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Introductory Essay

The Caribbean Low-Level Jet: relationship with climate and weather in Central America

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Abstract

Central America and the Caribbean region is conformed by countries and islands among the poorest of the Americas and even of the world. Many of their inhabitants are living in prone areas to natural disasters. It has been highlighted that most of the natural disaster reports in the last years, have been related to hydro-meteorological phenomena. Indeed, it is well known that this region is constantly affected by a wide variety of meteorological events, such as traveling easterly waves, tropical cyclones, convective systems, cold surges coming from the northern hemisphere, the mid-summer drought (MSD), the warm pools, the trades, and an intense low-level jet over the Caribbean Sea. It should be understood that variations in the intensity of these elements in combination with the social factors (i.e. people living in risky areas), can increase the probability to suffer natural disasters. In that sense, within the framework of the Center for Natural Disaster Science (CNDS), the present essay reviews the current knowledge on one of the most striking climate features of the region, the Caribbean low-level jet (CLLJ). It has been found in the literature some mechanisms explaining its origin, maintenance, and structure, nevertheless more studies using modeling techniques and observations are required in order to validate such mechanisms. It is known that the large-scale convergence (divergence) at the jet exit (entrance) is associated with precipitation patterns in the Central American isthmus. It was found that discrepancies calculating precipitation in the General Circulation Models (GCMs) and reanalysis lead to a misrepresentation of the climate in this region, and these discrepancies are hand-to-hand with errors in the models estimating the low-level wind flow. Moreover, some studies highlighted the association of the CLLJ with the vertical wind shear (VWS), as a main mechanism of control, and the influence of the VWS on hurricanes and tropical cyclones. Furthermore, the influence of the large-scale climate modulators such as El Niño (ENSO), the North Atlantic Oscillation (NAO), the Madden-Julian Oscillation (MJO), upon the intensity of the jet have been related with changes in precipitation, sea surface temperatures (SST) anomalies in the Caribbean Sea, and tropical cyclone genesis. In spite that the jet has been identified as an important mechanism modulating the precipitation regime in Central America, the association with extreme phenomena such as droughts and/or heavy rain events is still not understood.

1. Introduction

Central America and the Caribbean regions are home of about eighty million of people. Formed by countries and islands among the poorest of the Americas and even of the world, many of their inhabitants are living in prone areas to natural disaster. Alfaro et al. (2010) pointed out that most of the natural disaster reports have been related to hydro-meteorological phenomena. Indeed, this region is constantly affected by a wide variety of meteorological events, like travelling easterly waves, tropical cyclones, convective systems, cold surges coming front the north hemisphere, the mid-summer drought (in Spanish *canícula or veranillo*), the warm pools, the trades, and an intense low-level jet over the Caribbean Sea. It should be understood that variations in the intensity of these elements in combination with the social factors (i.e. people living in risky areas), can increase the probability to suffer a natural disaster.

Nevertheless, it should be also said that all of these climate elements make the region worth studying in the context of an integrated program, including observations, modelling and processes studies. However, there is very little knowledge on these climate issues in the context of regional weather. One of the most relevant features of this region, practically unknown in regard to its nature, interaction with other regional phenomena of the tropics and extratropics, or in relation to its role in weather and climate, it is the Caribbean low-level jet.

The Caribbean Low-Level Jet (CCLJ) is a wind flow that appears over the Central Caribbean sea. Amador (1998) was the first documenting its presence during the boreal summer. He noticed that this jet develops in low levels of the atmosphere, between 925 and 700 hPa in the central part of the Caribbean ($15^{\circ}\mathrm{N}-75^{\circ}\mathrm{W}$). Later, Poveda and Mesa (1999) have also documented the existence of an air current with similar characteristic to the CLLJ. Its potential interaction with transients, such as easterly waves, makes the jet an important mechanisms for convective activity during the summer months.

In recently studies, a winter component of the CLLJ has been reported (e.g. Amador et al. 2006; Wang 2007; Wang and Lee 2007; Amador 2008; Muñoz et al. 2008). However, no relation between the wind flow and the convective activity has been found during the first months of the year within this region. Like many other LLJs in the world, the CLLJ is also an important mechanism for moisture transport (Amador 1998; Amador et al. 2006; Wang 2007; Amador 2008; Gamble and Curtis 2008; Muñoz et al. 2008; Whyte et al. 2008; Durán-Quesada et al. 2010; Amador et al. 2010).

Although the interest for the CLLJ has increased during the last years, and more features and mechanisms explaining its origin, structure and maintenance have been suggested, there still is uncertainty due to lack of observations (at surface and upper air levels) in the Central America and Caribbean regions. Thereby, campaigns to collect observed data and modeling techniques, such as those documented in Amador (2008), have been suggested in order to improve the understanding on the jet.

In the scope of natural hazards, the relationship of the CLLJ with precipitation gains relevance, as previously mentioned, in Central America most of the natural disasters are related with extreme hydro-meteorological events (Alfaro et al. 2010; Maldonado et al. 2013). In that regard, some studies have analyzed the interaction of this low-level wind flow with rainfall in the Caribbean islands (Gamble and Curtis 2008; Whyte et al. 2008; Martin and Schumacher 2011a) and continental United States (Wang 2007; Martin and Schumacher 2011b; Muñoz and Enfield 2011). Nevertheless, its relationship with precipitation in Central America and mainly with extreme precipitation events is still not well understood. Other studies have analyzed the association of the wind shear and the CLLJ with tropical cyclone activity (Wang 2007; Wang and Lee 2007; Wang et al. 2007; Amador et al. 2010).

Within the framework of the Center for Natural Disaster Science (CNDS, Halldin et al. 2011), the present essay reviews the current knowledge on the CLLJ as modulator of the climate in the Central America and Caribbean regions. Thus, this paper is organized as follows. In Section 2, an overview of the mean conditions of the wind field, humidity and precipitation, is given in order to understand the climate in the region. In Section 3 the most important variability indexes such as El Niño Southern Oscillation (ENSO), and the Atlantic Multi-decadal Oscillation (AMO), are studied in order to describe how they could change the mean conditions for the jet during its maxima. Furthermore, the regional- and local-scale variability elements are also discussed. General remarks on the LLJs, main features and the importance of the CLLJ for the climate in Central America are exposed in Section 4. In Section 5 the data sets available (observations and models), and the mathematical tools such as the dynamical and statistical downscaling techniques are examined, in order to know their abilities and drawbacks doing regional and local climate and weather studies. Section 6 sets forth the future work and research plan expected to be developed within the SIDA-funded research project – the Central America Disaster Mitigation program.

2. Overview of the climate in the Intra-Americas Seas

The geographical location of Central America plays a significant role to describe the climate, and hence, the variability of the region (Fig. 1). Surrounded by two large water masses, the eastern tropical Pacific (ETPac) ocean at the west side (Fiedler and Lavín 2006), the Caribbean Sea and North Atlantic ocean at the east side; along with the Gulf of Mexico, the entire area is known as the Intra-Americas Seas (IAS, Amador 2008). The IAS is also important in the global energy balance, because it receives large amounts of radiant energy coming from the Sun into the Earth surface, and the regional waters act like an energy reservoir.

The IAS region is sensitive to the effect of both large- and regional-scale dynamical systems acting in its vicinity, such as the North Atlantic Subtropical High (NASH), strong easterly winds, a large latent heat belt, warm sea surface temperature SST and intense precipitation. All of these elements depict the general frame of the regional climate and variability (Wang et al. 2008) in the region. Furthermore, the formation of the Western Hemisphere Warm Pool (Wang and Enfield 2001, 2003; Wang and Fiedler 2006), cyclogenesis over the Caribbean Sea (Goldenberg et al. 2001), the influence of the Inter Tropical Convergence Zone (ITCZ) along with mid-summer drought (MSD, Magaña et al. 1999; Karnauskas et al. 2013) on precipitation, and the CLLJ form part of the vast variety of regional climate components present in this area. It should be noticed that many of the dynamical and physical mechanisms and interactions with the climate of the region are still not fully understood (Amador 2008).

2.1. The wind field

The trade winds are the low-level tropospheric flow, classically known as part of the equatorward branch of the Hadley cell transporting a large amount of the moisture convergence flux at the low latitudes. They are responsible of the convective activity and

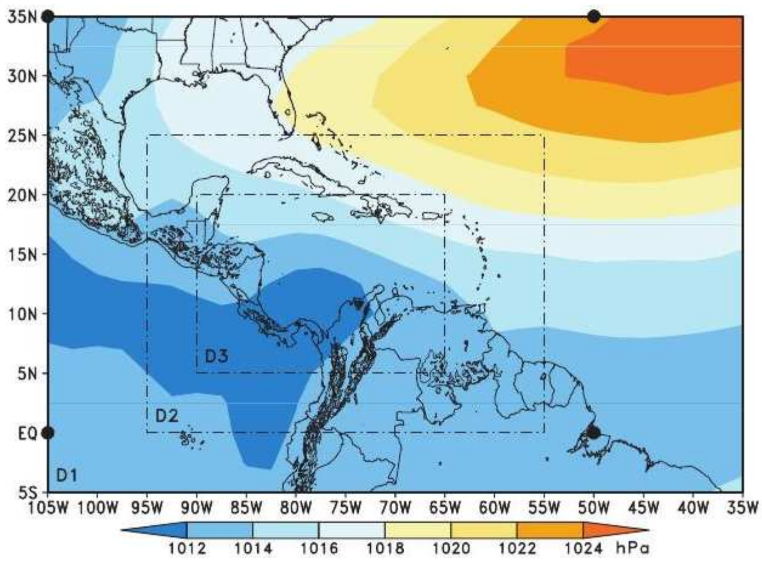

FIG. 1. The Intra-Americas Seas region is represented by the square with large black dots (D1), as defined in Amador (2008). The dashed squares are nested modeling domains used in Amador (2008). In this figure, the July long-term mean (LTM) sea level pressure hPa (1958 – 1999) over the IAS region, is also shown. Note the large northeast-southwest pressure gradient over the Caribbean Sea.

associated precipitation distribution that take places near or within the ITCZ. At local scales, the interaction with the topography helps to explain the temporal and spatial rainfall variability in some areas of Central America (Amador et al. 2003). Figure 2 show the annual mean of the sea level pressure and the annual mean of the horizontal wind field. Note that the most striking feature on this figure is the subtropical high-pressure systems located near 30° N, in both the Pacific Ocean (the North Pacific High) and in the North Atlantic Ocean (the Bermudas or Azores High). Notice that in average, there is a relative strong meridional pressure gradient between the subtropics and tropics, which accelerates the air masses towards the equator, the trades winds.

Monsoonal systems such as the North American Monsoon System (NAMS) and the South American Monsoon System (SAMS) interact in different ways with the trade winds, and are very important mechanisms to explain the precipitation during the warm season in the ETPac. Vera et al. (2006b) review both systems with more details. According to Amador et al. (2006), the most relevant for ocean-atmosphere dynamics in the ETPac is the NAMS.

During the northern hemisphere winter, the ITCZ is at its southernmost position (Srinivasan and Smith 1996), and SSTs over the adjacent areas of the Caribbean and the Pacific are relatively uniform, with values usually below 28° C (Amador et al. 2003). Trade winds are intensified and a frequent southward displacement of air masses occurs. From December – March cold air masses coming from Canada and the polar regions penetrate deep into the tropics and produce strong wind events associated with intense periods of rainfall (Schultz et al. 1997, 1998). Interaction with topography (i.e. wind is funneled trough topographical gaps in southern Mexico and Central America), and a strong gradient in the SLP between the basins, a low in the Pacific and a high in the Caribbean, due to the intrusion of the cold fronts, produce strong near-surface wind events that can to extend far in the ETPac. High sea level pressure (SLP) over the southwestern Caribbean generates northerly surface winds across the Isthmus of Panama that extend offshore of

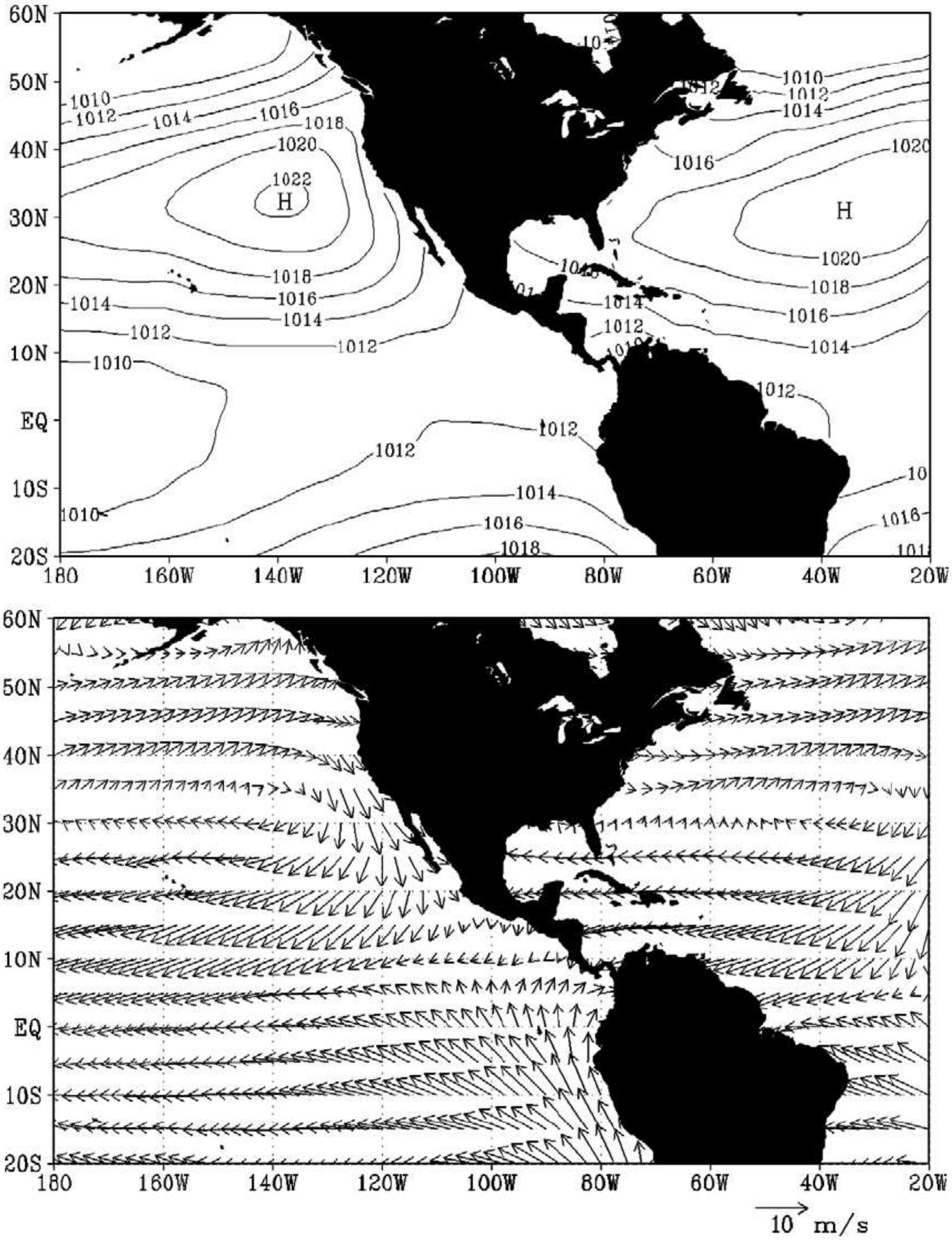

FIG. 2. The annual mean sea level pressure (hPa) showing the high-pressure systems near 30° N and southwest of Peru in South America (top). The pattern of annual mean horizontal wind vector (ms^{-1}), near the surface (bottom). From Amador et al. (2006).

the Gulf of Panama into the eastern tropical Pacific region (Amador et al. 2006). It is noteworthy that during this season these very intense winds over the Caribbean have been associated with the winter branch of the CLLJ (Wang and Lee 2007; Amador 2008), and its interaction with mountains forcing rainfall in Central America.

During the northern hemisphere summer, more complex circulations than those in winter develop in the IAS region. The NAMS, a monsoon-like circulation due to the seasonal thermal contrast occurs in this season over the ETPac, and the North, Central and South America lands to the east (e.g. Higgins et al. 2003). The NAMS has been shown to be associated with the summertime precipitation of the region (Mock 1996; Higgins et al. 1997). It starts to develop during May – June, and associated with heavy rainfall in late May or early June over southern Mexico, moving north to the south-western United States. The mature phase is reached during July-August and September, and the precipitation regime on North America is related to NAMS (Higgins et al. 1997). Surges of maritime tropical air that move northward and that are associated with rainfall over the Gulf of California and southwestern United States (Douglas and Leal 2003). The Decay phase (late September – October) can be characterize as broadly the reverse of the onset phase, though at a slower rate (Higgins et al. 2003).

Other low-level flows occur during the boreal summer, particularly the Gulf of California low-level jet (Douglas 1995; Douglas et al. 1998), the Chocó low-level jet (Poveda and Mesa 2000), and the Caribbean low-level jet (Amador 1998, 2008). The former is parallel to the Gulf axis and has a well defined diurnal and synoptic scale variability. It is located below 2000 m at the northern end of the Gulf of California with wind velocities of $5 - 7 \mathrm{ms}^{-1}$. It is a relevant mechanism for moisture transport into the interior of the continental regions during the monsoon (Douglas and Leal 2003).

The second mechanism, the Chocó jet (CJ) develops in the western coast of Colombia near 5° N. It reaches its maximum by October-November, then decreases its intensity until being almost absent during the period February-March. Low-level warm air and moisture convergence associated with the CJ, low surface pressure and orographic vertical motion on the western Andes, contribute to deep convective activity, which is organized as mesoscale convective complexes.

The third low-level flow, the CLLJ, develops during May-June, and reaches its maximum in July and weakens in September (Amador 1998; Amador et al. 2003; Amador 2008). It is barotropically unstable, and has potential interaction with transients, such as easterly waves. From May till July, easterly waves loose energy and momentum strengthening the mean current and causing the low-level jet to peak in July. The CLLJ is an important element to explain the convective activity during July through November, and contribute to understand the climate of the region. Furthermore, it is also an important mechanisms for moisture transport from the Caribbean sea to Central America and the Gulf of Mexico during summer and winter (Durán-Quesada et al. 2010; Gimeno et al. 2012). Since one goal of this essay is precisely to examine the current knowledge about the CLLJ, more details about this jet will be described in Section 4.

According to Amador et al. (2006), the wind stress curl has a strong seasonal cycle with a marked meridional migration of the upwelling (positive curl) and downwelling (negative curl) areas that is mostly associated with the southeast and northeast trade patterns and its north–south migration in the eastern tropical Pacific. The distribution of this variable is almost zonal, and that most of the large-scale features are located north of the equator. In general, the fluctuations in the magnitude of the wind stress curl in the tropics are mainly related to seasonal atmospheric circulation and the ITCZ.

2.2. Moisture

Durán-Quesada et al. (2010) have employed the ERA-40 reanalysis data and the Lagrangian method to trace particle trajectories, in order to identify the main moisture sources for Central America. They found two oceanic sources. The first one and the most important is located in the Caribbean. The second appears to exist near the equatorial Pacific region. These results were highlighted also by Gimeno et al. (2012) whose stress that the major amount of humidity in the Continental part of Central America is originated in the nearby oceans. A third source of moisture was found by Durán-Quesada

(2012) over Venezuela due to water recycling processes (Fig. 3).

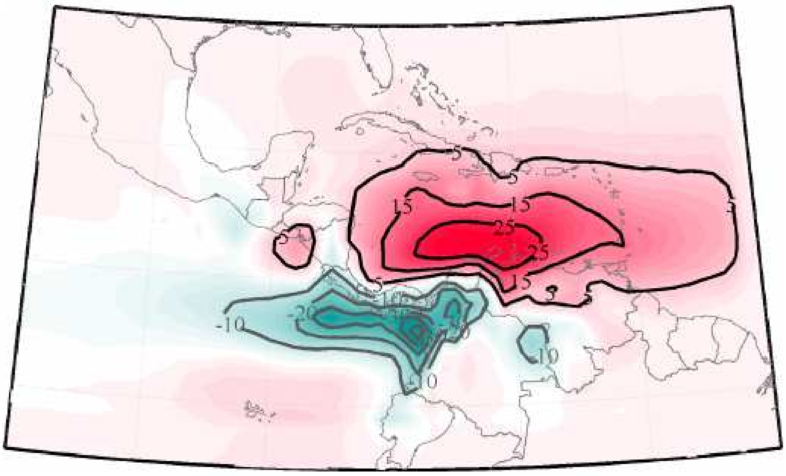

FIG. 3. Long term mean of the six backward days integrated conditional net fresh water flux (Evaporation − Precipitation)$^{-6}$ field in mm/day starting in −10 mm/day. Sources of moisture for precipitation over Central America are indicated by positive contours. The main climatological source associate with precipitation over Central America is the Caribbean Sea (Durán-Quesada 2012).

The intensity and extent of the moisture sources vary throughout the year (Fig. 4). The Caribbean source (CS) does not vary significantly along the year, except a slight displacement towards the Gulf of Mexico during winter. In contrast, the Pacific source (PS) shows significant variation throughout the year, and disappears as a source during winter and spring, mainly due to the influence of the ITCZ (Durán-Quesada et al. 2010). Seasonally, during the boreal summer the humidity transport from the CS is more effective and contributes with the precipitation in Central America, while the moisture that departs from PS is not even able to reach the entire Central America region, contributing only to precipitation in its southernmost portion, specifically at Costa Rica. Through the boreal winter a reduction in the moisture transport consistently with the precipitation pattern over this region, exhibiting dry conditions (mainly in the Pacific basin) during this season. Moisture convergence in Central America during winter is clearly less important than during spring (March, April, May) and autumn (September, October, November), when the moisture flux over the continental area of Central America becomes more relevant.

In order to understand the role of the CLLJ for moisture transport, it is worth to say that the CLLJ acts not only as a moisture belt, but also as a humidity collector that is capable of modulating surface evaporation as a result of its moisture content (Wang et al. 2007). Many studies found that the core of CLLJ wind jet is consistent with the maximum nucleus of moisture gain over the Caribbean Sea (Wang 2007; Wang and Lee 2007; Wang et al. 2007; Durán-Quesada et al. 2010). The major contribution occurs during boreal summer, and for the case of the CS region this is in good agreement with the maximum observed winds in the core of the CLLJ. However, the second maximum of the CLLJ in boreal winter is not associated with any important transport, mainly due to the incidence of the dry season, which is characterized by less intense jet winds than in summer and a minimal amount of precipitable water.

Durán-Quesada et al. (2010) highlight that the contribution of moisture to Central America that originates in the PS region is partly determined by the presence of the CJ,

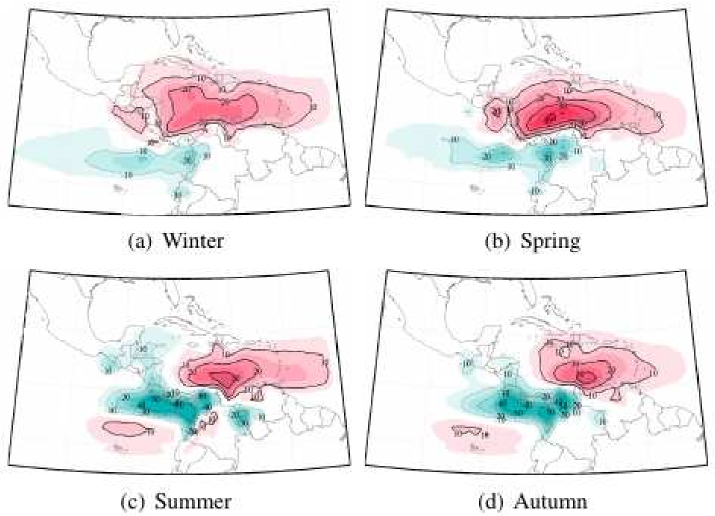

FIG. 4. As in Fig. 2, but for each season corresponding to the northern hemisphere. The Caribbean Sea is a permanent source of moisutre with intensity fluctuations while on average a source is identified over the ETPac region (Durán-Quesada 2012).

which in turn allows the development of deep convection in the region. This contribution is more noticeable over northern Colombia when it appears to be combined with the effect of orographic lifting as described by Poveda and Mesa (2000). The importance of the PS region is greatest during those parts of the boreal summer and autumn that coincide with the maximum velocities within the core of the CJ. A significant part of the moisture transported by this jet is unable to reach Central America completely, mainly as a result of the loss of moisture in the ITCZ and the presence of a mountain range in Costa Rica.

2.3. Precipitation

According to Taylor and Alfaro (2005) the most relevant synoptic influence present in the Caribbean and Central America area is the subtropical high located in the north Atlantic Ocean. The latter produces strong easterly trades winds towards the equator, being in Central America the dominant wind regime. The interaction of this trades with the topography, and the location of Central America (i.e. high mountains across the isthmus, which in time is a narrow land bridge surrounded by two oceans) imprint upon the region particular climate and weather features. The influence of the former on climate and weather, is not only manifested in the interaction with trade winds, which produces two regional climates – Pacific and Caribbean (Taylor and Alfaro 2005; Amador et al. 2006), but also this complex orography is capable to produce very local climate and weather patterns, noted mainly in the high variability observed in precipitation with respect to the altitude (Fernández et al. 1996; Amador et al. 2003).

As mentioned above, according to this wind regime, one finds the Pacific and the Caribbean coasts at lee and windward side, respectively. These climate regions show a noticeable contrast in the annual cycle of precipitation (Taylor and Alfaro 2005; Amador et al. 2006).

The Pacific coast presents a bimodal rainfall distribution around the year (Fig. 5). A first maximum occurs in May-June, due to the migration north of the ITCZ. However, it should be noted that the migration of the ITCZ cannot explain the generalized deep convection during the rainy season over the whole region since the ITCZ is not found at latitudes 10°–12° N (Alfaro 2000), neither explain the beginning and the ending of the rainy spell. Additionally to this migration, the SST of the neighboring seas have warmed reaching about 29° C, deep convection activity is developed along with a subtropical lower-tropospheric cyclonic circulation anomaly over the subtropics. Then, a relative minimum of precipitation through July till August upon this slope, is due to the MSD (Magaña et al. 1999; Karnauskas et al. 2013). During the occurrence of the MSD in these months, the convective activity diminishes, due to a decrease of about 1° C in the SST, the cyclonic circulation anomaly weakens, corresponding to an anticyclonic acceleration of the low-level flow and, therefore, to an intensification of the trade winds. This leads to a formation of divergence anomalies that inhibit deep convection activity, and the strengthening of the easterlies, forcing upward motion and intense precipitation over the Caribbean side, and subsidence and clear skies upon the Pacific slope.

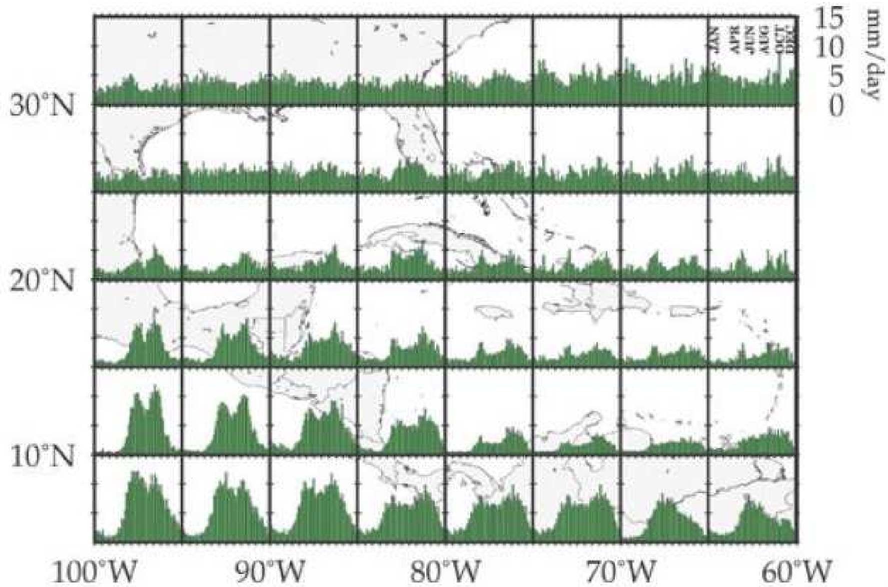

FIG. 5. Distribution of climatological *5-day* mean precipitation rates (mm day^{-1}) for contiguous 5° × 5° area. Note the bimodal distribution, especially in the ETPac, showing the MSD (Amador 2008).

The second maximum peaks during August through October due to the presence of fewer deep clouds, which would produce an increment of incoming solar radiation heating the SST above 28° C. Then, this warming in the SST produces an increase of evaporation from the oceans to the atmosphere; in addition, weakened trade winds and a low-level convergence anomaly lead to enhanced deep convection. Normally, this season presents the highest frequency of extreme events over the Pacific slope (Alfaro et al. 2010). While in the Caribbean coast, rainfall decreases during these months, owed to a decrease in the strength of the trade winds (Taylor and Alfaro 2005; Amador et al. 2006). From December to March, the Pacific slope of Central America presents warm and mostly dry conditions. During these moths, the ITCZ is at its southernmost position (Srinivasan and Smith 1996).

It is worth to mention, then, that the Caribbean side shows a different rainfall mode, as pointed out by Taylor and Alfaro (2005), in which is very difficult to define a dry season since rainfall is nearly homogeneous between January and the middle of October. Precipitation during the winter months along this coast, is mostly related to humidity convergence, mid-latitude air intrusions (Schultz et al. 1997, 1998), and less frequent low-level cloud systems traveling from the east (Velazquez 2000). During July there is no rainfall minimum, instead a relative maximum exists, which might result from the interaction between topography and the CLLJ (Amador 1998, 2008).

3. Variability elements in the Intra-Americas Seas

3.1. Large-scale variability modulators

El Niño Southern Oscillation (ENSO) is the most influential large scale index on the climate of the IAS (Wang and Fiedler 2006). Nevertheless, the influence of other variability modes in the Atlantic Ocean such as the Atlantic Multi-decadal Oscillation (AMO), and the North Atlantic Oscillation (NAO) have shown to have an important contribution in the variability of the precipitation field of Central America (Enfield and Alfaro 1999; Alfaro 2007; among others). Recently, the influence of the Pacific Decadal Oscillation (PDO) on rainfall over the Central America (Mantua and Hare 2002; Fallas-López and Alfaro 2012), the influence of the Madden-Julian Oscillation (Martin and Schumacher 2011b) on precipitation in the Caribbean islands have also been studied.

3.1.1. El Niño Southern Oscillation

The ENSO is a two-component physical mechanism that describes the coupling of SST anomalies in the tropical Pacific Ocean and the fluctuation in tropical SLP gradient in the Western and Eastern Pacific Hemispheres, also known as the Southern Oscillation (SO). El Niño is associated with unusual strong warming events that occur every two to seven years in concert with basin-scale tropical Pacific anomalies (Wang and Fiedler 2006). ENSO variability in the eastern tropical Pacific is centered along the equator, but is closely related to variability of the tropical WHWP. Authors such as Chen and Taylor (2002) and Bell and Chelliah (2006) suggest that ENSO forces variations in the SST field and vertical wind shear that trigger the inter-annual variability of the hurricane season during ENSO events. Moreover, an increase (decrease) of precipitation has been associated with cold (warm) ENSO phases (Dai and Wigley 2000; Giannini et al. 2000).

Variability of the surface winds has been also observed related to ENSO. The flow over surface has been found to increase during El Niño events while a reduction occurs in the opposite phase. Examples of that, the changes in the jet core of the Chocó LLJ (Poveda and Mesa 2000) and the CLLJ (Wang 2007; Amador 2008). The core intensity of the CLLJ varies with ENSO phases in such way that during warm (cold) events the jet core is stronger (weaker) than normal in the boreal summer, surface wind stress and wind stress curl area expected to be stronger (weaker) than normal in the easternmost portion of the eastern tropical Pacific. Contrary to what happens in summers, the jet core is weaker (stronger) than normal during warm (cold) ENSO phases in winter (Amador 2008). These variations in the wind field influence the vertical wind shear and cyclone-genetic processes (Amador 2008; Amador et al. 2010).

It can be followed that ENSO is able to affect precipitation distribution by modulating two different mechanisms: a) evaporation variability linked to SST and surface drag variations and b) transport of moisture due to wind flow modulation.

3.1.2. The Madden-Julian Oscillation

The Madden-Julian Oscillation (MJO) has been identified as a 40-50 days oscillation in zonal wind anomalies in the tropical Pacific (Madden and Julian 1994). Zhang (2005) makes a review of the basics of the Madden-Julian Oscillation (MJO). Recent studies have analyzed both the ocean-MJO relationship (Webber et al. 2010), and the interaction between the MJO and ENSO (Tang and Yu 2008; Moon et al. 2010). Some findings have shown that this oscillation has effects on the tropical troposphere and strong impact on tropical convection. The MJO-related zonal wind anomalies in the ETPac region might be associated with the increase of rainfall in the North America Monsoon region. Also, the amplification of easterly waves as a trigger of gulf surges development that may be related with the variability of the moisture sources that feed the North America Monsoon System (Lorenz and Hartmann 2006). These authors also suggested that westerly phase of the MJO are associated with the enhancement of favorable conditions for Mesoscale Convective System (MCS) development in the region. Barlow and Salstein (2006) found a positive (negative) phases of the MJO are associated with the increase (decrease) of precipitation in the region. Recently, Martin and Schumacher (2011b) showed evidence about a link between the intensity of the CLLJ and the MJO, and it is suggested to lead changes in precipitation. Both of these last studies find some relation between the occurrence of extreme events and the MJO phase.

3.1.3. The Pacific Decadal Oscillation

The Pacific Decadal Oscillation (PDO) is a long-lived El Niño-like pattern of the variability of the Pacific climate (Zhang et al. 1997). Mantua et al. (1997) highlight the PDO as the dominant pattern of Pacific Decadal Variability (PDV). Even when the PDO is referred as an El Niño-like pattern, it differs from ENSO in the time scale of the persistence of events and its fingerprint is more noticeable in the extratropics rather than the tropics (Mantua and Hare 2002). The pattern of a warm PDO phase is featured by cooler than normal SSTs in the central North Pacific and warmer than normal SSTs along the west coast of the Americas. Symmetry in SST anomalies patterns between northern and southern hemispheres exhibited by the PDO has been noted by Evans et al. (2001). The mechanisms of the PDO are complex and still and open issue, however some studies suggest the importance of the tropical coupling for the existence of the PDO (Feng et al. 2010). Schneider and Cornuelle (2005), propose the PDO to evolve from a composition between the forcing due to El Niño 3.4 (see Trenberth 1997 or for definition of El Niño regions), and the changes of the Aleutian low (interannual frequencies) and the Kuroshio-Oyashio Extension (decadal time scales). The importance that the PDO may have for global climate is related to a correlation between the PDO index (Mantua et al. 1997) and precipitation anomalies. Warm PDO phases have been found associated with anomalously dry periods in the eastern coasts of Eurasia, Northwest Pacific of USA, Central America and northern South America. While the same phase seems to be related to wetter than normal conditions in the Gulf of Alaska, South-west USA and Mexico, South-east Brazil, South central South America and western Australia. Lately, Fallas-López and Alfaro (2012) have studied the influence of the PDO in combination with other variability modes (i.e. ENSO, NAO, AMO) on the precipitation field in Central America.

3.1.4. The North Atlantic Oscillation

Greatbatch (2000) made a detail revision of the North Atlantic Oscillation (NAO). It is the most important variability mode in the North Atlantic Ocean. One way to define and index for NAO is like the difference between normalized mean winter (December to March) SLP anomalies at Lisbon, Portugal and Stykkisholmur, Iceland (Hurrell 1996). NAO is an element of the Arctic Oscillation (AO) patterns, then the former is linked with SST anomalies in the East/South-east of Greenland.

NAO has been also associated with the strengthen of the NASH and the North Eastern trade winds, affecting thus the circulation in the IAS. Thereby, the impact of the NAO is associated with the climate features of the IAS, SST in the Tropical North Atlantic, the size of the WHWP and the CLLJ. Seasonal variations of the NAO may be reflected in its influence on the IAS easterly winds and precipitation patterns (Wang 2007). Malmgren et al. (1998) found that during the boreal summer the NAO index has an inverse relation with the observed precipitation patterns over Puerto Rico.

3.1.5. The Atlantic Multi-Decadal Oscillation

The Atlantic Multi-Decadal Oscillation (AMO) is a 70 years period mode in SST (Delworth and Mann 2000; Kerr 2000). This signal has also been found in different modeling analysis (Delworth et al. 1993; Delworth and Greatbatch 2000; Latif et al. 2004) and it was detected as the first rotated EOF with a large response in the North Atlantic SST (Mestas-Nuñez and Enfield 2001). The AMO signal modulates precipitation variations in different regions such as the IAS (Giannini et al. 2000). Wetter (drier) conditions over Central America (north-east Brazil) during JJA (DJF) were found by Zhang and Delworth (2006). AMO has been also determined to be of importance modulating the impact of ENSO on drought. Connection between AMO and other regional features such as the Atlantic Warm Pool (AWP) has been also studied. Wang et al. (2008) show that warm (cool) phases of the AMO are associated with repeated large (small) AWPs, suggesting the relationship between the AMO and Atlantic tropical cyclones. The latter is in agreement with results that indicate the presence of multidecadal variations in hurricane activity due to the Atlantic SST (Goldenberg et al. 2001). AMO is then of importance related to the low frequency variability of precipitation as it modulates the distribution of moisture and extreme rainfall events (Maldonado et al. 2013).

3.2. Regional-scale modulators

3.2.1. Regional circulations systems

Two important regional circulation systems modulate the weather and climate in the region. The first mechanism is the CLLJ over the Caribbean Sea, and the second the Chocó LLJ.

As previously mentioned, the CLLJ is important for moisture transport in the region, and also to explain the convective activity during the summer months. During the winter, it has a second maximum but has been found related to convection in this period (Amador et al. 2006; Amador 2008). More details on this jet will be discussed later in Section 4. On the other hand, the Chocó LLJ peaks in October-November, and contributes to the moisture transport for the southernmost part of Central America (Durán-Quesada et al. 2010). Furthermore, the Chocó jet is associated with deep convection activity over the western Andes region in Colombia (Poveda and Mesa 2000).

3.2.2. SST of neighboring oceans

The influence of the SST anomalies in the precipitation variability field has been widely studied (Enfield and Alfaro 1999; Alfaro 2000, 2007). According to these authors the beginning and ending of the rain spells are related to fluctuations in the SST of the Atlantic and Pacific Oceans, and, these anomalies are related to the magnitude of rainfall and frequency of rainy days (Maldonado and Alfaro 2010a, 2011; Maldonado et al. 2013). Amador et al. (2006) pointed out that the seasonal cycle of SST is important in defining key climatological features, especially during summer-autumn, such as the Western Hemisphere warm pool (WHWP) development (Wang and Enfield 2001; Wang and Fiedler 2006), the appearance of MSD (Magaña et al. 1999; Karnauskas et al. 2013), and favorable areas for cyclogenesis (Goldenberg et al. 2001). During the northern winter, SST isotherms over the Caribbean and the eastern tropical Pacific are mostly zonally distributed, with values usually below $28 - 29°$ C, except in the central eastern tropical Pacific, and to the west of Central America, where there is a maximum of SST all year. As result of this, and a relatively strong vertical trade wind shear, and reduced evaporation, no major convective activity occurs in most of the Pacific coast of Central America during this season. Also, during boreal winter, the ITCZ is at its southernmost position (Srinivasan and Smith 1996).

During boreal summer, a large warm pool dominates the SST distribution over most of the eastern tropical Pacific region (Magaña et al. 1999; Wang and Enfield 2001, 2003). In the Caribbean warm pool, organized activity is barely observed, due mainly to strong vertical wind shear and strong subsidence associated with regional scale circulations, such as those associated with the low-level jet described above.

As cited before, during warm (cold) ENSO phases, the CLLJ shows stronger (weaker) than normal wind speeds (Amador et al. 2003, 2006; Amador 2008). This fluctuation is reflected in SST anomalies over the Caribbean Sea, north coast of Venezuela; a strong (weak) jet results in negative (positive) SST anomalies over this region due to strong (weak) Ekman transport. In this way, the jet may have a role in coupling SST anomalies in eastern Pacific during El Niño or La Niña events with anomalies over some regions of the Caribbean during summer. Variations in surface variables (precipitation and temperature) in different sectors of Mesoamerica, including its west coast, are the result of a combination of fluctuations in the equatorial tropical Pacific and in the tropical north Atlantic (Amador et al. 2006). Studies such as Alfaro et al. (1998); Alfaro and Cid (1999a,b); Enfield and Alfaro (1999) show that the strongest rainfall signal occurs when tropical north Atlantic and tropical Pacific SST anomalies are in a configuration of meridional dipole (antisymmetric) across the ITCZ, that is, when this anomalies have an opposite sign. The rainy season in south Central America tends to start early and end late in years that begin with warm SST in the tropical North Atlantic. Ending dates are also delayed when the eastern tropical Pacific is cool.

3.3. Local-scale modulators

Topography is the main local modulator of the variability in the region. Interaction between the terrain with the induced flow coming from the Pacific ITCZ, due to of some prevailing synoptic system, produces a type of disturbance that contributes to precipitation in Central America, which is named the “temporales” (Hanstenrath 1991; Fernández et al. 1996). They are periods of weak-moderate nearly continuous rain, lasting several days and affecting a relatively large region. Their definition includes the condition that the wind must be weak; however, Amador et al. (2003) have shown that in some cases winds can be intense and long-lasting.

Fernández et al. (1996) identifies at least four synoptic conditions that can eventually generate conditions for a temporal, such as a deep lower and middle troposphere troughs in the easterlies (no hurricanes included), intrusions of an upper troposphere troughs in mid-latitude westerlies, outbreaks of cold air from North America, and the direct and indirect effect of hurricanes. Velásquez (2000) also found that westward- traveling, low-level cloud system over the Caribbean reaching the Pacific, which are not necessarily associated with mid-latitude cold air intrusions. The frequency of these events presents a great deal of interannual an intraseasonal variability, and their relationship to ENSO or to other large-scale climatic signal is still unclear (Amador et al. 2006).

The sea-breeze circulations are other relevant regional modulators of the climate, in island and peninsulas since it favors the development of convective system. Marked by a diurnal cycle due to the thermal contrast between the coastline and the sea, it can be related to a diurnal cycle of precipitation. However, in coastal regions with nearby complex topographical, like Central America, the induced flow can impact the temporal a spatial distribution of the meso-scale features in rainfall. Besides that, there is evidence in regions with similar characteristics to the Central American isthmus (i.e. west coast of Colombia), that the sea breeze can penetrate over near-coastal mountains into a valley until approximately 100 km (Warner et al. 2003). Nevertheless, note that such a feature of the sea breeze circulation has not been reported in Central America yet, but it might be a local modulator on many regions of this area.

4. The Caribbean low-level jet

4.1. General remarks on low-level jets and climate

A low-level jet (LLJ) can be defined as a wind speeds maximum that occurs in planetary boundary layer or in the lower troposphere. Such a type of wind flow has been observed in many regions in the world, in a wide variety of large-scale environments and in all season (Stensrud 1996). The Great Plains LLJ (Hoecker 1963; Bonner 1968), the Caribbean LLJ (Amador 1998; Wang 2007; Amador 2008), the Chocó LLJ Poveda and Mesa (2000), and the South American LLJ (Vera et al. 2006a), in the Americas just to mention some jets, that have somehow relation with the climate in the IAS, or specifically related with the American Monsoon System. In his work, Stensrud (1996) points out that in midlatitudes, the LLJs are more frequent during summer than in winter, and tend to develop over night with a marked diurnal component, being the maximum near the dawn.

For global and regional climate LLJs are an important circulation mechanisms to provide the moisture transport and also are related with convective activity. In modeling, LLJs are a relevant meso-scale weather element in regional and global spatial scales and seasonal timescales. Deep convection and its effects on soil moisture influences strongly the uncertainties present in the models due to cloud-radiation interactions and surface energy budget. Also LLJs influences the global climate by the transport and melting of the ice-packs in the Antarctic Peninsula during winter. It has been shown that low-level jets associated with drainage in the Antarctic appear to be essential in prescribing the larger-scale circulations near the South-Pole.

There exist many physical mechanisms by which a LLJ forms. Inertial oscillation, shallow baroclinicty, and terrain effects are among the mechanisms to explain the origin of the LLJs. See Stensrud (1996) for details.

Inertial Oscillation: during the day the planetary boundary layer is coupled with the surface layer by mean of the frictional forces (turbulence mixing) causing the wind to be subgeostrophic. By the end of the day the frictional force decreases significantly, and the wind above the the shallow nocturnal inversion, and within the residual layer from the PBL formed earlier in the day, are decoupled from the surface layer, and are not longer in balance. This imbalance between the Coriolis force and the pressure gradient produces a inertial oscillation of the wind. This mechanism is used to explain the nocturnal low-level jet. It has been document a similar behavior in a marine boundary layer, in which warm air coming from inland reaches sea waters with cooler surface temperature, producing the same thermal stratification (Källstrand 1998).

The shallow baroclinicity: this feature is present in regions with differences in the horizontal distribution of sensible and latent heat, for instance the coastal zones. These differences produce strong low-level baroclinity within the PBL. These regions of shallow baroclinicity produce a LLJ through strong geostrophic forcing. This type of LLJ can be relatively constant through the day if the gradient temperature between the surfaces is nearly constant (i.e. sea-ice boundaries or gradients in sea surface temperature) or can have diurnal changes if the fluxes over the surfaces have a diurnal component (i.e. sea-land boundaries). Also, LLJ can occur due to baroclinicity produced by sloping terrain, regions of development and evolution of extratropical cyclones, and by the presence of a mountain range blocking the low-level flow of a cold, stable air mass and channelling this air mass along the mountain slopes.

Terrain Effects: In the case of very complex terrain the diurnal heating can produce both slope and valley wind that can develop into a LLJ. Furthermore, terrain features can create boundary currents, as in the case of the GPLLJ, and the Somali LLJ.

Isallobaric forcing: Some low-level streams appear to develop in association with synoptic-scale forcing, have a minimal diurnal oscillation, and often extend above the PBL depth. It has been found that this kind of low-level stream can be intensified due to an increase of the isallobaric wind.

Vertical parcel displacement: it has been reported that wind speeds in excess 30 ms^{-1} develops in response to the vertical displacement of air parcels within a baroclinic enviroment. As parcels approach the developing cyclone from the northeast, a change in the pressure gradient force is experienced. While this change is modest in the horizontal direction, it is quite large in the vertical as parcels move through the baroclinic region associated with the coastal front. As parcels are displaced vertically, a rapid increase in the ageostrophic wind component occurs that subsequently leads to acceleration of the parcel and the rapid development of a LLJ. This kind of processes are manifested within the transverse ageostrophic circulations associated with an upper-level jet streak as the jet is modified by diabatic processes and frontogenesis within a strongly baroclinic environment.

It should be noted that none of this mechanisms alone can explain the observations, but instead many authors have used a combinations of those forcing to describe the origin of some observed LLJs cases.

4.2. Features of the Caribbean low-level jet

The Caribbean low-level jet (CLLJ), also known as the Intra-Americas low-level jet (IALLJ) is located in the Western Caribbean Sea. It has a semi-annual cycle with two maxima, in summer and winter (Fig. 6). The jet axis is usually found at 15°N, and between 70°W and 80°W, and the wind maximum peaks near the 925 hPa (Amador 1998, 2008). The CLLJ develops during May-June, and reaches its first maximum in July and weakens in September (Amador 1998; Amador et al. 2003). A second wind speed maximum is also reached during the boreal winter in February (Amador et al. 2006; Wang 2007; Whyte et al. 2008; Amador 2008; Muñoz et al. 2008).

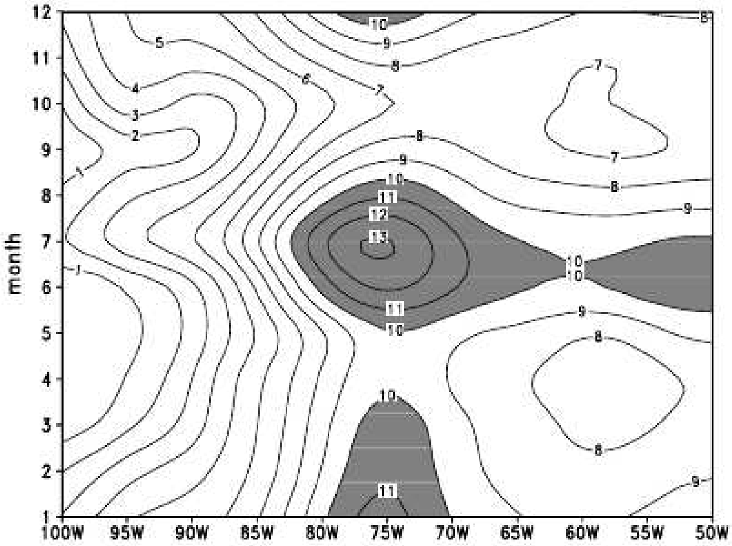

FIG. 6. Time-longitude cross section of the monthly mean wind speed (ms^{-1}) at 925 hPa averaged from 12.5 to 17.5° N from NCEP/NCAR reanalysis (Amador 2008).

Figure 7 shows the vertical profile of monthly wind speed for February and July. During summer, the CLLJ is barotropically unstable, so, disturbances can grow, and hence it has a potential interaction with transients, such as easterly waves (Amador 1998). From May till July, easterly waves loose energy and momentum strengthening the mean current and causing the low-level jet to peak in July. From September until early November trades winds are relatively weak, vertical wind shear over the Caribbean is reduced, hurricane activity peaks, and rainfall spreads almost all over the IAS.

According to Amador (2008), the winter component of the jet appears to be compressed below the 850 hPa with values up to 10 ms^{-1}, with a strong vertical wind shear, an element unfavorable for convection. These features makes the jet an important element to understand the convective activity during the second half of the year and contribute to the understanding of the regional climate but are not applicable for February when no major convective activity is observed. The intensification of the trades during boreal winter and the corresponding peak of the CLLJ in February may be from the strengthening of the meridional pressure gradient in the Atlantic and to the thermal contrast between the Caribbean Sea and the SAMS.

Wang (2007) found that during the jet maximum in summertime there is an associated SLP maximum. The semi-annual cycle of the CLLJ also coincides with both the semi-annual variations in the meridional gradient of SST, and SLP. Besides that, it is also highlighted that the CLLJ anomalies vary with the Caribbean SLP anomalies that are connected to the variation of the NASH. Wang et al. (2007) found that the SSTs of the Gulf of Mexico, the Caribbean Sea, and the western tropical North Atlantic (also known as the Atlantic Warm Pool, AWP) can modulate the intensity of the jet core through

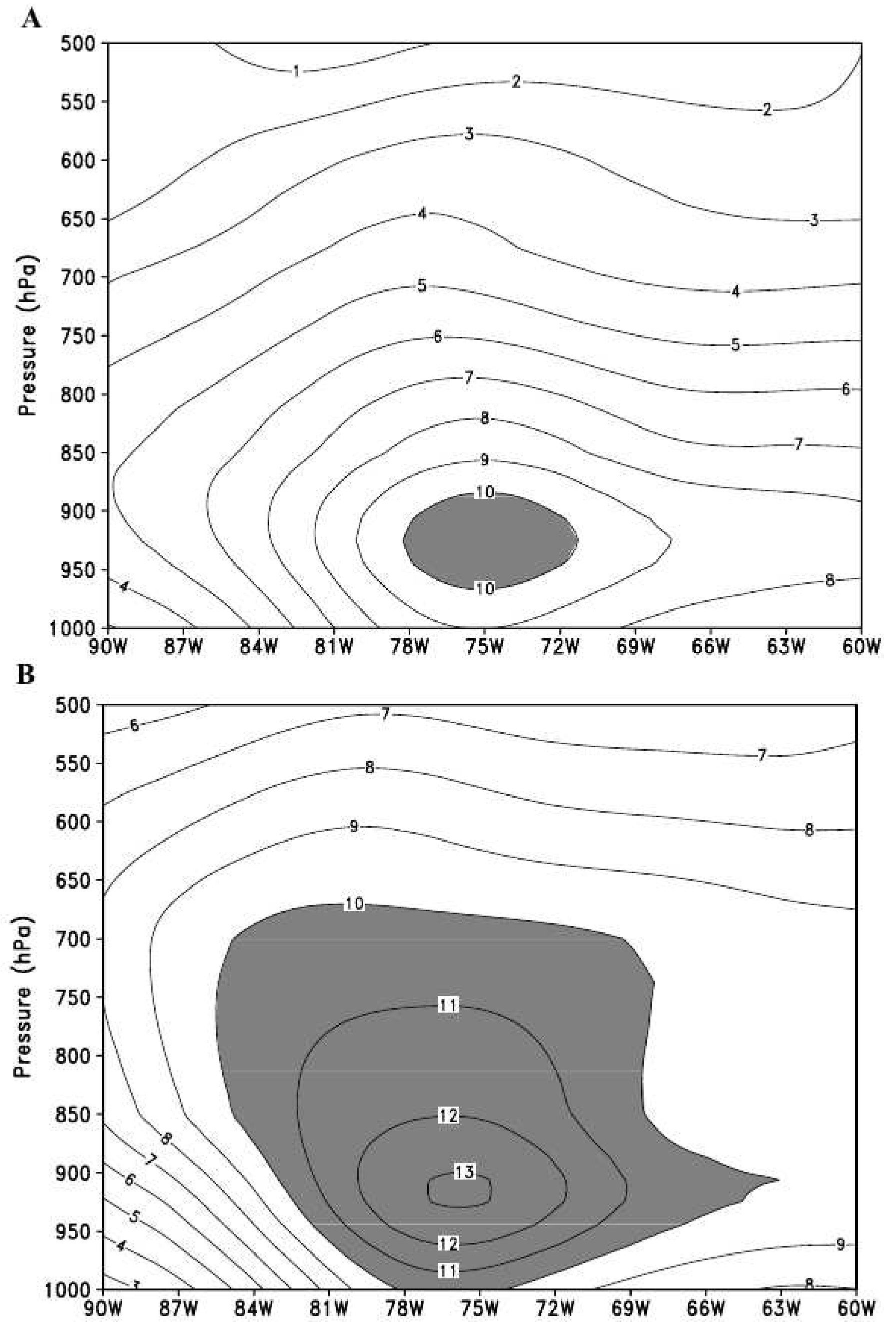

Fig. 7. Vertical profile of monthly wind speed (ms^{-1}) averaged from 12.5°N to 17.5°N from NCEP/NCAR reanalysis for (A) February and (B) for July. Taken from Amador (2008).

pressure changes. The AWP tend to weaken the summertime NASH and to strengthen the continental low in the NAM region.

Another striking difference between the summer and winter components of the CLLJ was noted by Amador (2008) and Muñoz et al. (2008). Figure 8 shows the February and July climatology of the horizontal wind field at 925 hPa (base period 1979 − 2001) from the ERA-40 reanalysis (Uppala et al. 2005). During the northern summer the CLLJ splits into two branches, one towards the Gulf of Mexico, and the other goes through Central America and continues to the Pacific Ocean. Whereas in winter, the CLLJ turns south, at the entrance of the South American subcontinent and after crossing the Central American land bridge. Amador (2008) points out that the trades and so the CLLJ are responding to thermal contrasts during the summer season of the corresponding subcontinent, and it is a natural component of the American monsoon system (AMS).

At diurnal scale, Muñoz et al. (2008) and Cook and Vizy (2010) noted that during July and February, the 925 hPa easterly wind peaks at 0300 (2300 LT) and 1200 (0800 LT) UTC, with average winds of 15.1 ms^{-1} in July and 14.4 ms^{-1} in February. These two maxima (at 0300 and 1200 UTC) are divided by a relative minimum that extends from 0600 (0200 LT) to 0900 (0500 LT) UTC. The absolute minimum is at 2100 UTC (1700 LT), with an easterly wind of 12.5 ms^{-1} in July and in February, and represents a semidiurnal variability (Muñoz et al. 2008). Cook and Vizy (2010) suggested that the semidiurnal cycle (with minima at about 0400 and 1600 LT) is caused by semidiurnal cycling of the meridional geopotential height gradient in association with changes in the westward extension of the NASH. A diurnal cycle is superimposed, associated with a meridional land–sea breeze (solenoidal circulation) onto the north coast of South America, so that the weakest jet velocities occur at 1600 LT.

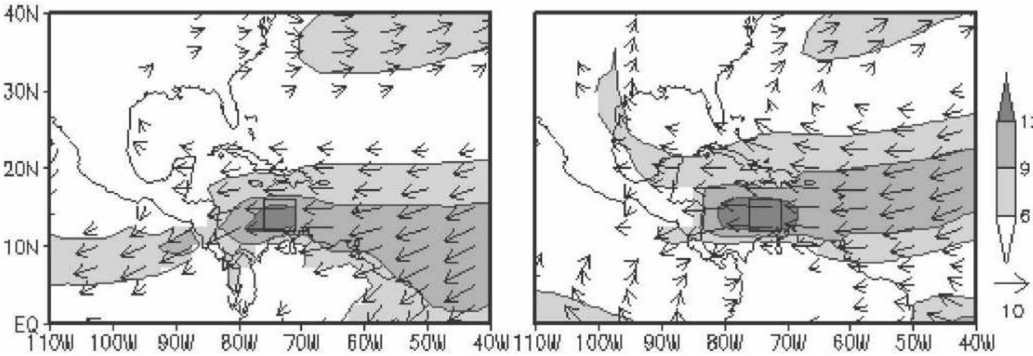

Fig. 8. February (left) and July (right) climatology of the horizontal wind field at 925 hPa. The base period is from 1979 - 2001. The reference arrow is 10 ms^{-1}. The shadings correspond to the magnitude of the vectors (ms^{-1}). Vector with wind speed with less than 3 ms^{-1} are not shown. The rectangle delineates the area of the maximum wind speed from 12° to 16° N and from 71° to 76° W. Data from ERA-40 reanalysis (Muñoz et al. 2008).

For regional climate, the IALLJ influences strongly the precipitation over the IAS. During summer at the exit of the jet (located at the eastern side of Nicaragua and Costa Rica), there is large-scale convergence of the wind field, vertical movement is developed which help to explain part of the rainfall during July over flat terrain in that area, where a maximum of precipitation is observed. In the entrance zone, the low-level flow diverges and associated coastal upwelling produces cooling of the waters, scarce precipitation is detected and dry conditions over northern coast of Venezuela and Netherlands Antilles are found (Amador 1998, 2008). In the boreal winter the situation is different, the SSTs are 1 - 2° C colder than in summer, and the CLLJ has a more defined vertical structure, generating conditions by which convection is inhibited, being this season drier than summer in the Caribbean.

Moreover, either the summer or the winter branch of the jet contributes with the vapor water transport to each subcontinent. In the first case, together with the GPLLJ helps to the humidity transport that supplies moisture in the central United States, meanwhile during the austral summer supplies moisture to rainfall associated with the SAMS. Durán-Quesada et al. (2010); Durán-Quesada (2012); Gimeno et al. (2012) found that, precisely, the CLLJ is the main element in moisture transport from the Caribbean sea to Central America, the Gulf of Mexico and the northern South America.

Amador (2008) has also documented some particular things happening in the boundary layer during the summer and winter components. For instance, as the jet passes through the mountain gaps of Central America and reaches the Pacific ocean, it seems to be accelerated. In Managua, Nicaragua, the jet can be very strong with mean values greater than 16 ms^{-1}. The height of the wind maxima is also apparent to increase from the western Caribbean through the Pacific, although there is a lot of vertical fluctuation in the jet core (See Figs. 6 and 7 in Amador 2008). Amador (2008) also reported another striking feature within the boundary layer in summer. From direct observations just below the jet core, he noted that the fluxes of latent heat increase compared to the surroundings, despite that the SSTs just below the jet core are cooler than the nearby waters (See Figs. 17 and 21 in Amador 2008). He said that this behavior could be expected since the evaporation beneath the jet axis increase due to the strong wind. This fact can be also associated with humidity transport done by the CLLJ towards the north (Gulf of Mexico) and to the east (Central America). the latter was also pointed out by Wang et al. (2007).

The relationship with the ENSO has been highlighted by Amador (1998); Amador et al. (2006); Wang (2007); Amador (2008). During a warm (cold) ENSO episode, the jet core has shown to be strong (weak) than normal. The latter also has been with precipitation anomalies over the IAS, being positive (negative) in the western Caribbean near Central America, and negative (positive) in the central IAS region. Furthermore, it has been suggested that the CLLJ is an important dynamical mechanisms coupling the SST in the Pacific Ocean with SST in the Caribbean, plus it acts as a modulator of the hurricane seasons since it is as the main controlling mechanism of the tropospheric vertical wind shear. For instance, Wang and Lee (2007) showed that a weaken easterly wind CLLJ, together with the AWP-induced chage of the upper-level wind, reduces the tropospheric vertical wind shear that favors hurricane formation and intensification during August-October. Amador et al. (2010) noticed that for warm ENSO events, low-level wind increases at the jet core strengthening the low level convergence near Central America at the jet exit, and low-level divergence in the central Caribbean at the jet entrance. The descending motion associated with the latter acts as an inhibiting factor for convection and tropical cyclones development.

Other large-scale indexes such as NAO and MJO have been studied and its link with the variability of the CLLJ. Wang (2007) also mentions that the CLLJ varies in phase with NAO since a strong (weak) NASH is associated with a strengthening (weakening) of both CLLJ and NAO.

Martin and Schumacher (2011b) found that intraseasonal variability of rainfall is in phase with MJO. They mention that anomalies in precipitation are also related with anomalies in the wind field at 925 hPa, and more evident in the region where the CLLJ is prominent. The magnitude of the jet varies in phases with MJO.

5. Methods to study climate

5.1. Climate data sets

5.1.1. General Circulation Models

Atmospheric general circulation models (AGCMs) are numerical models in which the average circulation is computed for specific external forcing. Holton (2004) classify them as a boundary condition problem. Similar to large-scale weather prediction models, AGCM attempt to simulate synoptic-scale weather disturbance. Most of the time the SST is treated as specific forcing, however, it is noteworthy that in reality, there are strong interactions between the atmosphere and the ocean, such as the wind drive currents, which influence the SST distribution, which in turns influences the atmosphere. Nowadays AGCMs are couple to ocean for more reliable SSTs. These kind of models are named Atmospheric-Oceanic Circulation Models (AOGCM). AOGCMs use a geometrical computational domain of the globe employing some spatial and temporal representation of the variables (finite difference and/or spectral methods), and, mostly, including implicitly small-scale processes in the equations. Although the AOGCMs make a gross representation of the climate features for a specific region, due to is typically resolution of 100–300 km, they can generate reliable information for authorities, decision-makers and scientists (Amador and Alfaro 2009). To exemplify that, consider the models utilized by the IPPC in their assessment reports on climate change (Randall et al. 2007). Among the AGCMs one can mention the Hadley Centre Coupled Model version 3 or HadCM3,(Gordon et al. 2000; Pope et al. 2000; Reichler and Kim 2008), the Canadian Centre for Climate Modelling and Analysis or CCCma (Scinocca et al. 2008), and the Community Earth System Model or CESM (Boville and Gent 1998; Kiehl et al. 1998), among many others.

Other type of global model are the Global Numerical Weather Prediction systems (GNWPs). They used the same large-scale dynamical equations as the AGCM. However, the GNWPs are cataloged by Holton (2004) as an initial conditions problem. These models are used to make prediction for few days, in other words, they employ short time scales. The Globel Forecast System (GFS) model (Environmental Modeling Center 2003) from National Oceanic and Atmospheric Administration (NOAA), National Weather Service (NWS) and National Center for Environmental Prediction (NCEP) is an example of this models. When in a NWP real-data assimilation is implemented, the products are known as the analysis of the NWP. See for instance the Global Data Assimilation System or GDAS (Whitaker et al. 2008), which is the analysis of the GFS. It has been found that this technique can improve the forecasts.

5.1.2. The reanalysis projects and observations

A reanalysis can be described as an assimilation system, in which historical observations of the atmosphere field variables are assimilated in a global model, in order to improve and support the needs of the research and climate monitoring communities. The 40-year NCEP/NCAR reanalysis (Kalnay et al. 1996), and now its updated version the 50-year NCEP/NCAR reanalysis (Kistler et al. 2001), the ERA-40 (Uppala et al. 2005), and the ERA-interim (Dee et al. 2011) reanalysis projects are among the most often used reanalysis data products used in the scientific community. Differences among the different reanalysis products can be found owed to the incorporation of the satellite data in the reanalysis models. Trenberth et al. (2011) examine the current reanalysis models according to the representation of moisture transport and energy fluxes, important parameters in the hydrological cycle. Previously, issues in the NCEP/NCAR reanalysis have been already documented, such as substantial problems that limit its use, particularly for climate change and variability studies. In tropical zones, like Central America and/or the IAS, this reanalysis data have shown deficiencies in representing some key features of the climate for those regions. Janowiak et al. (1998) discovered that the NCEP/NCAR reanalysis is not correctly representing the migration of the ITCZ neither in the Pacific nor in the Atlantic oceans. Also, it misses to render the transitions from warm to cold phases of the El Niño South Oscillation. Amador (2008) showed that the NCEP/NCAR reanalysis is not capable of capturing the correct the vertical wind structure of the CLLJ. Additionally, reanalysis data should be taken with caution in regions with scarcity of data such as the IAS, mainly over the Caribbean Sea. Despite of that, the NCEP/NCAR reanalysis data still is a good enough for representing mean features of the atmospheric fields, and remains as the most used reanalysis product used for this region. On the other hand, few studies have been performed using the Europeans reanalysis in the IAS. Durán-Quesada et al. (2010); Durán-Quesada (2012); Gimeno et al. (2012) studied the moisture transport over the IAS using the ERA-40. Muñoz and Enfield (2011) studied the variability of the wind field over the IAS using the same reanalysis. They do not report major related issues using this reanalysis product.

5.2. Downscaling techniques

For generating information at local or regional scales (10 to 50 – 60 km, or even less), two approaches are commonly used – interpolation methods and downscaling techniques. These methods have advantages and disadvantages, which will not be discussed here, but some comparison can be found in Amador and Alfaro (2009). However, two aspects should be highlighted; first, the statistical relations are not causality relations implying less ability for description and understanding of the physics and dynamics of the interactions among the climate system elements. These statistical relations must be sustained by a conceptual physical model. Second, downscaling has the potential to generate climate scenarios taking into account regional spatial variations and the interaction among the climate system elements. The latter approach can use statistical or dynamical methods. Nowadays, both statistical and dynamical downscaling, have shown good skill for prediction of the atmospheric variables, under the same climate conditions (Gershunov et al. 2000; Wilby and Dawson 2007).

5.2.1. The dynamical downscaling

The dynamical downscaling uses the Limited Area Models (LAMs) or Regional Climate Models (RCMs) which are also known as meso-scale models. This approach generates high resolution information from the GCM outputs. RCM can employ sub-regional domains of few kilometers (10–50 km), but associated uncertainty could be expected due to increase resolution, and less understanding of physical processes. There is evidence that these models simulate better the regional climate and meteorology than GCMs (Giorgi and Mearns 1999), and, especially over mountains, important issue in the spatial domain in Central America (Amador and Alfaro 2009). Furthermore, RCMs incorporate the state-of-the-art of the physics and dynamics of the climate system, and can simulate adequately the interaction among the elements of the climate system. But, in spite of their great potential, RCMs require a high computational power, good knowledge of the climate of the region, and experience using climate models. Some of these models are widely used for climate and forecast such as the MM5 (Dudhia et al. 2005) and the WRF (Skamarock et al. 2008). Since the latter is planed to use in this research, a brief description on the basic of WRF is giving later in this section.

In Central America, lately, it has been an increase in climate modeling using LAMs. Warner et al. (2003); Hernandez et al. (2006); Amador (2008); Rivera and Amador (2009); Maldonado and Alfaro (2010a); Maldonado (2012) have investigated some features of both the climate in the IAS, and the performance of regional models estimating some climate parameters such as precipitation and wind. It has been noticed that precipitations results are very sensitive to the physical parametrization schemes for cumulus due to the inherent sensitivity of those schemes to seasonal patterns (Rivera and Amador 2009; Maldonado 2012). It was noted a good performance of the models estimating the mean field of horizontal wind, temperature, but not precipitation. Furthermore, Amador (2008) found evidence that using regional models, the representation of the vertical wind field is enhanced compared to reanalysis, but still is not in agreement with observations.

Weather, Forecast and Research model system (WRF) The meso-scale model WRF version 3 (Skamarock et al. 2008) a Numerical Weather Prediction (NWP) designed for both research and operational applications. It has been widely used in tropical regions around the world, for instance Vaidya (2007); Davis et al. (2008); Ray et al. (2011); Maldonado (2012) showed good results in precipitation estimation.

This model has some important characteristics such as: (a) a fully compressive set of equations; (b) Euler non-hydrostatic with a run-time hydrostatic option available; (c) prognosis variables such as horizontal velocity (u,v) in Cartesian coordinates, vertical velocity (w), and perturbation of potential temperature; (d) the vertical coordinate system is terrain-following, dry hydrostatic-pressure with vertical grid stretching allowed; and (e) top of the model is a constant pressure surface; (f) for horizontal grid the Arakawa C-grid staggering is used. Furthermore, it allows performing one-way interactive, two-way interactive and moving nests. For physics, it employs microphysics schemes ranging from simplified physics to sophisticated mixed-phase physics suitable for process studies and NWP. Also, it uses cumulus parametrization with adjustment and mass-flux schemes for meso-scale modeling. Multi-layer land surface models from simple thermal model to vegetation and soil moisture models are used for surface physics. Non-local K schemes are used to parametrize physics in the planet boundary layer, and longwave and shortwave schemes with multiple spectral 8 bands and a simple shortwave scheme suitable for climate and weather prediction applications.

5.2.2. The statistical downscaling

The statistical downscaling can generate information from GCM outputs to either station or regional scale and in different temporal scales – daily, monthly or seasonal scales. Any statistical downscaling scheme uses empirical relationships among the predictors (large scale variables) and the predictants (local or regional scale variables). These methods have some advantages such as they require a low computational cost, and sometimes the relations between predictor-predictant are non-stationary (Wilby and Dawson 2007), simulating better the real behavior of the climate system Amador and Alfaro (2009). However, this approach requires a good quality of observed time series for the calibration. Furthermore, the results are dependent in the selection of the predictors, and the empirical transfer functions.

The statistical methods often used for this approach belong the multivariate statistics field, e.g. the multiple linear regression, principal component analysis or empirical orthogonal functions, the canonical correlation analysis (Wilks 2005; Amador and Alfaro 2009; Navarra and Simoni 2010). Besides, there are some tools designed for climate change or climate variability studies whose employ these methods such as the Statistical Downscaling Model or SDSM, (Wilby and Dawson 2007), or the Climate Predictability Tool (CPT) developed by the International Research Institute for Climate and Society (IRI). Some studies such as Maldonado and Alfaro (2010b, 2011) and Maldonado and Alfaro (2012) have used these tools focused in climate change and climate variability in the IAS.

For the analysis of the variability of the horizontal wind field over the IAS, and so, of the CLLJ, and its influence on precipitation over central United States during spring, Muñoz and Enfield (2011) have utilized an approach integrated by EOF and CCA. It that study, it is highlighted that this method can capture the influence of the main variability modes on the horizontal wind field in the region. Besides that, in other studies, such as Maldonado and Alfaro (2010b, 2011) and Maldonado et al. (2013), a combination of EOF and CCA has been used to related the extreme precipitation events in Central America with the SST anomalies, showing very good performance in both capturing the main variability modes of the region such as ENSO or AMO and in the estimation of the precipitation distribution. Thereby, both techniques are thought to be used in this research, in order to analyze the variability of the jet and its influence on the distribution of precipitation in Central America at seasonal and daily scales. A brief description on both methods comes in rest of this section.

Principal Component Analysis (PCA): The idea behind the use of PCA is in its capacity to reduce a data set containing a large number of variables to a data set containing fewer (hopefully many fewer) new variables. These new variables are linear combinations of the original ones, and these linear combinations are chosen to represent the maximum possible fraction of the variability contained in the original data Wilks (2005). Data for atmospheric and other geophysical fields generally exhibit many large correlations among the variables, and a PCA results in a much more compact representation of their variations. Here PCA has the potential for yielding substantial insights into both the spatial and temporal variations exhibited by the field or fields being analyzed, and new interpretations of the original data can be suggested by the nature of the linear combinations that are most effective in compressing the data.

Usually, PCA are calculated using the anomalies vector. Let's say that the field is represented by the vector $\mathbf{x}$, so the anomalies are $\mathbf{x}' = \mathbf{x} - \bar{\mathbf{x}}$, where the bar represent the mean. The total number of stations or observations are represented by K. PCA find the total M vectors $\mathbf{u}$ such as $M \ll K$. This is aimed to eliminate redundant information from the observations.

Calculating the eigenvectors of the covariance (unstandardized) or correlation (standardized) matrix of $\mathbf{x}'$, $\mathbf{S}$ or $\mathbf{R}$ respectively, each new variable or principal components, u_m, is defined accounting the maximum amount of the joint variability. In particular each u_m is obtained as the projection of the data vector x' on the m^{th} eigenvector, e_m.

$$u_m = \mathbf{e}_m^T \cdot \mathbf{x}' = \sum_{k=1}^{K} e_{km} x_k', \qquad \text{m = 1,..., M.} \tag{1}$$

Then, the m PC, u_m is that linear combination of $\mathbf{x}'$ having the larges variance, subject to the condition that they are uncorrelated with the PC having lower indices. The result is that all the PCs are mutually uncorrelated. Other properties of the PCA are discussed in Wilks (2005); Navarra and Simoni (2010).

When physical interpretation rather than data compression is a primary goal of PCA, it is often desirable to rotate a subset of the initial eigenvectors to a second set of new coordinate vectors. Rotated eigenvectors are less prone to the artificial features resulting from the orthogonality constraint on the unrotated eigenvectors. They also appear to exhibit better sampling properties than their unrotated counterparts. One cost is that the dominant-variance property of PCA is lost. The first rotated principal component is no longer that linear combination of the original data with the largest variance.

When the analysis of multiple field is sought, the combined EOF technique is applied. The correlation matrix is formed by two or more fields having L variability and K locations. Application of PCA to this kind of correlation matrix will produce principal components successively maximizing the joint variance of the L variables in a way that considers the correlations both between and among these variables at the K locations. This joint PCA procedure is sometimes called combined PCA, or CPCA.

The combination of fields in this way requires some care to handle different units and quantities. Different data have widely different numerical values corresponding to the different units that are used to measure them. These differences could generate systematic deviations in the resulting patterns that do not correspond to real variability patterns. The problem can be overcome by transforming the data to values of the same order of magnitude by using suitable scales, making the data adimensional.

Canonical Correlation Analysis (CCA): CCA is a statistical technique that identifies a sequence of pairs of patterns in two multivariate data sets, and constructs sets of transformed variables by projecting the original data onto these patterns. In CCA, the patterns are chosen such that the new variables defined by projection of the two data sets onto them exhibit maximum correlation, while being uncorrelated with the projections of the data onto any of the other identified patterns. That is, CCA identifies new variables that maximize the interrelationships between two data sets, in contrast to the patterns describing the internal variability within a single data set identified in PCA (Wilks 2005).

Canonical correlation analysis can also be viewed as an extension of multiple regression to the case of a vector-valued predictand variable. CCA looks for pairs of sets of weights analogous to the regression coefficients, such that the correlations between the new variables defined by the respective dot products with $\mathbf{x}$ and $\mathbf{y}$ are maximized.

A CCA transforms pairs of original centered data vectors $\mathbf{x}$ and $\mathbf{y}$ into sets of new variables, called canonical variates, v_m and w_m, defined by the dot products

$$v_m = \mathbf{a}_m^T \cdot \mathbf{x}' = \sum_{i=1} a_{m,i} x'_i, \qquad \text{m = 1,..., min(I,J)} \tag{2}$$

and

$$w_m = \mathbf{b}_m^T \cdot \mathbf{y}' = \sum_{i=1} b_{m,i} x'_i, \qquad \text{m = 1,..., min(I,J)} \tag{3}$$

This construction of the canonical variates is similar to that of the principal components u_m, in that each is a linear combination (a sort of weighted average) of elements of the respective data vectors $\mathbf{x}$ and $\mathbf{y}$. These vectors of weights, a_m and b_m, are called the canonical vectors.

The canonical vectors are the unique choices that result in the canonical variates having the properties

$$\text{Corr}[v_1, w_1] \geq \text{Corr}[v_2, w_2] \geq ... \geq \text{Corr}[v_M, w_M] \geq 0; \tag{4}$$

$$\text{Corr}[v_1, w_1] = \begin{cases} r_{C_m}, & \text{k=m} \\ 0, & \text{k} \neq \text{m.} \end{cases} \tag{5}$$

More properties can be consulted in Wilks (2005); Navarra and Simoni (2010)

6. Future work

This project is part of the Centre for Natural Disaster Science, which has among its aims, to improve the management of natural disasters by an increased understanding of their reasons and origins. In that sense, the planed future research is to focus on CLLJ and its relationship with extreme precipitation events. As has been previously mentioned in this essay, the association between the CCLJ and droughts or heavy rainfall events in Central America is still unclear.

In order to achieve that goal, my research aims at using a meso-scale model as an important tool to produce high resolution data to improve the understanding of the CLLJ, as a key issue of the climate in Central America. As an added value, it is worthy to mention that this type of approach and the information derived from it, can be used as an input of data in regions with scarcity of observations such as the IAS, however, with the caveat of the uncertainties due to a) there is not enough knowledge of the physical processes in the atmosphere, and precisely b) to the lack of observations in this region.

In addition, as it was noticed by Amador (2008), Amador et al. (2010), and Martin and Schumacher (2011a), it should be investigated how the CLLJ is rendered in the GCMs or reanalysis data (see for instance Fig. 13 in Amador 2008). These authors have noticed that, either a wrong representation of the vertical wind structure, or the wind speed maximum in July, or the jet at all, leads to differences, even deficiencies, in the estimation of the precipitation by the general circulation models.

Moreover, it has been documented that variations in the wind field linked to some large-scale indexes such as ENSO, NAO or MJO, might be reflected with precipitation anomalies, and so, it might be associated with extreme events such as the previously mentioned, or with the tropical cyclone activity in the Caribbean Sea and North Atlantic Ocean. Thereby, in order to understand how is this association, a statistical approach, employing methods such as the PCA and/or CCA, is sought. Using these techniques, then, information on how such large-scale modulators interfere with the wind field in the Caribbean Sea, and so, with precipitation in Central America will be generated. Thus, the research plan has four objectives describe as follows:

- To study the physical and dynamical mechanisms of the Caribbean low-level jet. In order to achieve this goal, the first part of this study will focus in analysing large-scale forcing features during the jet maxima (February and July), and thereby, how the CLLJ is captured in the reanalysis data such as the NCEP/NCAR and the ERA-Interim;

- The study is aimed at advancing the understanding of the regional physical mechanism that help to maintain the annual cycle of this jet, using the regional climate model Weather Research and Forecast (WRF). In this part, I might be working in collaboration with the Center for Geophysical Research at University of Costa Rica, in the experimental design and relevant study cases to aboard during this stage. Thereby, it is expected to coordinate what kind of information, i.e. time and spatial resolution, is need by researchers in both Earth and social sciences areas;

- The third objective is linked to the previous one, it aims in the analysis of the issues that imply the use of two-way nesting in regional models, such as the WRF and MM5 models. It has been documented (see e.g. Maldonado 2012) some issues in the estimation of precipitation, related with the physical parametrization schemes, such as the cumulus schemes, and they appear at the boundaries of the nested domains;

- The fourth objective is to study the relationship between the CLLJ and the precipitation distribution over Central America. Also, how this jet varies according to El Niño Southern Oscillation (ENSO) episodes, and Atlantic Multi-decadal Oscillation phases. An index for the horizontal wind field at 925 hPa, and the use of PCA and CCA is seeking to employ in order to study such variability modulator and how they revamp the horizontal wind field at 925 hPa and precipitation, at daily and monthly scales. The latter still needs to be examined. As it has been mentioned, part of this work will be focused on anomalous CLLJ and its relation with extreme precipitation events; thereby, it is planned to coordinate collaborative other CNDS PhD candidates, in particular on-going work focusing on droughts in Central America.

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