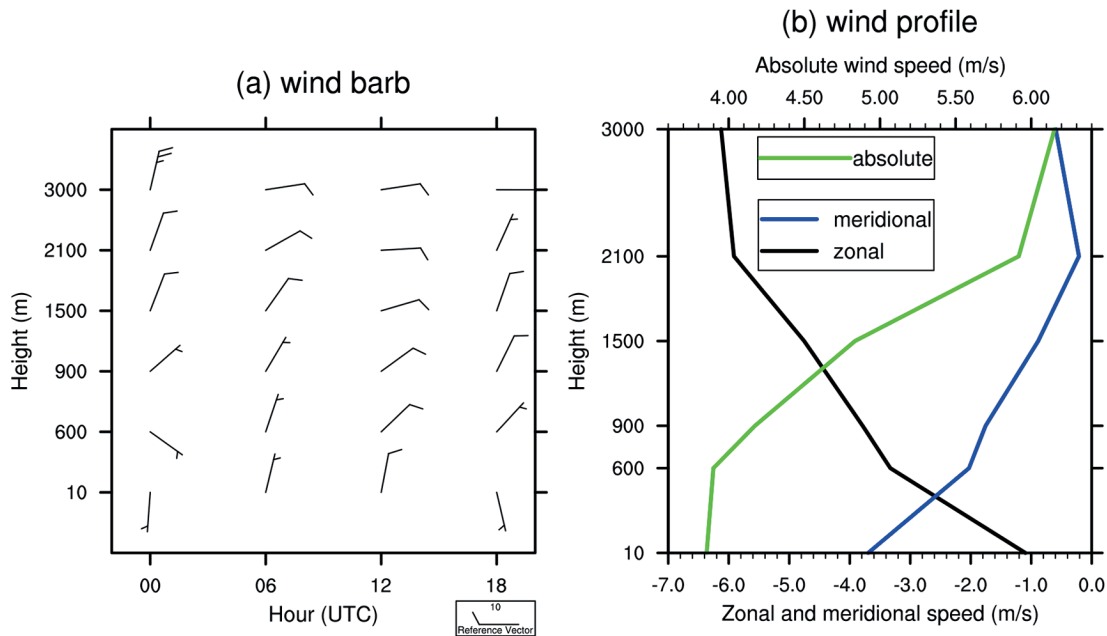
Figure 3: Example of Harmattan situation at Cotonou: (a) Diurnal cycle of low-level winds and (b) vertical profile of the zonal and meridional wind components at 12 UTC, both averaged for December 2015.

- Filter 2: The relative humidity reaches values of less than 60 % in the afternoon (between 12 UTC and 15 UTC).

2.2.3 NLLJ detection

The detection of the NLLJ is mainly based on the results of previous studies in West Africa. Many authors (e.g. Parker et al., 2005; Lothon et al., 2008) revealed that higher wind speeds occur over several locations in West Africa during the night and in the early morning than during other times. The observed wind speeds are often greater than $10\,\mathrm{ms}^{-1}$ and centered around $400\,\mathrm{m}$ height (Lothon et al., 2008) with high values also found at $600\,\mathrm{m}$ height in PILOT observations. During daytime, convective turbulent mixing is large and limits horizontal transport of air masses. In the late afternoon, the heat flux decreases, and the circulation responds to the surface pressure gradient force imposed by the heat low without being slowed down by friction. The winds are accelerated towards the low pressure during the night, reaching their maximum speed in the morning hours. Due to the proximity to the equator, the influence of the Coriolis force is small. Exemplarily Figures 4a and b show the average speed of the zonal and the meridional components of the NLLJ at Cotonou in September 2015 at 06 UTC. The nocturnal peak observed at $600\,\mathrm{m}$ height in PILOT soundings is clearly visible. The average wind is much stronger during nighttime and can reach up to $10\,\mathrm{ms}^{-1}$ on individual days. To detect the NLLJ, upper-air winds at $600\,\mathrm{m}$ height are used, which is mainly southwesterly at Cotonou station. The threshold is fixed to $10\,\mathrm{ms}^{-1}$ and days with wind speeds (at 00 UTC or 06 UTC) greater than this threshold are classified as NLLJ days.

2.2.4 Monsoon flow detection

The monsoon flow, characterized by southwesterly winds from the ocean, is observed in coastal regions nearly the whole year except for the short period when the northerly Harmattan winds reach the coast. So these two regimes are mutually exclusive. It has been the subject of several studies during the last 10 years, mainly through the African Monsoon Multidisciplinary Analysis (AMMA, see Redelsperger et al., 2006) project.

The depth of the monsoon layer is closely linked to the vertical structure of the ITD, the sloped boundary between the southerly and northerly winds over West Africa, and its seasonal migration. The monsoon season in West Africa starts with the propagation of the ITD from the Guinea Coast towards the north during the months of March and April. This defines the beginning of the first rainy season south of 10° N (Sultan and Janicot, 2003), when the vertical extent of the monsoon flow commences to increase. At this time, the average monsoon layer depth, identified by West African operational meteorologists as the height of zero meridional wind, is generally less than 1500 m. Figure 5 shows an example for April 2015 clearly demonstrating the marked shift from southerlies in the lowest three levels to easterlies at 2100 and 3000 m at all times of day.

Statistics on the monsoon depth at Cotonou station are presented here using available winds from PILOT sounding observations. A given day is either assigned to a shallow monsoon flow, when the vertical extent of southerly (i.e. positive) meridional wind direction is less than 1500 m height, or deep monsoon flow, when it is greater than 1500 m height. It shall be stressed that several authors use the depth of a positive zonal (i.e. westerly) component as an indicator of the depth of the mon-
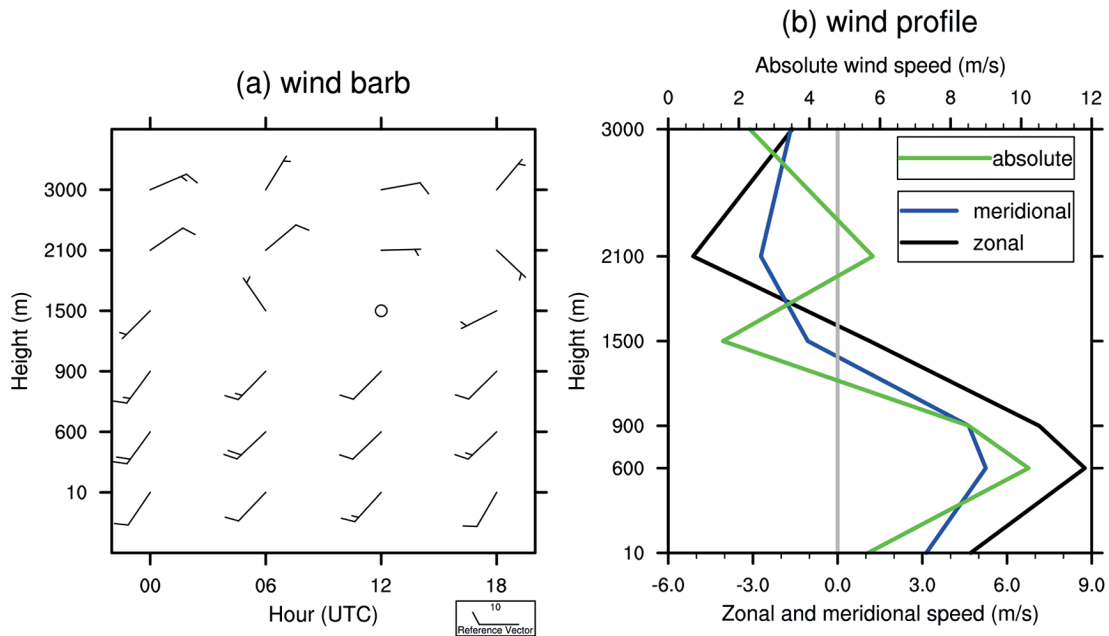
Figure 4: Example of NLLJ situation at Cotonou: (a) Diurnal cycle of low-level winds and (b) vertical profile of the zonal and meridional wind components at 06 UTC, both averaged for September 2015.
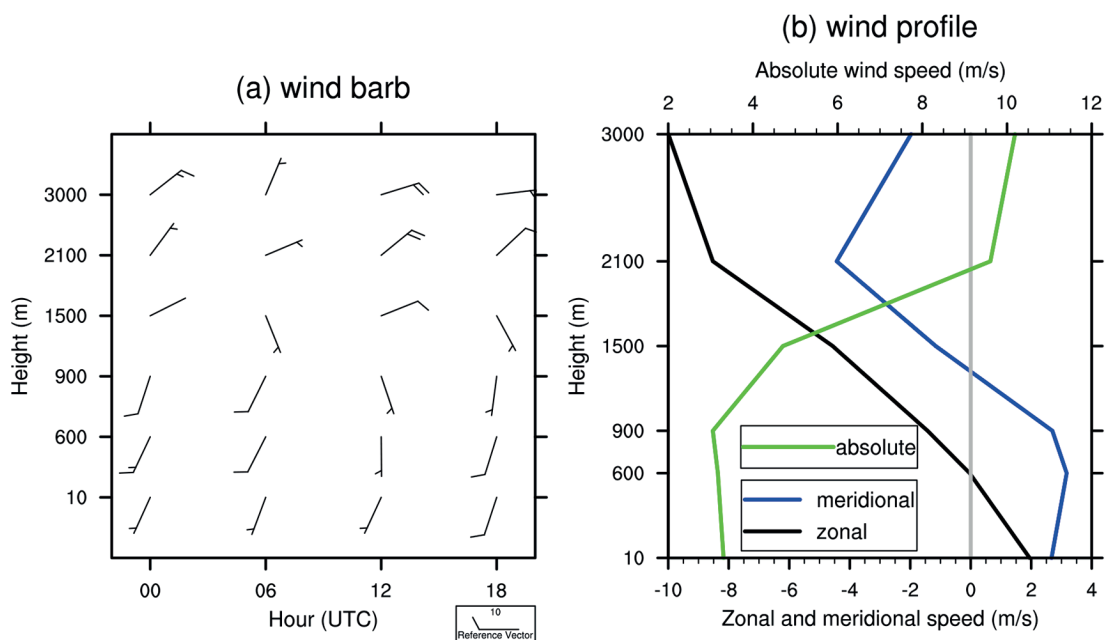

Figure 5: Example of monsoon situation at Cotonou: (a) Diurnal cycle of low-level winds and (b) vertical profile of the zonal and meridional wind components at 12 UTC, both averaged for April 2015.

soon flow. As can be seen in Figure 5 for the example of 12 UTC in April 2015, our definition based on the meridional wind is better suited to characterize the origin of air from the tropical Atlantic below 1500 m and from the continent above this height. However, for 18 UTC in September 2015 (Fig. 4), the difference is not as clear-cut. In the layer where winds turn from low-level monsoonal southwesterlies to mid-level easterlies, weak wind velocities and thus variable directions are observed, rendering both definitions somewhat uncertain in this layer. However, over larger periods a reli-able assessment of the monsoon layer depth is possible. All non-monsoon days are Harmattan days by definition (see above).

3   Results

This section will present the results of the objective analysis of wind regimes in Cotonou. The first sub-section contains a general climatology followed by a detailed discussion of the four individual regime
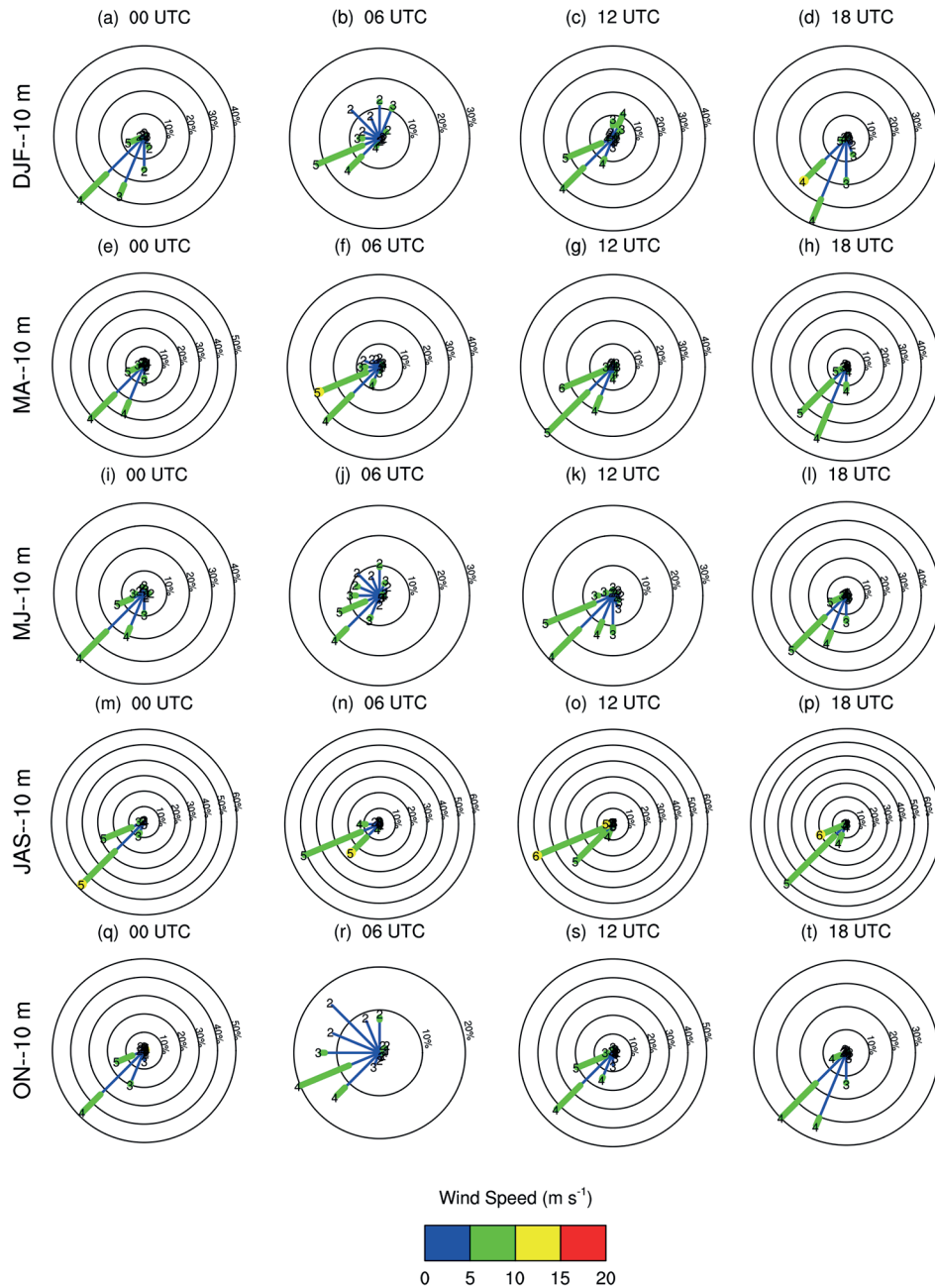
Figure 6: Diurnal cycle of 10 m wind roses at Cotonou for the five main seasons averaged over the period 2008–2015.

types in Section 3.2. The analysis is based on monthly means and considers the following five seasons in the Guinea coastal region: (1) the long dry season in December–February (DJF), (2) the beginning of the long rainy season in March-April (MA) when precipitation amounts are relatively low, (3) the long rainy season in May–June (MJ), (4) the little dry season in JAS and (5) the short rainy season in October-November (ON). Depending on data availability either the period from 2006 or from 2008 to 2015 is analysed.

3.1 Climatology of low-level wind regimes

Figure 6 shows the diurnal cycle of 10 m wind speed and direction at Cotonou for the five main seasons us-

ing wind roses. The LB is clearly observable at 06 UTC when light northwesterly to northerly winds occur. It is most pronounced in DJF but is also detectable in MJ and ON. It vanishes in MA and JAS when the monsoon flow is strong. The situation in MA will be discussed in more detail below. In DJF, northerly winds are also observed at 12 UTC but these are more likely related to Harmattan episodes. It is unlikely that the LB persists until that time of day when the difference between the land and sea temperatures should be positive. With regard to the vertical extent of the LB system, Figure 8 shows that it does not reach 600 m, and in the absence of wind observations with higher vertical resolution, the exact determination of its depth is impossible. At 12 and 18 UTC, winds are dominated by southwesterly direc-
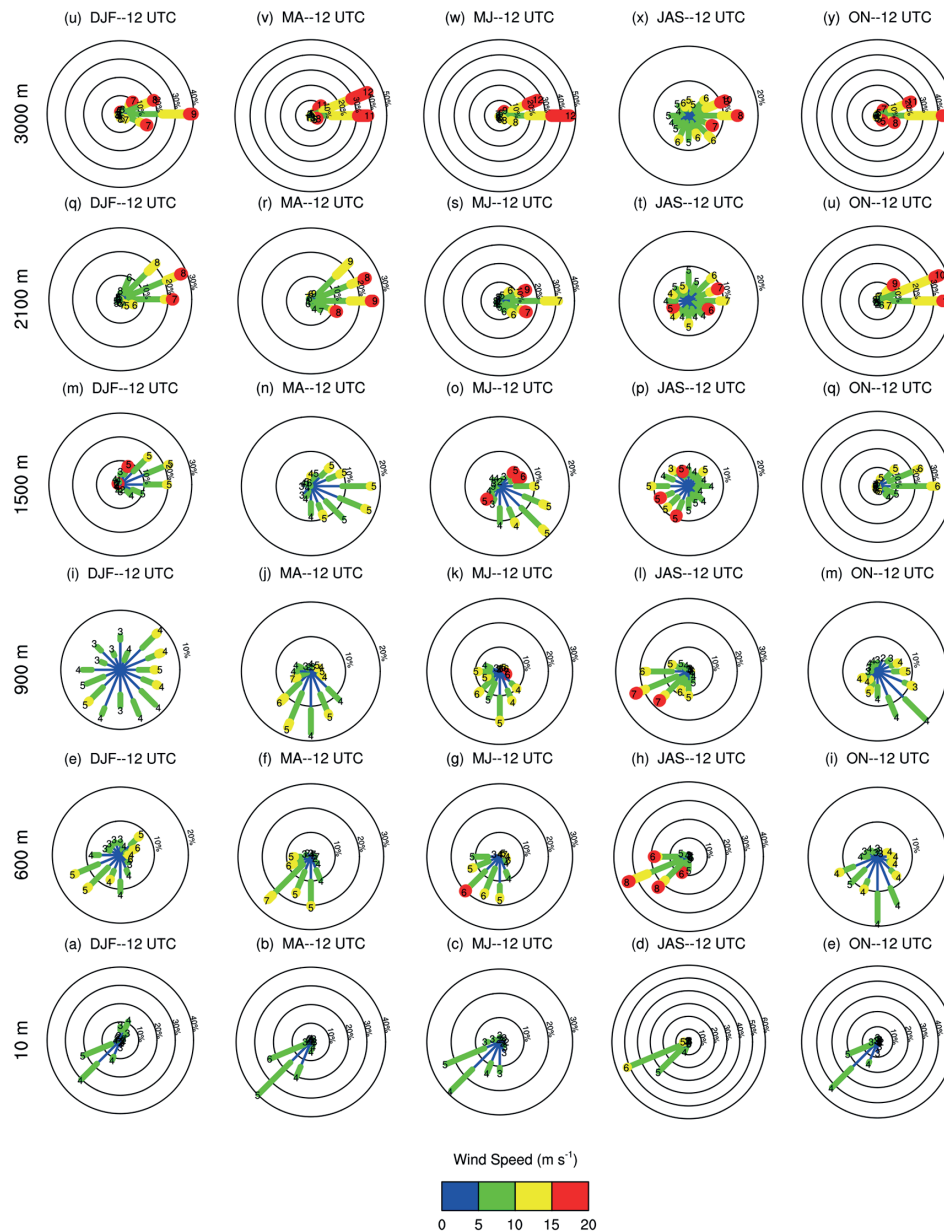
Figure 7: Vertical profile of wind roses for 12 UTC at Cotonou for the five main seasons averaged over the period 2008–2015.

tions throughout the year (right columns in Fig. 6). This is a combination of the large-scale monsoon flow with the more local SB circulation. Wind speeds peak at the height of the WAM in JAS and show similar magnitudes during the rest of the year. Winds at 00 UTC (left column in Fig. 6) largely resemble those at 18 UTC with a very slight shift to more westerly directions.

Figure 7 shows the vertical structure of the wind distribution for all seasons for the 12 UTC launches. As discussed above, the occasional presence of northeasterly winds at the surface suggests the occurrence of Harmattan winds at the coast in DJF (left column), when the ITD reaches its most equatorward latitudinal position. These synoptic-scale winds are also present (and stronger) at higher levels up to 1500 or even 2100 m height. Southwesterly winds indicating a monsoon or SB-type flow are restricted to the levels below 900 m,

indicating a shallow monsoon layer. MA (second column from left in Fig. 7) shows a stronger concentration to the southerly and southwesterly sector at low levels, most evident at 900 m, above which the flow is dominated by easterlies. This signifies a first deepening of the monsoon layer. MJ (middle column in Fig. 7) shows rather moderate changes in direction and speed relative to MA. More marked changes occur towards JAS (second column from right in Fig. 7) when the three lowest layers are clearly dominated by strong and stable southwesterlies. The three top levels show very variable wind directions and a slow tendency to increasingly easterly flow with height, reflecting the day-to-day variations in monsoon layer depth. Finally, ON (right column in Fig. 7) shows wind structures resembling MJ but with a slight shift to more easterly flow, particularly at 1500 m.
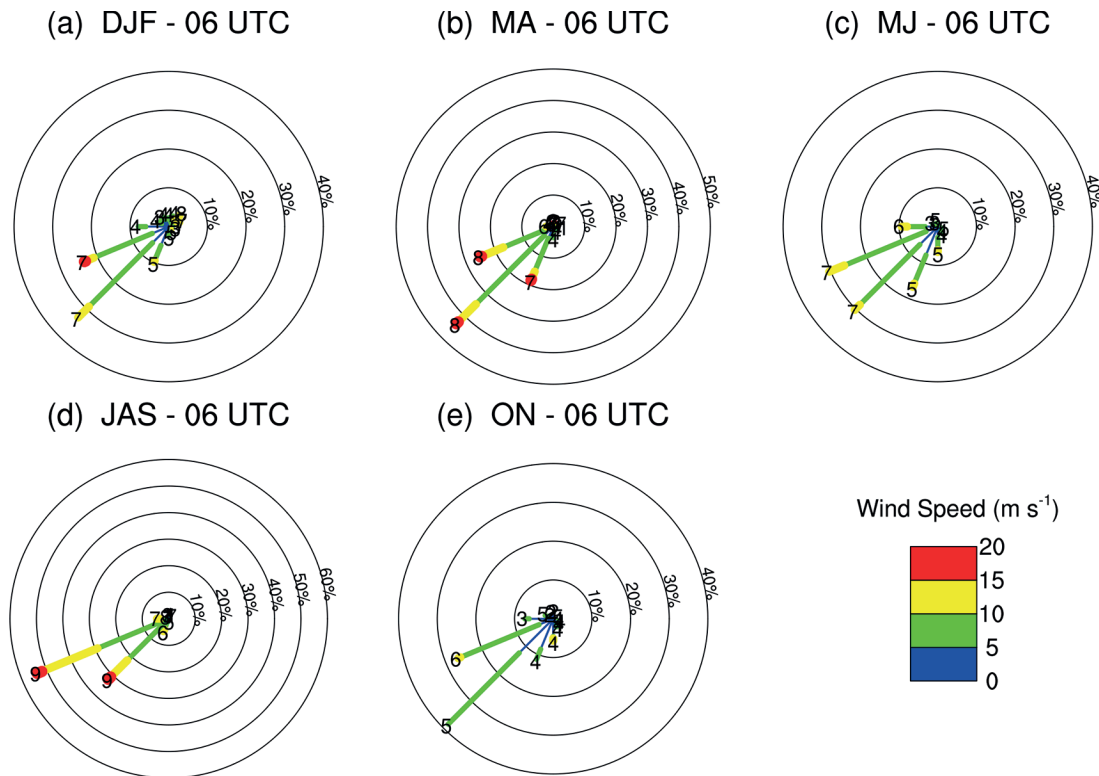
Figure 8: Seasonal cycle of 600 m wind roses for 06 UTC at Cotonou averaged over the period 2008–2015.

Another feature of the low-level wind regime is the formation of NLLJs. This is best illustrated by the wind velocity at 600 m in the morning hours (i.e. at 06 UTC, Fig. 8). In Cotonou, the average velocity of the NLLJ at 600 m exceeds $7\,\mathrm{ms}^{-1}$ in many seasons except in ON when it remains relatively weak. Highest wind speeds are observed in MA and JAS. The dominant direction is southwesterly in all seasons but JAS when the maximum shifts to west-southwesterly.

3.2 Objectively identified statistics

Using the detection criteria defined in Section 2, here we investigate the characteristics of the four main low-level wind regimes in the coastal areas of Benin with regard to their seasonal fluctuations and the physical processes which influence their activity.

3.2.1 LB and SB

Using the detection criteria outlined in Section 2, 996 SB and 1291 LB events are identified over the period 2006–2015. This difference shows that the two do not always coincide. Figures 9a and b show the relative frequency distribution of LB and SB winds at 06 UTC and 15 UTC, respectively, at Cotonou. LB winds occur between 270° and 40° with low speeds (up to $5\,\mathrm{ms}^{-1}$). SB winds preferably blow from 180° to 240° with a maximum frequency at 220° (45 %) with velocities that can reach $10\,\mathrm{ms}^{-1}$ or more.

The monthly variations of LB, SB and combined LSB occurrences are presented in Figure 9c. SB events show a rather unimodal behaviour with a peak in November-December (more than 50 %) and a marked minimum in MJ around 15 %. In sharp contrast, LB events have a bimodal distribution with maxima during October–January and during the long rainy season MJ. December stands out with frequencies well above 70 %. The minimum with less than 10 % is reached at the height of the WAM in August. Particularly in MJ and JAS, LB and SB frequencies differ to a great amount suggesting that they do not regularly coincide. In fact the frequency of joint occurrence (green line in Fig. 9c) has a clear minimum in boreal spring and summer. The maximum occurs during the long dry season with a peak in December (up to 37 %). Apart from August, the frequency of joint occurrence is always well below the minimum of either of the two curves, revealing that in almost all seasons there are many days with either SB or LB occurrence.

The monthly variation of SB described by BAJAM-GNIGNI GBAMBIE and STEYN (2013, their Fig. 3) is rather similar to the monthly behaviour of LB shown in Figure 9 (the blue curve). Since the detection of SB events by the authors is based on the reversal of the wind direction, a SB day is detected only if it is a LB day. Thus their SB frequency curve nearly coincides with the LB frequency curve that we obtained. It is well known that the LSB systems are often influenced by large-scale synoptic flows. In order to identify SB passages, the requirement of offshore winds in the early morning, usually
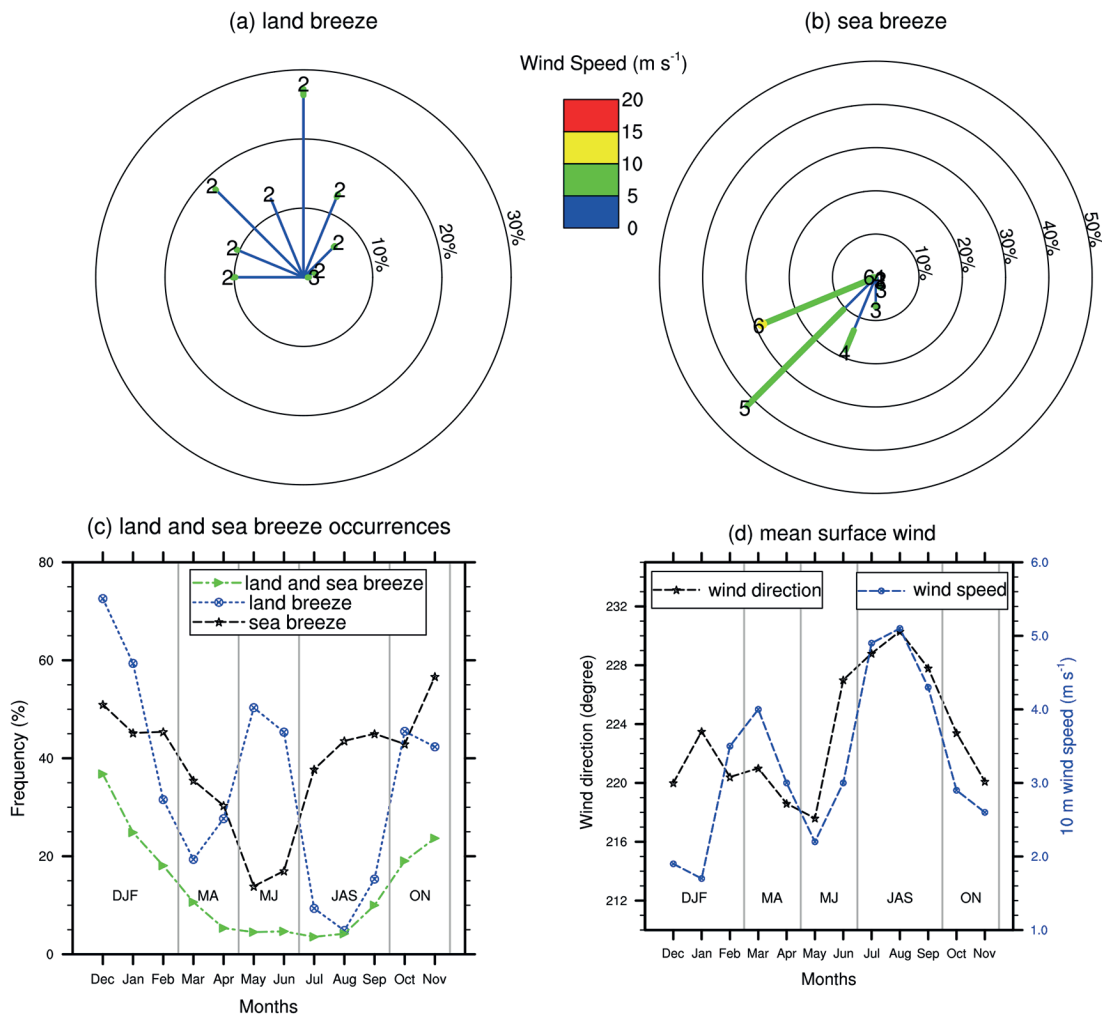
Figure 9: LB and SB climatology at Cotonou during 2006–2015. Wind roses of all detected (a) LB and (b) SB events. (c) Monthly average frequency of occurrences of LB, SB and combined LSB events. (d) Monthly mean 10 m wind direction and speed. The five main seasons are marked by vertical lines.
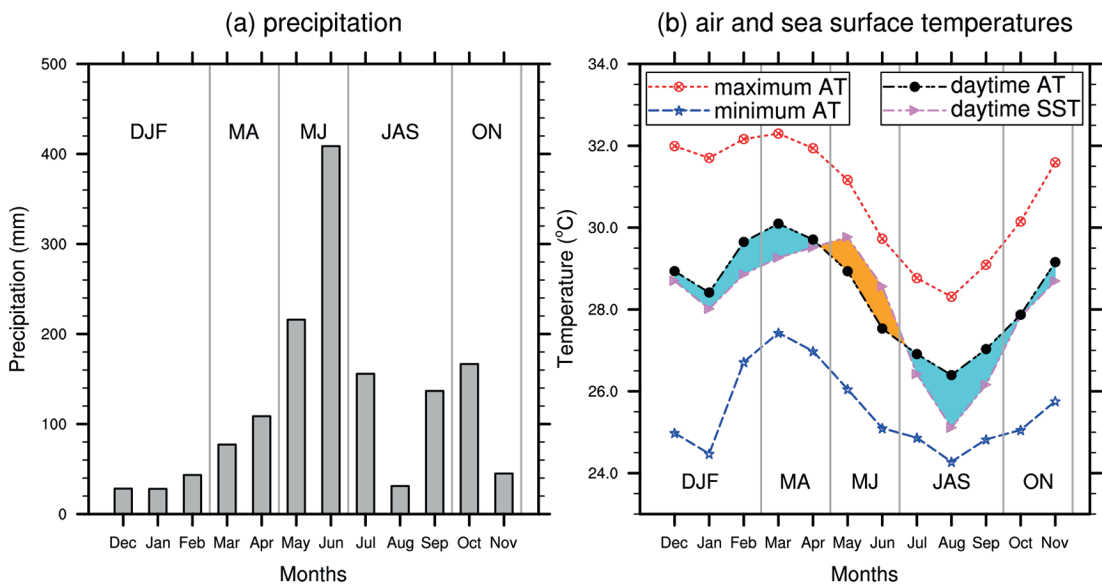

Figure 10: Monthly average (a) precipitation, (b) 2 m air temperature (AT) and sea surface temperature (SST) at Cotonou station over the period 2006–2015.

employed by many authors, is not appropriate in coastal regions of Benin.

Figure 9d shows the seasonal cycle of 10 m wind speed and direction at Cotonou. It shows a clear anti-correlation between wind speed and LB frequency, most evident in JAS when mean winds reach 5 ms^{-1}. As the mean direction varies by less than 15 degrees, this figure suggests that the stronger the southwesterlies the less likely is a LB detection. Such a straightforward interpretation does not exist for SB events.

Figure 10 shows seasonal cycles of precipitation and the differences between 2 m air temperature over land (maximum, minimum and daytime mean) and SST. From November–January, mean land-sea temperature differences are small but positive. These months are characterized by small precipitation amounts (Fig. 10a) and moderate SSTs. Despite the insolation minimum, daytime temperature maxima are fairly high but night-time minima are low during dry conditions. Mean wind speeds are low (Fig. 9d) and LB, SB and joint LSB events are frequent (Fig. 9c). On these "classical" LSB days, winds turn clockwise during nighttime to reach the LB northwesterly direction in the morning and then turn counterclockwise to take the characteristic south-westerly SB direction in the afternoon (not shown). Infrequent Harmattan bursts that often occur in the morning could contribute to increasing the frequency of LB events, as no filter was used to differentiate Harmattan winds from the LB system.

Responding to the increase in insolation, February and March show the highest daytime maxima, warmer SSTs and much warmer nights, the latter probably related to a moister atmosphere with more clouds and precipitation (Fig. 10). This leads to a large land-sea temperature contrast, overall strong winds (Fig. 9d), a sharp drop in LB occurrence and still relatively frequent SBs (Fig. 9c). After a transition in April, the MJ rainy season is characterized by a sharp drop in daytime temperature maxima, partly related to cloudy skies and land surface cooling through precipitation and evaporation, while the ocean remains relatively warm (Fig. 10). The associated reversal of land-sea temperature differences leads to overall weak winds, a sharp increase in LB events and a drop in SB events (Fig. 9c and d). In the following little dry season (JAS), SSTs drop to their annual minimum due to coastal upwelling along the Bight of Benin, leading to a strongly positive thermal forcing despite the relatively low daytime 2 m air temperature over land (Fig. 10b). This explains, in combination with the peak monsoon wind strength, the very strong mean winds associated with the overall LB minimum and frequent SBs (Fig 9c). Finally, October marks a transition to the dry-season conditions described above.

Another important feature of the LSB system is the inland penetration of the SB front that can lead to the development of convective systems. It depends on several factors including the pressure gradient and the difference in temperature between the coast and the inland. To analyse this aspect, Figure 11 shows the diurnal cycle of monthly averaged air temperature and mean sea-level pressure (MSLP) between the stations of Cotonou and Savè (located at 8° 02′ N) that are 185 km apart. It shows that the temperature difference between the inland and the coast is negative during nighttime (Fig. 11a and c) and positive during daytime, except in JAS at 12 UTC (Fig. 11e and g). The absolute value of this difference in temperature is greater between December and April than during the other months of the year. During DJF it is likely amplified by the occurrence of Harmattan cool and dry air, while in MA it is rather linked to the strong warming at the surface away from the coast. During MJ, JAS and ON, Cotonou is affected by the relatively cold SSTs, while inland areas tend to be cloudier in the afternoon.

The right column of Figure 11 shows the impact on MSLP. At 00 and 06 UTC, MSLP at Savè is slightly higher than at Cotonou during the entire year. At 12 and 18 UTC, a striking feature is the large daytime drop in MSLP at Savè during February to April, when the surface pressure trough is located just north of the coast at 8° N (Fink et al., 2010). This results in a tighter pressure gradient and suggests a potentially deep penetration of the SB. After the monsoon jump in June, when the pressure trough moves over the Sahel, the MSLP gradient between the coast and the inland decreases, due to the strong activity of the St. Helena High, which strengthens the southwesterly flow. It shall be stressed, however, that some caution is necessary in the interpretation of small pressure differences since barographs in West Africa are often not well calibrated and systematic errors in the reduction of station pressure to MSLP may occur.

3.2.2 Harmattan

In Cotonou, Harmattan synoptic winds are preferably from northerly to northeasterly direction (between 0° and 45°) and their average speed is about 4 ms^{-1} (Fig. 12a). Their occurrence is generally preceded by a rapid increase in MSLP over the Sahel (not shown). The phenomenon only appears in DJF (Fig. 12b) with about 30 % frequency in January followed by just under 20 % in December and about 7 % in February. During this periods the air temperature is low in the early morning and high in the afternoon (Fig. 10b), leading to a significant diurnal thermal amplitude in an often cloudless sky, which, however is often covered by aerosol, e.g. dust from the Sahara and local pollution.

3.2.3 Monsoon flow

The depth of the monsoon flow is crucial, as the associated moisture transport from the ocean contributes to modulating the diurnal variation of temperature, cloudiness and rainfall at the coast. The classification of the monsoon flow shown in Figure 13 confirms the expected strong contrast between the dry season with a shallow monsoon (85 % frequency in December) with values as low as to 500 m height and the rest of the year, when the
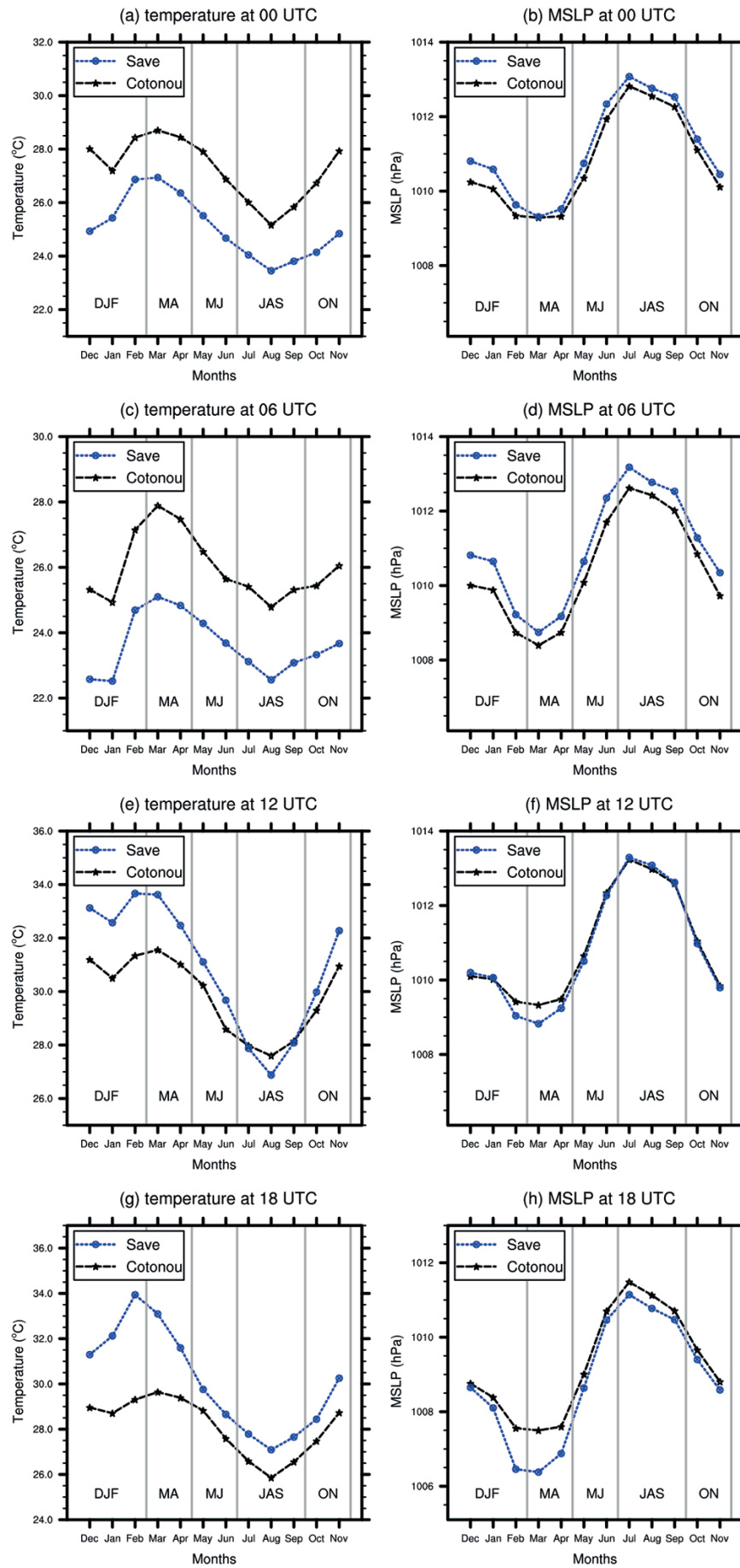
Figure 11: Diurnal cycle of monthly averaged 2 m air temperature and MSLP at Cotonou and Savè stations over the period 2006–2015.
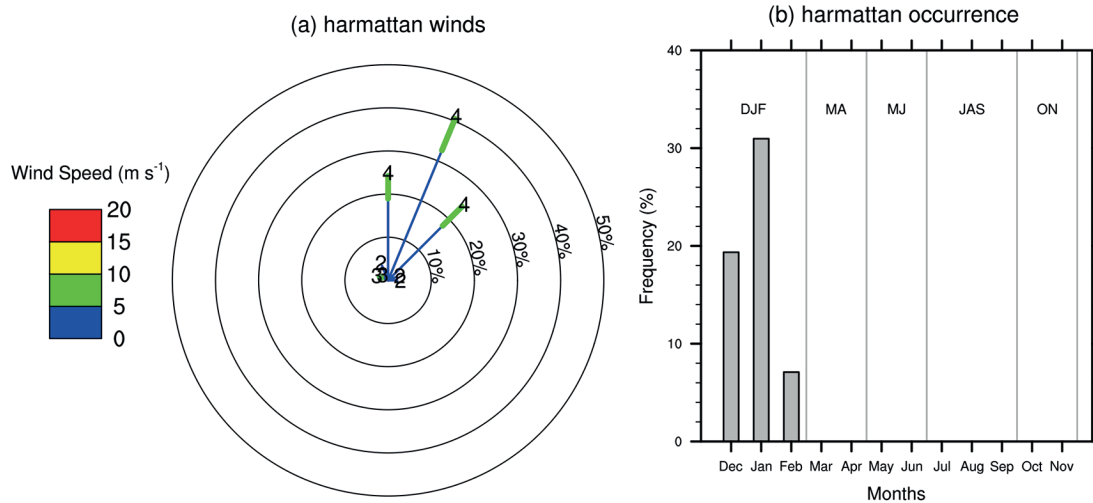
Figure 12: Harmattan climatology at Cotonou for the period 2006–2015. (a) Wind rose at 12 UTC for all detected events and (b) monthly average frequency of Harmattan occurrence.
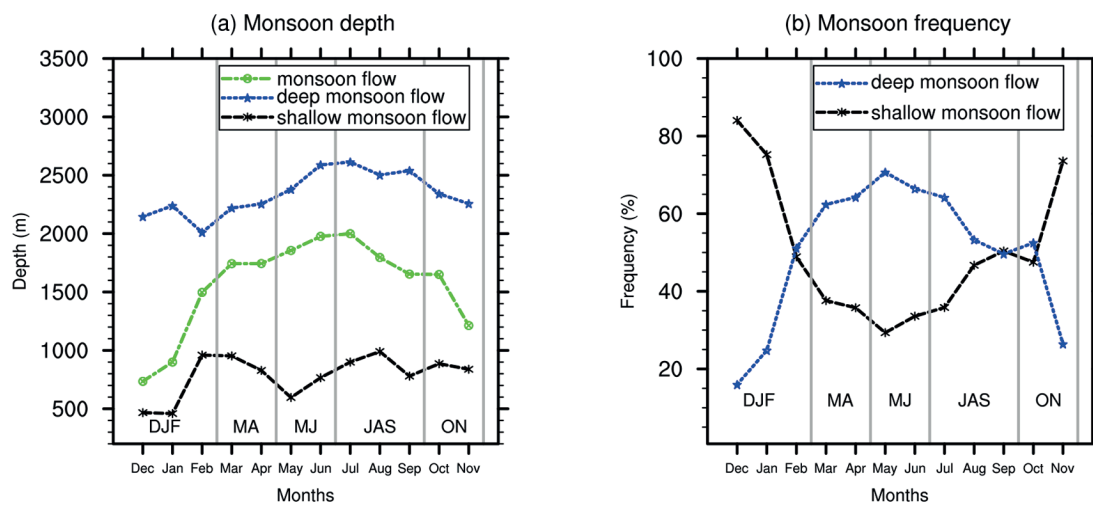

Figure 13: Monthly averaged (a) depth of occurrence and (b) frequency of shallow and deep monsoon flows at Cotonou over the period 2006–2015.

monsoon is much deeper and usually exceeds 1500 m in height. In May during the long rainy season, the monsoon layer depth exceeds 1500 m in more than 70 % of all days and in June and July a peak altitude of 2500 m is reached. Clearly, Figure 13a shows that the average monsoon depth gradually increases until July and decreases until December.

3.2.4 NLLJs

The transport of water vapour from the south at lower levels is accelerated by the NLLJ towards the ITD during night (e.g. PARKER et al., 2005). Statistics on the NLLJ at 600 m height at Cotonou show that on average, its strength is around 10–12 ms^{-1} but can reach 15 ms^{-1} on individual days (Fig. 14a), while the direction is exclusively from the southwest. The NLLJ is consistently observed every month (Fig. 14b) with a maximum activity during the periods of relative stability encountered in February–March, at the end of the long dry season, and

in JAS during the little dry season. Apart from these periods, the occurrence of the NLLJ declines significantly to reach a relative minimum in May in the middle of the long rainy season and in November at the end of the short rainy season. It is conceivable that during the rainy seasons the vertical mixing of momentum associated with frequent moist convective activity and precipitation inhibits the NLLJ occurrence. If in addition to the exceedance of 10 ms^{-1} in wind speed, a second filter criterion is adapted that the overnight wind speed at 00 or 06 UTC is 3 ms^{-1} higher than at 18 UTC, then the NLLJ peak in August decreases to be about equal to the frequency of the second peak in March (Fig. 14a). We prefer the extended definition, as it prevents the detection of many NLLJs in the high-wind period of JAS on days, when there is no distinct diurnal cycle.

Figure 14c shows the monthly evolution of the wind velocity at 600 m and 3000 m height and the wind shear between these two levels. The monthly evolution at 600 m (blue solid line) corresponds well to the clima-
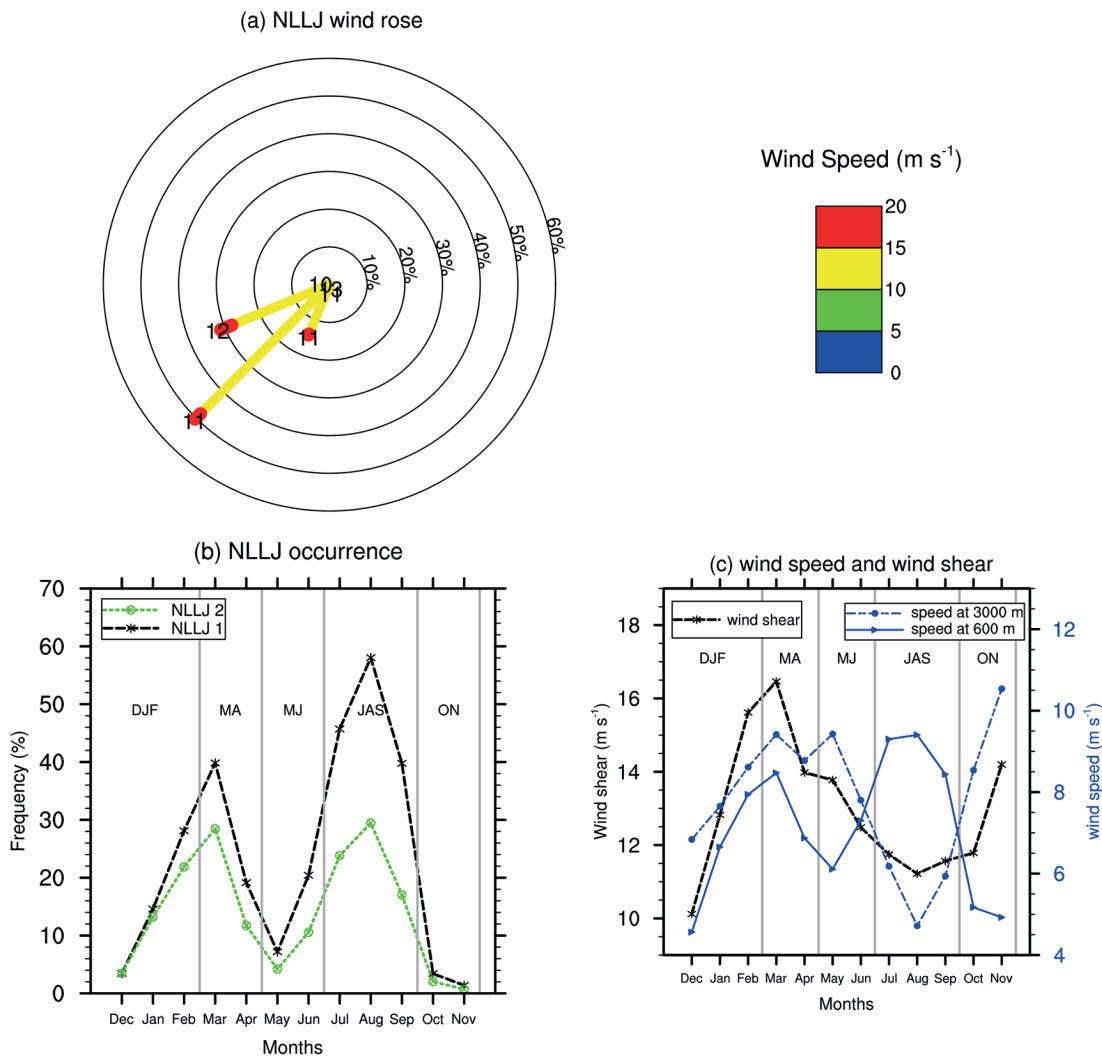
Figure 14: Climatology of NLLJ at Cotonou over the period 2006–2015. (a) Wind rose of all detected NLLJ events. (b) Monthly average frequency of NLLJ occurrence computed using the exceedance filter of $10\,\mathrm{ms^{-1}}$ in wind speed (NLLJ 1) adapted with a second filter that the overnight wind speed at 00 or 06 UTC is $3\,\mathrm{ms^{-1}}$ higher than at 18 UTC (NLLJ 2). (c) Monthly average wind shear between 3000 m and 600 m and monthly average velocity at 600 m and 3000 m.

tology of NLLJ1 shown in Figure 14b with two peaks in March ($14\,\mathrm{ms^{-1}}$) and August ($15\,\mathrm{ms^{-1}}$). The wind shear (in $\mathrm{ms^{-1}}$) is computed as the magnitude of the mean monthly vector difference between 3000 m and 600 m height, which corresponds roughly to the shear between 600 hPa and 925 hPa widely used to assess the likelihood of the development of squall line systems in the region (e.g. LAFORE et al., 2017). The shear has a bimodal distribution with peaks in March and (MARANAN et al., 2018). The first peak is the result of directional shear between a strong southwesterly low-level and a strong midlevel easterly flow. In contrast, the very pronounced second peak is almost solely created by the swift passage of the midlevel easterly jet aloft.

3.2.5   Annual statistics

In this section we discuss the frequency of days when certain wind regimes coincide during the period 2008–2015 (Table 2).

As noted above, the large-scale Harmattan and monsoon winds are mutually exclusive. In Benin coastal regions, the annual frequency of the monsoon flow varies year-to year between 88.2 % and 98.4 % while the annual frequency of the Harmattan flow ranges between 1.6 and 11.8 %. Years 2008 and 2015 have experienced the longest Harmattan episodes. The occurrence of NLLJ ranges from a minimum of 9.9 % in 2010 to a maximum of 26.5 % in 2015. These values do not change for the frequency of joint occurrence of monsoon flow and NLLJ, confirming that the coastal NLLJ is a part of the WAM system.

With regard to the mesoscale LSB wind system, the frequencies are about 30.1–43.0 % for the SB and 31.5–42.9 % for the LB. It is worth noting that these mesoscale winds occur in Benin coastal regions on average 1 day out of 3 during the year. The frequency of joint occurrence of LB and SB on classical LSB days, which contribute to reinforce the recirculation of pollutants, is about 9.6–8.6 % during the 8-year period.

Table 2: Annual frequency and joint frequency of occurrence of wind regimes over the period 2008–2015.

| Year	Monsoon	Harmattan	NLLJ	NLLJ and Monsoon	SB	LB	LB and SB	LB and Harmattan	SB and Monsoon	LB and Monsoon
2008	91.8	8.2	20.0	20.0	39.3	42.9	18.6	6.6	35.8	36.3
2009	96.4	3.6	15.3	15.3	43.0	32.1	14.0	2.7	40.3	29.3
2010	98.4	1.6	9.9	9.9	36.2	38.4	15.9	1.6	35.6	36.7
2011	94.5	5.5	25.1	25.1	41.4	37.5	17.0	5.2	39.5	32.3
2012	96.4	3.6	24.8	24.8	42.1	31.7	11.2	3.6	40.7	28.1
2013	95.3	4.7	20.3	20.3	36.7	31.5	10.4	4.7	35.3	26.8
2014	96.2	3.8	17.9	17.9	30.1	37.5	10.7	3.8	29.0	33.7
2015	88.2	11.8	26.5	26.5	36.7	31.5	9.6	11.0	35.1	20.5
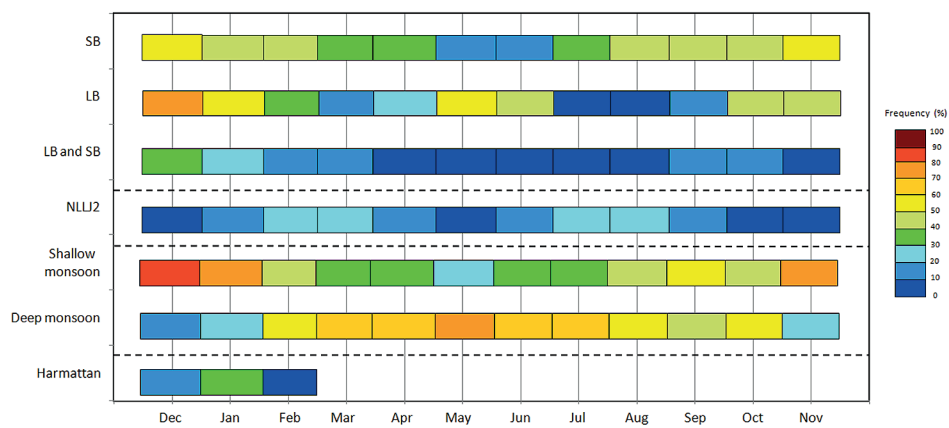
Figure 15: Bar chart schematic representation of the frequency of occurrence of the different low-level wind regimes in Benin coastal areas.

Another interesting feature is the daily joint occurrence of large-scale and mesoscale winds, both of the same direction. In the case of Harmattan and LB, their joint frequency ranges between 1.6 % (2010) and 11.0 % (2015), indicating that from morning until early afternoon, the coastal regions remain under the influence of northerlies in an often cloudless sky. With regard to monsoon and SB flows, their joint occurrence increases the speed of southwesterly winds and fluctuate between 29.0 and 40.7 %. In contrast, the occurrence of LB in the early morning can be followed by the reversal of the wind direction. When the southwesterly winds that occur following this change of direction do not fulfill the SB filters set above, they are classified as monsoon flow. In this case, the joint occurrence of monsoon flow and LB is estimated between 20.5 and 36.7 % during the period 2008–2015.

4 Summary and conclusion

Seasonal and diurnal climatologies of four coastal wind regimes (the monsoon and Harmattan winds, the NLLJ and the LSB systems) were presented for the station of Cotonou over the period 2006–2015.

For the seasonal climatology, wind roses were utilized to describe the diurnal cycle and the vertical structure of the low-level wind regimes using winds from PILOT and TEMP reports for 2008–2015. The southwesterly monsoon flow is observed in the coastal region throughout the year, except during short periods in DJF when the northerly Harmattan winds occur. Within the monsoon flow, the NLLJ is best represented at 600 m a.g.l with a seasonal average velocity of $7\,\mathrm{ms}^{-1}$. These large-scale flows are often reinforced in Cotonou by the occurrence of mesoscale LB and SB winds and it is difficult to clearly separate them in case of joint occurrence. The LB offshore wind is sharply observable at 06 UTC with speeds of less than $5\,\mathrm{ms}^{-1}$, while the southwesterly SB flow during the afternoon reinforces the large-scale monsoon flow by up to $5\,\mathrm{ms}^{-1}$.

The climatology is based on detection methods using a set of filters to assign a given day to one (or several) wind regimes or to classify it as "unspecified". On the basis of these statistics, the bar chart presented in Figure 15 shows a summary of the frequency of occurrence of these different low-level wind regimes in Benin coastal areas. A visual analysis of this figure confirms that the monsoon flow, the LB and the SB are the most frequent wind regimes encountered throughout the year, in contrast to the Harmattan winds and the NLLJ2. However, it is worth noting that the joint occurrence of LB and SB (LSB) systems is relatively low, especially during the rainy seasons. The study has demonstrated that:

- LB and SB days do not always coincide: in almost all seasons there are many days with either SB or LB occurrence (Fig. 15);
- The monthly variation of the frequency of SB events shows a unimodal behaviour with a peak in

November-December (more than 50 %) and a minimum in MJ around 15 %, while the LB curve shows a rather bimodal distribution;

- When southwesterly winds strengthen, the LB is less likely detected, since the surface wind speed and the LB frequency show a clear anti-correlation, most evident in JAS, when mean winds are 5 ms^{-1};
- Many processes such as insolation, the tightening of the pressure gradient between the coast and the inland trough (8° N) in MA, the precipitations in MJ and coastal upwelling in JAS contribute to modulate or disrupt the occurrence of the LSB systems.

The LB system is also disrupted by the Harmattan winds in DJF mainly in the early morning but no distinction has been made between the LB and the Harmattan winds at this time of day. The occurrence of Harmattan winds in DJF is followed by an increase in the diurnal amplitude of temperature and a decrease in humidity at the coast. Apart from these Harmattan episodes, high humidity is almost permanent, as is the monsoon flow that transports it. The humidity contents in the lower atmosphere depends in part on the depth of the monsoon flow and the classification made in this study has shown that during periods of heavy rainfall the monsoon depth exceeds 1500 m on average. Within this monsoon flow, the activity of the NLLJ increases during periods of relative stability in the dry seasons when turbulent mixing is lower. At the daily scale, the NLLJ consistently shows speeds up to 15 ms^{-1}, and even above on individual days. These values of speed are much higher than the peak velocities (6–8 ms^{-1}) between 400 and 500 m above ground level (a.g.l.) showed by Schuster et al. (2013, their Fig. 4) using mean low-level wind profiles from radiosondes for JAS 2006.

Annual statistics on the occurrence and joint occurrence of certain wind regimes have shown that the frequencies of mutually exclusive large-scale flows fluctuate year-to-year from 88.2–98.4 % for monsoon and 1.6–11.8 % for Harmattan winds. On average, LSB systems occur 1 day out of 3 during the year. While Harmattan and LB winds imply offshore flows with a joint frequency of 1.6–11.8 %, the potential joint occurrence of monsoon and SB winds fluctuated between 29.0 and 40.7 % during investigation period. Overall, this study helps to improve the understanding of low-level wind at the Guinea Coast. It has demonstrated that the reversal criterion in surface winds to define LSB days, as used in Bajamgnigni Gbambie and Steyn (2013), may obscure SB occurrence during periods of strong background winds as for example at the peak of the West African monsoon region season. Changes in the strength, or even the absence of the coastal upwelling in boreal summer along the Guinea Coast might change the seasonal cycle of the LSB system, as indicated in Abayomi et al. (2007). Future studies shall look at other locations along the Guinea Coast with the limitation of lacking upper-air observations. Lower-tropospheric wind shear that is known to be relevant for the oc-

currence of linear convective systems, i.e. squall lines (Lafore et al., 2017), is largest during March-April due to strong NLLJ and AEJ winds. Consistent with this, Maranan et al. (2018) found the strongest and most organized systems at the Guinea Coast at this time of the year. Thus, the seasonal cycle of the low-level wind system impacts on organization of convection at the coast. In addition, afternoon convection gets triggered at the SB front. The present study provides a foundation for more comprehensive study of these aspects and other phenomena associated with coastal wind systems, such as the dispersion of air pollution.

Acknowledgement

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