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Temporal and spatial characteristics of rainfall patterns in the Northern Sierra of Peru – A case study for La Niña to El Niño transitions from 2005 to 2010

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Abstract

The climatic conditions of the northern Sierra of Peru are marked by the interaction of different macro- to mesoscale climatic features such as the El Niño Southern Oscillation (ENSO) or Mesoscale Convective Complexes (MCCs) and the seasonally shifting Intertropical Convergence Zone (ITCZ), but also by local scale climatic features such as inhomogeneous topography and local wind fields. The region under investigation, located in the vicinity of the South America Continental Water Divide (CWD), provides the opportunity to study interactions of western and eastern disturbances in a high mountain environment and their effects on rainfall variability. In general, rainfall variability is related to diurnal convection patterns, enhanced by valley breeze systems and modulated by local scale wind anomalies. Spillover of low-level air masses of Pacific origin passing over the Andean ridges is frequent. Although direct effects of ENSO on high Andean rainfall variability are in debate, our findings show that the majority of rain gauges used in this study follow an El Niño/dry and a La Niña/wet signal. However, high elevation areas on the western escarpment of the Andes benefit from abundant nocturnal rainfall that partly offsets the rainfall deficits during El Niño. Our data suggest that the spatial extent of this easterly wet pulse is limited to areas located above 3000 m asl. ENSO cycles contribute to rainfall variability near the CWD in the northern Sierra of Peru by modulating the seasonal rainfall regime and causing a positive temperature anomaly.

Keywords: Rainfall variability, Mountain environments, Continental Water Divide, Peru, ENSO.

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Características temporales y espaciales de los patrones de precipitaciones en la sierra norte del Perú. Un estudio de caso de las transiciones de La Niña a El Niño desde 2005 a 2010

Resumen

Las condiciones climáticas de la sierra norte del Perú están marcadas por la interacción de diferentes rasgos climáticos a escalas macro y meso, tal como el fenómeno ENSO o los Complejos Convectivos a Meso-escala (MCC) y la estacionalmente móvil Zona de Interconvergencia Intertropical (ITCZ), pero también por rasgos climáticos a escala local tales como la heterogeneidad topográfica y los campos de vientos locales. La zona de estudio se ubica en las cercanías de la divisoria de aguas continentales de Sudamérica (CWD), provee la oportunidad de estudiar las interacciones de las disturbancias occidental y oriental en un ambiente de alta montaña y sus efectos en la variabilidad de las precipitaciones. En general la variabilidad de las precipitaciones se relaciona con patrones de convección diurna, reforzada por los sistemas de brisas de valle y modulada por anomalías de vientos a escala local. El desplazamiento de masas de aire a baja altura provenientes del Pacífico que remontan los Andes es frecuente. Aunque los efectos directos del ENSO en la variabilidad de las precipitaciones andinas está todavía en debate, nuestras investigaciones muestran que la mayoría de las mediciones hechas para este estudio siguen una señal El Niño/seco y La Niña/húmedo. Sin embargo, las zonas altas de las escarpas occidentales de los Andes se benefician de abundante lluvia nocturna que parcialmente completa los déficits de precipitaciones durante El Niño. Nuestros datos sugieren que la cobertura espacial de este pulso húmedo del Este se limita a áreas que se ubican encima de los 3000 msnm. Los ciclos ENSO contribuyen a la variabilidad de las precipitaciones cerca del CWD en la sierra norte del Perú mediante la modulación del régimen estacional de lluvias que causa una anomalía de temperatura positiva.

Palabras clave: variabilidad de precipitaciones, ambientes de montañas, divisoria continental de aguas, Perú, ENSO.

Introduction

The climate of tropical and subtropical South America is characterized by a seasonal cycle following the meridional migration of the Intertropical Convergence Zone (ITCZ) due to changing seasonal insolation and thermal inertia of the land (Hoffmann, 1992; Poveda *et al.*, 2006). In austral winters (June-September), the ITCZ is located north of the equator, and the southern hemispheric high pressure cells and the cold Humboldt Current off the western coast of South America reach their northernmost position. A bridge of high pressure between the high pressure cells over the southern Pacific and southern Atlantic hinders convection, thus leading to a pronounced dry season over southern America (Bendix & Lauer, 1992; Garreaud, 2009; Garreaud *et al.*, 2009). At the beginning of the austral summer (October-May), there is a rapid southward shift of the zone of convection, and a continental low forms over the Chaco region (~25°S), so that a broad area of heavy precipitation emerges (Horel *et al*., 1989; Seluchi *et al*., 2003; Garreaud *et al*., 2009).

In the area west of the South American Continental Water Divide (CWD), hereafter named the Pacific realm, annual rainfall increases with distance from the coast. Owing to enhanced insolation during day-time, a sea-land wind system develops, transporting offshore air masses over land. Normally the intruding air masses are dry owing to relatively cold sea surface temperature (SST) off the western coast of South America. Heating of intruding air masses over coastal deserts is not strong enough to destabilize the cold and dry air masses, thus resulting in an arid to hyperarid climate. However, when the air masses reach the Piedmont area the sea-land wind system is displaced by a valley wind system, which enables rainfall due to orographic uplift (Caviedes and Endlicher, 1989; Bendix & Lauer, 1992).

The area east of the CWD, hereafter named the Atlantic realm, and the eastern Andean slopes are more affected by easterlies transporting moist air over the Andes (Vuille *et al*., 2000; Ronchail & Gallaire, 2006; Garreaud, 2009). Espinoza *et al*. (2009) investigated rainfall distribution within the Amazon basin and reported that rainfall increases from the lowlands to the foothills of the eastern Andes owing to orographic uplift. On the basis of 391 rain gauges, the authors showed that rain gauges registering more than 3000 mmy^{-1} are located below 1500 metres above sea level (m asl). At higher altitudes inadequate moisture supply results in a decrease in rainfall (Espinoza *et al*., 2009).

On shorter timescales, cloudiness and heavy rainfall over tropical and subtropical South America, associated with convective activity, follow a diurnal cycle (de Angelis *et al*., 2004; Poveda *et al*., 2005; Bendix *et al*., 2006). Garreaud and Wallace (1997) studied the diurnal march of convective cloudiness over America by investigating satellite derived data and found several parallel bands of enhanced cloudiness, one of them located along the subtropical Andes. These bands of enhanced cloudiness build up rapidly in the afternoon and decline gradually during night-time. Many studies on weather dynamics and regional climatic conditions for the subtropical Andes confirmed that high rainfall and high rain rates occur during the afternoon and are most pronounced close to sunset. Poveda *et al*. (2005) analysed 51 rain gauges with records spanning between 22 and 28 years in the tropical Andes of Colombia and found diurnal (24h) and semidiurnal (12h) cycles with precipitation maxima occurring in the afternoon or around midnight, or both. Bendix *et al*. (2006) used a K-band rain-radar profiler to distinguish stratiform and convective rainfall in the Rio San Francisco valley in southern Ecuador. They found most rainfall to be of stratiform character; however, a large amount of rainfall is due to convective activity. Two rainfall maxima have been reported: a predawn/dawn and an afternoon maximum. Bendix *et al*. (2006) ascribed the afternoon maximum to enhanced thermally induced convection, intensified by an up-slope breeze system. The predawn/dawn maximum, which they report to be highly variable regarding rainfall and rain frequency, is associated with larger-scale atmospheric instabilities such as mesoscale convective systems (MCSs) or mesoscale convective complexes (MCCs). MCSs are long-lived, multicellular structures with distinct regions of both convective and stratiform precipitation (Houze Jr., 2004). A special case of MCS is the mesoscale convective complex (MCC), defined by Maddox (1980) as a long-lasting, quasi-circular cold-cloud shield (Houze Jr., 2004), which occurs primarily over land and is found to be nocturnal (Laing & Fritsch, 1997). MCCs usually develop in the mid-to-late afternoon. By around sunset, the cold-cloud shields have amalgamated into a coherent mesoscale structure. The cold-cloud shield reaches its maximum extent after midnight and dissipates a few hours after sunrise (Laing & Fritsch, 1997). Durkee and Mote (2010) analysed 330 MCCs that occurred in the subtropical South American region during the austral warm period (October-May) between 1998 and 2007. They found that MCCs occur most frequently during December and January and that the largest MCCs, expressed by maximum cloud-shielded area, occur during late spring (AM) and late fall (OND). MCCs develop most frequently east of the Andes between 20°S and 30°S.

1. ENSO and rainfall variability

Many studies on climate variability and thus rainfall variability in tropical South America focus on the phenomenon of El Niño Southern Oscillation (ENSO) (Horel & Cornejo-Garrido, 1986; Goldberg *et al.*, 1987; Aceituno, 1988; Caviedes & Endlicher, 1989; Bendix & Lauer, 1992; Ropelewski & Halpert, 1996; Garreaud & Battisti, 1999; Vuille, 1999; Vuille *et al.*, 2000; Poveda *et al.*, 2006; Ronchail & Gallaire, 2006) because climate and rainfall variability over much of South America is closely related to coupled ocean-atmosphere interactions of ENSO (Kayano *et al.*, 1988; Ropelewski & Halpert, 1996; Vuille *et al.*, 2000; Garreaud, 2009). Owing to the non-linear chaotic nature of the ocean-atmosphere system, however, each phase–e.g. El Niño (EN) or La Niña (LN)–varies in its time of initiation, severity and spatial extent. In consequence, various indices of the ENSO phenomenon vary in their sign and strength, just as inter-annual variability in precipitation varies in space (Waylen & Poveda, 2002).

A direct and strong effect of ENSO variability is observed over coastal Ecuador, Peru and northern Chile; thus, more rainfall than normal is observed during EN (Horel & Cornejo-Garrido, 1986; Goldberg *et al.*, 1987; Tapley & Waylen, 1990; Bendix, 2000). The intrusion of relatively warm water off the coast displaces the colder Humboldt current and destabilizes the low-level inversion over the cold surface waters of the

Humboldt current, so that enhanced convection over the coastal plains (< 100 m asl) of Ecuador and northern Peru triggers devastating rainfall (Horel & Cornejo-Garrido, 1986; Goldberg *et al.*, 1987; Caviedes & Endlicher, 1989; Bendix & Lauer, 1992). For more elevated regions such as the Andean Piedmont (up to 1000 m asl), rainfall anomalies are less intense. With increasing elevation and thus increasing distance from the coastal plain, rainfall anomalies are hardly distinguishable from normal year to year variability of rainfall (Caviedes & Endlicher, 1989; Bendix, 2000).

In fact, the effect of ENSO on rainfall variability in the higher Andes is not as conclusive. Rossel and Cadier (2009) could not establish a direct link between ENSO and the annual rainfall of inter-Andean valleys and eastern Andean slopes in Ecuador. Celleri *et al.* (2007) report that for inter-Andean depressions in Ecuador small-scale features such as topography or local convection patterns contribute significantly to rainfall variability. In the central Peruvian Andes, considerable fluctuations in rainfall are related neither to the ENSO phase nor to its strength. Kane (2000) observed a negative or mixed response to El Niño phenomena by studying rainfall variability at Huancayo in the central Peruvian Andes. The author reported either a deficit in DJFM rainfall or an excess in DJ rainfall, followed and preceded by deficits during EN phases. On the mid-elevation eastern Andean slopes (> 1500 m asl) of Bolivia as well as on the Altiplano, a LN wet signal is observed (Tapley & Waylen, 1990; Vuille, 1999; Garreaud & Aceituno, 2001; Ronchail & Gallaire, 2006). By contrast, EN triggers lower than normal rainfall and thus dryer conditions in that area.

2. Objectives

Starting with this contrasting information, we intend to investigate temporal and spatial patterns of rainfall within the La Niña-El Niño transitions of 2005 to 2010 to fill the knowledge gap relating to seasonal distribution, rainfall abundance, rain rates and the interrelation with low-level wind fields in the vicinity of the CWD in the northern Sierra of Peru.

Until now very few studies have focused on rainfall variability in mesoscale headwater areas close to the CWD (Buytaert *et al.*, 2006; Celleri *et al.*, 2007; Romero *et al.*, 2007). However, such headwater areas are crucial, being essential for water availability, upland and lowland irrigation purposes, human and industrial (e.g. mining) demand and the use of hydro power, especially for the arid west coast of South America. Understanding rainfall variability is indispensable in order to estimate water yield, assess water scarcity issues and evaluate measures to improve systems of sustainable water consumption. Moreover, knowledge of annual precipitation and its temporal and spatial distribution is of great interest for the affected population, for public stakeholders and for hydrological modellers to calibrate and validate hydrological model accuracy. However, accurate measurement and estimation of spatial and temporal distribution of precipitation is difficult (Buytaert *et al.*, 2006). This is especially true for remote mountainous environments where continuous data sets are scarce and temporal and spatial patterns of rainfall are strongly influenced by the irregular topography (Celleri *et al.*, 2007; Espinoza *et al.*, 2009; Rollenbeck & Bendix, 2011).

3. Study area and related environmental problems

The object of this study is to investigate rainfall variability at meso- to local scale in the northern Peruvian highlands within a La Niña (LN)–El Niño (EN) transition. The research was carried out in the region of Cajamarca in the northern Peruvian highlands (= Sierra), between 7°01' to 7°19' S and 78°19′ and 78°45' W, at elevations from 1005 to 4200 m asl (Fig. 1). Rainfall variability and rainfall characteristics are in general important for the people in the region and especially in the city of Cajamarca. At present, water availability in the austral winter period (June to September) does not meet the demand in the city of Cajamarca. Water scarcity issues due to population growth emerged over the past decades; however, a significant amount of water is lost owing to insufficient retention potential of the river catchments serving the city of Cajamarca. Rainfall characteristics, steep slopes, shallow soils and non-existing or inadequate land use strategies lead to landscape degradation due to soil erosion by surface runoff. In many parts of the catchments surrounding the city of Cajamarca, topsoil is already irreversibly eroded, and thus soils are no longer able to retain rainfall, which progressively intensifies water scarcity issues for the Cajamarca region. It is of special interest to learn more about rainfall characteristics and variability in the vicinity of the CWD as these remote areas are difficult to access but are very important for runoff generation; lack of knowledge is thus a major hurdle to assessing water related issues.

The CWD in the region of Cajamarca comprises the watersheds of the westward draining Rio Jequetepeque and eastward draining Rio Cajamarca (Fig. 1). The Rio Jequetepeque drains directly into the Pacific Ocean, whereas the Rio Cajamarca is part of the Amazon River basin and thus drains into the Atlantic Ocean. In the Peruvian highlands, several elevation zones, each with its unique flora and fauna, have been distinguished on the basis of their climatological conditions (Pugar Vidal, 1996). The study area encompasses the elevation zones of the Yunga (500-2300 m asl), the Quechua (2300-3500 m asl) and the Jalca (> 3500 m asl). The Jalca is described as a transition zone between the more humid Paramo in the north and the more arid Puna type landscape in the south (Sánchez Vega *et al.*, 2006). In general, zones of higher elevation receive more annual precipitation. Mean annual rainfall rises from > 100 mm at Yunga to > 900 mm at Quechua to > 1100 mm at Jalca (ONERN, 1975; 1977).

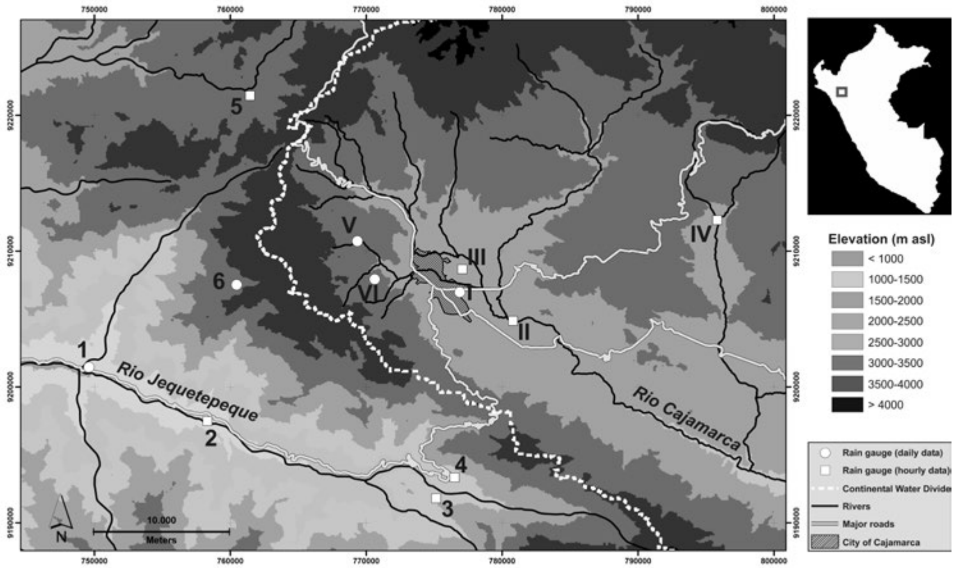

Figure 1. Location map of weather stations used in this study in the headwater basins of Rio Jequetepeque and Rio Cajamarca. Within the Pacific realm these are: (1) Huaquillas, 1034 m asl; (2) Magdalena, 1260 m asl; (3) Asunción, 2170 m asl; (4) San Juan, 2460 m asl; (5) Granja Porcón, 3180 m asl and (6) Alto Chetilla, 3422 m asl. Weather stations within the Atlantic realm are: (I) Cajamarca-UNC, 2536 m asl; (II) La Victoria, 2625 m asl; (III) Augusto Weberbauer, 2660 m asl; (IV) La Encañada, 2980 m asl; (V) Chamis, 3224 m asl and (VI) Ronquillo, 3325 m asl. See Table 1 for detailed characteristics of each weather station used in this study.

4. Methods

For this study we used seven weather stations providing daily precipitation data: Magdalena (1260 m asl), Asunción (2170 m asl), San Juan (2460 m asl), Augusto Weberbauer (2660 m asl), Granja Porcón (3180 m asl), La Encañada (2980 m asl) and La Victoria (2625 m asl). These rain gauges are maintained by Servicio Nacional de Meteorología e Hidrología del Perú (SENAMHI). We installed three DAVIS Vantage Pro II stations with a one-hour resolution of data storage (Huaquillas, 1034 m asl; Chamis, 3224 m asl; Alto Chetilla 3422 m asl). Within the past three years, local authorities in cooperation with SENAMHI and CARE installed two more weather stations with hourly data sets which are incorporated in our study as well (Ronquillo, 3325 m asl; Cajamarca-UNC, 2536 m asl). The area covered by the rain gauges is about 816 km², giving a network density of 1 per 68 km². Because of the relatively short time series, we limited our analysis to descriptive statistics methods and correlation analysis. For ascribing seasonality we used the SI Index first published by Walsh and Lawler (1981).

5. Results

5.1. General climatic conditions

The weather stations under investigation are located to the west and east of the CWD (Fig. 1). The general climatic conditions are presented in Table 1. Throughout the investigation period (2005-2010) annual precipitation varies from 437 mm to 1555 mm, depending on elevation and locality. The region under investigation is characterized by a mono- to bimodal rainy season. Rainfall from October to December is less abundant than rainfall from January to May. The Seasonality Index (SI), an indicator for seasonality in rainfall and thus a parameter for water availability, varies from 0.51 to 1.1, expressing seasonality from «rather seasonal with a short dry season» to «most precipitation in less than 3 months» (Walsh and Lawler, 1981). More seasonal conditions are observed for rain gauges at lower elevations; these are accordingly located more closely to the Pacific Ocean. The SI is lower for higher areas located closer to the CWD, indicating less seasonal conditions.

#	Name	Elevation (masl)	Realm	Max. Temp. (°C)	Min. Temp. (°C)	Mean Annual precipitation (mm)	Seasonality Index-SI	Years of record
1	Huaquillas (a)	**1034**	Pacific	33.2	**13.9**	**437**	1.02	2008-2010
2	Magdalena (b)	1260	Pacific	**34.8**	11.7	526	0.98- 1.10	2005-2010
3	Asunción (b)	2170	Pacific	30.0	6.2	1280	0.96- 1.02	2005-2010
4	San Juan (b)	2460	Pacific	34.0	8.4	**1157**	0.91- 1.00	2005-2010
5	Granja Porcón (b)	3180	Pacific	22.4	-6.4	1555	0.54- 0.76	2005-2010
6	Alto Chetilla (a)	**3422**	Pacific	**19.1**	0.3	1189	0.69	2008-2010
I	Cajamarca-UNC (a)	2536	Atlantic	25.3	0.6	---	---	2009-2010
II	La Victoria (b)	2625	Atlantic	26.0	**-7.0**	517	0.65- 0.98	2005-2010
III	Augusto Weberbauer (b)	2660	Atlantic	25.3	-2.3	719	0.61- 0.84	2005-2010
IV	La Encañada (b)	2980	Atlantic	26.3	-4.9	1022	0.62- 0.82	2005-2010
V	Chamis (a)	3224	Atlantic	19.4	2.1	1159	0.61	2008-2010
VI	Ronquillo (a)	3325	Atlantic	22.2	3.8	---	---	2008-2010

Table 1. General climatic features of weather stations under investigation (2005-2010). (a) Automatic rain gauges with a temporal resolution of one hour, (b) conventional weather stations with a temporal resolution of one day. Maximum and minimum values are marked in bold.

In high Andean mountain environments, spatial rainfall patterns are closely connected to main air flow trajectories, which are altered and canalized by topographic features (Buytaert *et al.*, 2006; Celleri *et al.*, 2007; Rollenbeck & Bendix, 2011). The spatial correlation of daily and monthly rainfall data presented in Figure 2 indicates that the spatial rainfall pattern is more diverse in the Atlantic realm than on the western escarpment of the

Andes (Pacific realm). In general the correlation of rainfall data decreases with distance; in the Pacific realm, however, higher correlation coefficients are observed.

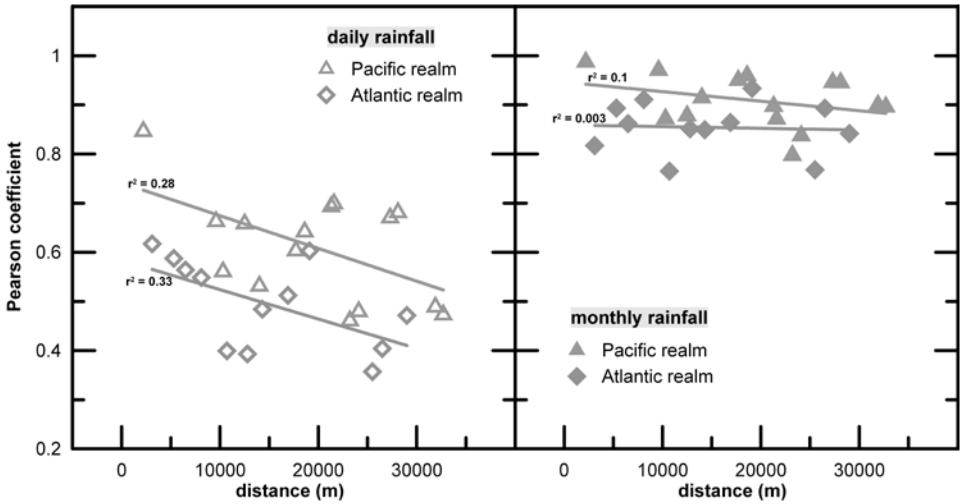

Figure 2. Spatial correlation of daily (left panel) and monthly (right panel) rainfall data for rain gauges in the Pacific realm (triangles) and the Atlantic realm (diamonds). Coefficient of determination (r^2) is given for each data array.

5.2. ENSO cycles and ASR rainfall variability

In this section we compare the occurrence of El Niño (EN) to the austral summer rainfall (ASR), which we define as rainfall during the six month period from October to March. We restricted our analysis to the austral summer rainfall period because 68 to 83% of annual rainfall is observed within this period. To identify representations of ENSO events, we chose the Oceanic Niño Index (ONI) from the United States National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC). The ONI is derived from SST anomalies in the equatorial Pacific Ocean using a three month running mean of ERSST.v3 SST anomalies in the Niño 3.4 region (5°N-5°S, 120°-170°W). Cold (LN) and warm (EN) episodes are defined when the threshold +/- 0.5°C is met for a minimum of five consecutive and overlapping seasons. Anomalies in austral summer rainfall (ASR) for seven rain gauges under investigation and the corresponding ENSO phase based on ONI are given in Table 2. ASR is lower in the case of EN (2006/2007 and 2009/2010) compared to LN (2005/2006, 2007/2008 and 2008/2009). The negative rainfall anomalies for rain gauges in the Pacific and Atlantic realms vary between 3% and 20%. In LN episodes (2007/2008 and 2008/2009), ASR is equal to or higher than normal for all rain gauges under investigation The positive rainfall anomalies vary between 0% and 20%. Only the rain gauge at Granja Porcón (3180 m asl), located in the Pacific realm, contradicts this observed pattern. Rainfall in the LN period of 2005/2006 is lower than normal, whereas rainfall in the EN period of 2009/2010 is slightly higher than normal (Table 2).

Besides anomalies in ASR due to ENSO episodes, a modified seasonal rainfall distribution during El Niño phases is observed for the rain gauges under investigation. The modification of the bimodal rainfall distribution is seen in a delay of the secondary minimum, which is expected to occur in December, but is delayed for one or two months and thus occurs in January to February. Correlation analysis of ONI and monthly rainfall confirms that during EN typically above-normal rainfall is observed in December, succeeded by below-normal rainfall in January and/or February (Table 3), whereas during LN typically below-normal rainfall is observed in December and above-normal rainfall in January and/or February. This consequence of EN-LN transition is more pronounced in the data record of rain gauges located in the Pacific realm (Fig. 3). In the Atlantic realm this interrelation exists; however, depending on locality, the correlation is significantly weaker (A. Weberbauer) or is missing (La Encañada). We suppose that at the rain gauges in the Atlantic realm the modulation of seasonal rainfall distribution is masked by local effects or that EN driven variability does not influence areas east of the CWD as much as areas west of the CWD, as they are beyond the reach of western disturbances.

		Pacific realm				Atlantic realm		
		Magdalena	Asunción	San Juan	Granja Porcón	La Victoria	A. Weberbauer	La Encañada
ASR (mean)		441	1023	961	1146	396	564	750
Year	Typ	difference to mean (%)						
2005-2006	LN	7.52	0.18	4.67	-10.48	17.31	5.41	18.74
2006-2007	EN+	-9.01	-20.80	-12.61	-9.17	-10.82	-20.16	-12.38
2007-2008	LN+	3.61	12.90	7.72	11.83	9.82	9.49	4.74
2008-2009	LN	18.96	26.12	14.77	7.53	2.54	8.13	11.23
2009-2010	EN+	-21.08	-18.40	-14.55	0.29	-21.44	-2.86	-18.73

Table 2. Austral summer rainfall anomalies and ENSO cycles. EN phases are framed. A plus (+) stands for a more pronounced ENSO phase based on Ocean Niño Index (ONI) being > +/-1

	Pacific realm				Atlantic realm		
Month	Magdalena	Asunción	San Juan	Granja Porcón	A. Weberbauer	La Victoria	La Encañada
October	0.011	0.045	0.114	0.051	-0.275	-0.104	-0.323
November	0.104	0.588	0.375	0.351	0.329	0.160	0.109
December	0.564	0.793	0.736	0.614	0.270	0.700	-0.253
January	-0.816	-0.738	-0.819	-0.479	-0.402	-0.643	-0.204
February	-0.487	0.620	-0.703	-0.535	-0.273	-0.671	-0.427
March	-0.196	-0.433	-0.309	0.111	0.095	0.087	0.182

Table 3. Pearson correlation coefficients of Ocean Niño Index (ONI) and monthly rainfall anomalies for seven rain gauges.

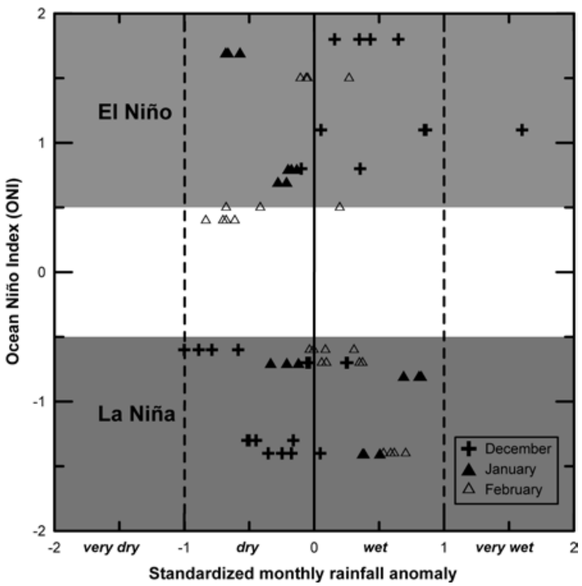

Figure 3. Standardized monthly rainfall anomalies for DJF rainfall in the Pacific realm (based on the rainfall data of the rain gauges Magdalena, Asunción, San Juan and Granja Porcón). During EN more rainfall is generally observed in December, whereas rainfall tends to be lower in January and February. By contrast, during LN rainfall tends to be lower in December and higher in January and February.

5.3. Rainfall characteristics of 2008/2009 (LN) and 2009/2010 (EN)

The detailed characteristics of the austral summer rainfall (October-March) of two subsequent years based on hourly datasets are presented in Table 4. The rain gauges at Huaquillas, Alto Chetilla, Chamis and Ronquillo are located far away from each other and at different elevations. Huaquillas (1034 m asl) and Alto Chetilla (3422 m asl) are part of the Pacific realm, whereas Chamis (3224 m asl) and Ronquillo (3325 m asl) are located east of the CWD and therefore in the Atlantic realm.

Austral summer rainfall in 2008/2009 (LN) is 402 mm, 945 mm, 876 mm and 533 mm for the rain gauges at Huaquillas, Alto Chetilla, Chamis and Ronquillo, respectively. The lowest ASR is observed at the low-altitude rain gauge at Huaquillas, whereas highest ASR is observed at the high altitude rain gauge of Alto Chetilla. Both rain gauges located east of the CWD recorded lower ASR due to rainout effects by moist air masses passing the high mountain ridges encircling the valley of Cajamarca.

ASR in 2009/2010 (EN) is 227 mm and 726 mm for Huaquillas and Chamis rain gauges, respectively, and thus is lower than in the preceding year. Rainfall at Alto Chetilla and Ronquillo rain gauges is 996 mm and 591 mm, respectively, and higher than in 2008/2009. However, Ronquillo rain gauge values are not appropriate for comparisons as October is not included in its data array. Similarly to rain gauge

Granja Porcón rain gauge Alto Chetilla contradicts the observed pattern of LN/wet and EN/dry. Rainfall in the LN period of 2008/2009 is lower, whereas rainfall in the EN period of 2009/2010 is higher (Table 4). The exceptional rainfall characteristics of Alto Chetilla rain gauge will be discussed in detail in section 6.6.

2008/2009 - LN (ONDJFM)	total precipitation (mm)	maximum monthly precipitation (mm)	maximum daily precipitation (mm)	maximun rain rate (mmh^{-1})	number of storms (>25 mmh^{-1})	percentage of days with rainfall
Huaquillas	402	135	26.2	67.8	18	46
Alto Chetilla	945	255	36.2	225.8	53	74
Chamis	876	297	32.6	169.4	57	67
Ronquillo*	533	211	33.2	---		54
2009/2010 - EN (ONDJFM)						
Huaquillas	227	100	30	86.6	12	40
Alto Chetilla	996	271	60.8	195.2	64	73
Chamis	726	182	62.6	174.6	43	68
Ronquillo*	591	193	66.9	---	---	57

Table 4. Austral summer rainfall data of rain gauges at Huaquillas, Alto Chetilla, Chamis and Ronquillo with a temporal resolution of one hour for LN-EN transition. (*) no data of rainfall intensity exists, and rainfall data exclude October 2008.

5.4. Diurnal pattern in rainfall and air flow regimes

The pattern of diurnal rainfall and air flow regimes is analysed on the basis of four rain gauges offering hourly datasets. Figure 4 shows the diurnal distribution of rainfall and wind frequencies for Alto Chetilla rain gauge, representing the Pacific realm, and Ronquillo rain gauge, representing the Atlantic realm. An afternoon/early evening maximum and two less pronounced peaks around midnight and predawn are observed.

In the Pacific realm, western wind directions dominate all day long, from about 08:30 to 19:30 local time (LT). They reach their maximum abundance at about 16:00 LT (Fig. 4a). During night-time hours easterly winds are more dominant, being most abundant (up to 90%) at 06:00 LT. The well-known diurnal wind reversal in mountain valleys is associated with the development of the thermally induced valley breeze system with up-slope breeze during day-time and down-slope breeze during night-time (Wagner, 1938; Defant, 1951; Flohn, 1969; Freytag, 1987; Vergeiner & Dreiseitl, 1987).

As shown in Figure 4a, daily rainfall distribution follows a diurnal cycle as well. A maximum of rainfall is observed in the afternoon and coincides with mainly westerly winds. Abundant rainfall, starting at about 12:00 LT and ending at about 19:00 LT, accounts for 70% of total rainfall. The temporal distribution of rainfall fits quite well into the diurnal pattern of enhanced convection induced by thermal heating due to the diurnal march of the sun and boosted by orographic uplift, thus resulting in deep convection. The influence of strong convective activity during day-time is seen in the data of rain rates (Fig. 5). At Alto Chetilla rain gauge, a peak in mean maximum rain rates is observed from 12:00 LT to 14:00 LT, which is followed by gradually decreasing maximum rain rates, indicating a rapid buildup of convective activity at noon and gradual decline during the afternoon. Tokay *et al.* (1999) investigated rainfall characteristics in the central equatorial Pacific Ocean and found that almost all rain rates exceeding 10 mmh^{-1} are associated with convective clouds.

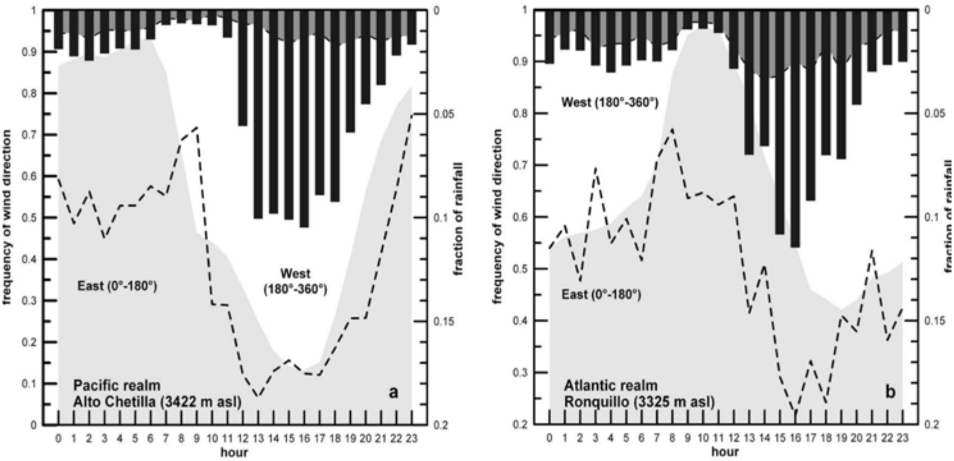

Figure 4. Diurnal wind frequencies and fraction of mean rainfall for rain gauges Alto Chetilla (a) and Ronquillo (b). Shaded areas indicate easterly wind directions; blank areas indicate westerly wind directions. The dashed line represents the diurnal frequency distribution of wind directions during rainfall. Areas above the dashed line represent westerly wind directions, and areas beneath the dashed line represent easterly wind directions. The bars representing the fraction of rainfall for each hour are underlain by a shaded area which represents the relative contribution of rainfall accompanied by easterly wind directions.

The amount of nocturnal rainfall is significantly less, but still accounts for 30% of total rainfall. During night-time easterly winds become more abundant, but in the case of rainfall, as seen in Figure 4a, the frequency of easterly and westerly winds is more or less equal between 00:00 and 07:00 LT. This in turn indicates that nocturnal down-slope breezes as well as nocturnal up-slope breezes occur, thus pointing to a complex interrelation of eastern and western air masses during night-time. Nocturnal down-slope breezes can be explained by the reversal of the wind directions within the Jequetepeque valley owing to nocturnal radiative cooling (Howell, 1953). However, nocturnal up-slope breeze points to convective activity due to up-scaling of storm size through the night and thus can be related to long-lived MCCs. Mean maximum rates of nocturnal rainfall are significantly lower (< 5 mmh^{-1}) than during day-time, thus suggesting mainly stratiform dynamics of rainfall (Fig. 5). However, mean maximum rain rates increase to about 7 mmh^{-1} at midnight (00:00 to 01:00 LT) and pre-sunrise (05:00 LT), thus suggesting that embedded shallow convection within a stratiform cloud cap results in mixed rain events. Kirshbaum and Durran (2004) showed that average rain rates for embedded convection range from 1.74 mmh^{-1} in a more stratiform cap cloud to 5.36 mmh^{-1} in a more convective case.

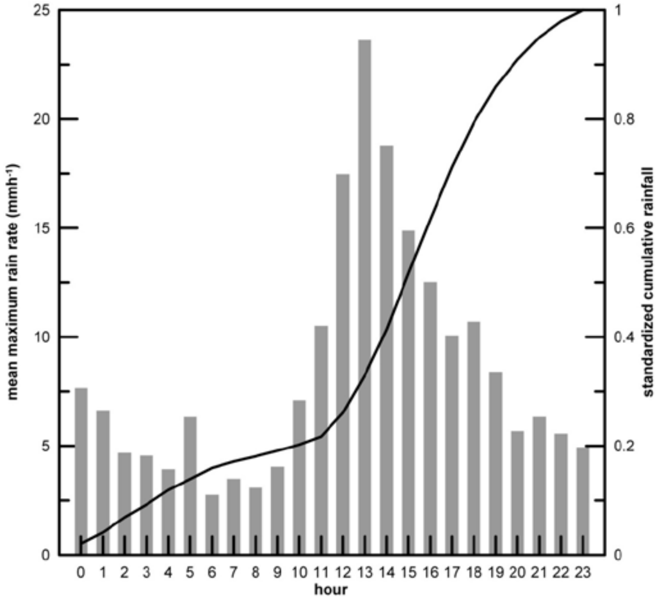

Figure 5. Diurnal mean maximum rain rate and standardized cumulative rainfall for Alto Chetilla rain gauge. Rain rates exceeding 10 mmh^{-1} are associated with convective activity; lower rain rates point to stratiform rainfall dynamics (Tokay *et al.*, 1999).

In the Atlantic realm the diurnal cycle of the wind system is more complex than in the Pacific realm (Fig. 4b). Easterly and westerly wind directions are more evenly distributed. During night-time easterly winds are slightly more frequent than westerly winds. Of special interest is the reversal of wind directions during the afternoon. In the morning to early afternoon (07:00 to 14:00 LT), eastern wind directions are most abundant with a maximum from 10:00 to 11:00 LT. They are a result of radiative heating due to the diurnal cycle of insolation and thus represent the thermally induced up-slope breeze. In the afternoon to evening (15:00 to 19:00 LT) the relative frequency of westerly wind directions increases until finally down-slope winds of westerly direction replace the easterly up-slope breeze. The overturn in wind direction is accompanied by a high abundance of rainfall. A mechanism for the observed reversal of wind directions during the late afternoon has been proposed by López and Howell (1967) and is illustrated in Figure 6. They argue that air masses of Pacific origin are moved landwards/up-slope by diurnal sea/valley breezes. When the air masses reach the mountain chain they eventually cross the lower ridges or mountain passes. The winds descend as katabatic flow and replace the easterly up-slope winds. The strength of these descending winds depends on the overflow depth and on the temperature gradient between the ridge and the valley floor to which they descend. By descending, the down-slope flow favours the formation of reversed convective cells over the valley floor, which adds an impulse to overall convective activity. In accordance with Kuettner (1959) the authors applied hydraulic formulas to describe the overflow mechanism in detail (López & Howell, 1967). The mechanism proposed by López and Howell (1967) offers an explanation for the deflection of the up-slope breeze that evolves during forenoon by air masses of Pacific origin spilling over the CWD during late afternoon and for convection that occurs during the afternoon and results in heavy precipitation.

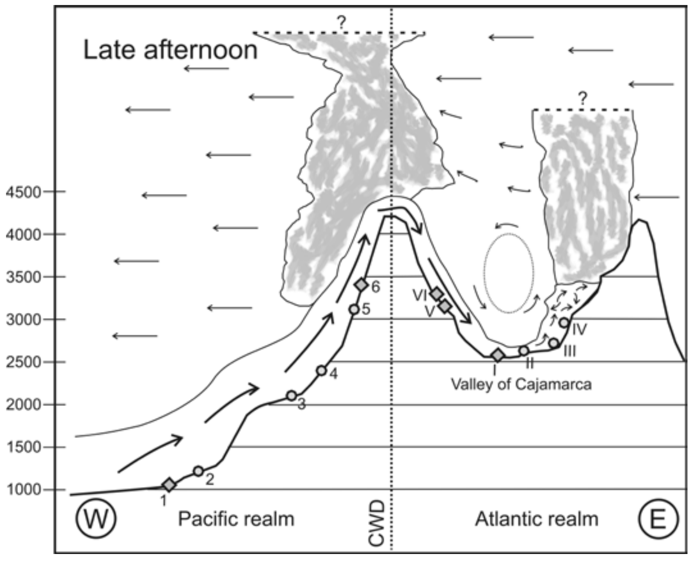

Figure 6. Schematics of the late afternoon flow. In the late afternoon westerly winds may overflow the ridges of the Andes. By descending they deflect the thermally induced up-slope breeze. At the eastern flanks of the valley a reversed convective cells develops, that strengthen convection and thus trigger rainfall. Numbers indicate rain gauges used in this study and their altitude as seen in Table 1. Note that rain gauges do not lie along a transect. Modified after López and Howell (1967).

The temporal distribution of rainfall at Ronquillo rain gauge, representing the Atlantic realm, is similar to that in the Pacific realm: 60% of rainfall occurs from

13:00 LT to 19:00 LT with a rainfall maximum from 15:00 LT to 16:00 LT (Fig. 4b). The late afternoon rainfall maximum is in good accordance with the diurnal cycle of convective activity observed over the subtropical Andes (Garreaud and Wallace, 1997).

The available data of maximum rain rates for Chamis rain gauge, representing the Atlantic realm (not shown) confirm the convective nature of rainfall as mean maximum rainfall rates range between 10 to 20 mmh^{-1} in the afternoon and are thus quite comparable to the values recorded on the western Andean escarpment. Similar to the observations made for Alto Chetilla rain gauge (see Fig. 5), nocturnal rain rates exceeding 5 mmh^{-1} suggest mixed rainfall at Chamis rain gauge that is partly of a stratiform and partly of an embedded convective nature. Nocturnal rainfall accounts for about 40% of total rainfall and is an important source of moisture, as evaporation losses are significantly lower than during day-time.

In summary, the findings attest a pronounced diurnal cyclicality in rainfall and in wind directions for both sides of the CWD. The rapid buildup of convective activity at noon and the gradual decline during the afternoon are in good accordance with the diurnal cycle of convective activity. The diurnal wind reversal is associated with the development of the thermally induced valley breeze system with up-slope breeze during the day and down-slope breeze during the night. However, a special case emerges for the valley of Cajamarca, as the thermally induced up-slope breezes of easterly direction are replaced by down-slope winds of westerly direction during the afternoon that cross the Andean ridges from western directions.

5.5. Features of inter-annual variability of LN 2008/2009 and EN 2009/2010

In this section we compare the diurnal march of relevant meteorological parameters of four rain gauges consisting of hourly datasets of two subsequent rainy seasons to deduce variations among them and discuss them against the backdrop of the LN/EN transition from 2008/2009 to 2009/2010. Figure 7 presents the data array of rainfall, mean wind speed, mean maximum wind speed and mean temperature for the rainy season of 2008/2009 and 2009/2010 for the four rain gauges with the highest temporal resolutions used in this study.

A common feature for all rain gauges is the existence of a bimodal distribution of diurnal rainfall. At all rain gauges, rainfall during afternoon/early evening is more abundant than nocturnal rainfall. It should be noted, however, that the temporal signature of rainfall distribution varies from season to season, as an offset in the temporal distribution of afternoon rainfall is noticeable. During EN 2009/2010 the temporal pattern of afternoon rainfall is characterized by a maximum at midday and a gradual decrease in rainfall during the afternoon. However, during LN 2008/2009 the temporal signature is reversed, so that rainfall starts at 13:00 LT but does not reach its maximum until 18:00 LT. This temporal signature of early afternoon rainfall maximum followed by a gradual decrease during EN 2009/2010 and late afternoon maximum preceded by a gradual increase during LN 2008/2009 is evident in the datasets of Alto Chetilla, Chamis and Ronquillo rain gauges (Fig. 7). This in turn indicates that at least the three rain gauges located at altitudes higher than 3000 m asl are subject to a common forcing. The temporal signature of rainfall at rain gauge Huaquillas in the day-time is quite similar for both seasons, however during LN of 2008/2009 hardly any nocturnal rainfall occurs, whereas during EN of 2009/2010 nocturnal rainfall becomes more important for the seasonal rainfall budget (Fig. 7).

It has been recognized that ENSO and its representation in SST anomalies affects air temperature over tropical and subtropical South America (Aceituno, 1988; Vuille *et al.*, 2000). Temperatures over the Altiplano in southern Peru tend to be higher (approximately 0.7°-1°C) during an EN episode (Vuille *et al.*, 2000), and higher temperatures in the Andes of Ecuador during EN have been associated with accelerated glacier melting (Francou *et al.*, 2004). Our data confirm earlier findings. We find a positive temperature anomaly of approximately 0.9°-1.5°C for EN of 2009/2010 compared to LN of 2008/2009 (Fig. 7). This temperature anomaly is also imprinted in the temporal signature of the wind speed, as winds compensate for differences in air pressure and/or air temperature. Mean wind speeds as well as mean maximum wind speeds for Alto Chetilla and Ronquillo rain gauges indicate an enforcement of atmospheric turbulence during the EN period of 2009/2010 (Fig. 7).

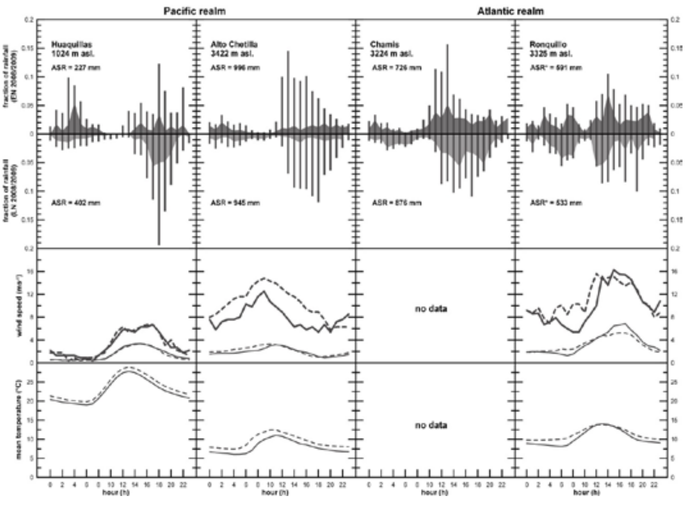

Figure 7. Diurnal course of rainfall, mean wind speed (thin line), mean maximum wind speed (thick line) and median humidity for Huaquillas, Alto Chetilla, Chamis and Ronquillo rain gauges for LN (solid line) and EN (dashed line) episodes. The shaded area represents the fraction of rainfall that coincides with easterly wind directions. * October 2008 not included in the data.

In summary, the findings show that EN 2009/2010 produces higher mean temperatures and enhances atmospheric turbulence. Moreover the convection is strongest in the early afternoon in contrast to the late afternoon maximum observed during LN 2008/2009. Strong convective activity results in the buildup of cloud systems, which in turn reduce thermal heating by reflecting incoming solar radiation. Accordingly the convective activity diminishes throughout the afternoon, and thus offers a mechanism for lower ASR observed during EN episode.

5.6. Offset of EN/dry and LN/wet scheme for high elevated rain gauges in the Pacific realm

One interesting feature of rainfall anomalies due to EN/LN remains: Why do all of the rain gauges in the Pacific realm record less ASR during EN and higher ASR during LN except for the rain gauges at Granja Porcón and Alto Chetilla? These rain gauges are the only rain gauges located at altitudes higher than 3000 m asl; thus, they are closer to CWD than any other rain gauge within the Pacific realm used in this study.

In general rain gauges of the Pacific realm are characterized by abundant rainfall during day-time in combination with westerly winds. Most rainfall is obviously associated with up-slope breezes. However, some rainfall occurs in combination with down-slope breezes of eastern direction (Fig. 4). A closer look at the seasonal distribution of diurnal and nocturnal rainfall and the prevailing wind direction at rain gauge Alto Chetilla (Fig. 8) confirms that during day-time (06:00 to 18:00 LT) most frequently rainfall coincides with westerly wind directions (~ 85%). This feature is independent of the ENSO mode. However, the relative occurrence of wind direction during night-time (19:00 to 05:00 LT) differs significantly from season to season. During EN 2009/2010 52% of nocturnal rainfall is accompanied by winds from eastern direction. Whereas during LN 2008/2009 only 29% of nocturnal rainfall coincides with easterly down-sloping winds. As down-slope winds tend to stabilize the atmosphere and thus inhibit rainfall, the presence of nocturnal rainfall indicates that atmospheric instability was transferred from the Atlantic realm in the form of MCCs developing east of the Andean mountains (Bendix *et al.*, 2006).

As seen in figure 9 easterly rainfall at Alto Chetilla rain gauge (3422 m asl) is slightly higher during EN 2009/2010 than during LN 2008/2009. However, this easterly wet pulse in 2009/2010 is not seen in the data of lower elevation rain gauge of Huaquillas, located downstream at an altitude of 1034 m asl. Rain gauge Huaquillas lacks the positive rainfall anomaly for easterly winds during EN 2009/2010 (Fig. 9). This finding suggests that eastern disturbances that trigger rainfall anomalies at Alto Chetilla rain gauge do not reach Huaquillas. Some kind of decoupling of these rain gauges is observed in the wind speed data record as well. At the low-elevation rain gauge at Huaquillas, only westerly winds appear to reach a mean wind speed > 2 ms^{-1}, whereas at the high-elevation rain gauge of Alto Chetilla the highest wind speed – with a mean maximum up to 15 ms^{-1} – is observed for easterly winds and thus for descending winds (Fig. 9). However, these gusty down-slope winds subside before they reach Huaquillas.

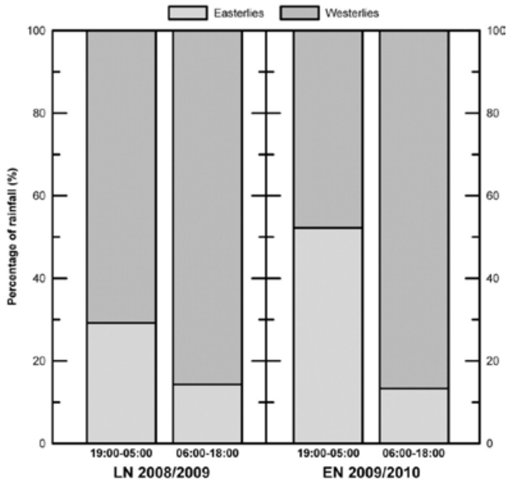

Figure 8. Seasonal distribution of prevailing wind direction for diurnal (06:00 to 18:00 LT) and nocturnal (19:00 to 05:00 LT) rainfall at rain gauge Alto Chetilla.

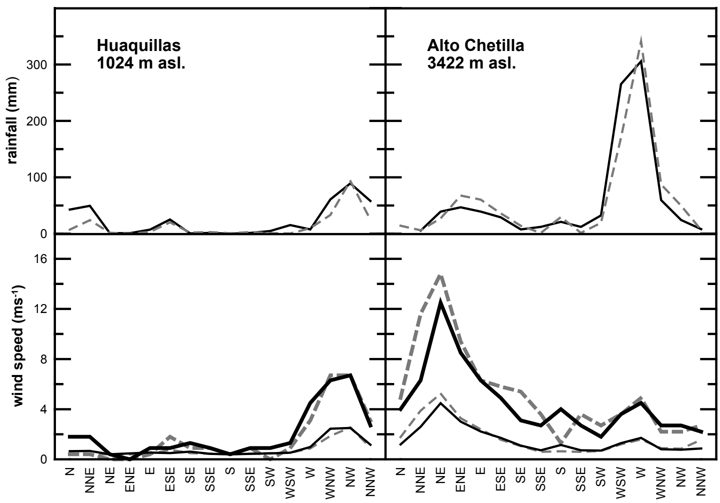

Figure 9. Cross-correlation of wind sectors and of rainfall, mean wind speed (thin line) and mean maximum wind speed (thick line) for Huaquillas and Alto Chetilla rain gauges for LN (solid line) and EN (dashed line) episodes of 2008/2009 and 2009/2010.

We assume that high-elevation areas in the Pacific realm close to the CWD, such as Granja Porcón (3180 m asl) and Alto Chetilla (3422 m asl), are benefiting from nocturnal rainfall due to the eastward penetration of MCCs. The surplus in rainfall due to easterly influence partly offsets the rainfall deficit. However, the spatial extent of this wet pulse and thus the positive rainfall anomaly seems to be limited to areas located higher than 3000 m asl.

6. Discussion and conclusions

In this study we analysed rainfall variability in the region of Cajamarca, focusing on EN/LN episodes of 2005 to 2010. We found that rainfall variability is due to local scale features such as topography and/or local wind fields and also to large scale circulation features such as ENSO or MCCs. The effect of local topography and wind fields on precipitation patterns is more apparent, whereas the influence of ENSO and MCCs appears to be more subtle.

Topography, represented by shape and aspect of valley systems and valley flanks, determines the main trajectories of moist air masses by canalizing valley wind systems. Poor spatial correlation of rainfall totals indicates that spatial and temporal distribution of rainfall varies for each valley, depending on its exposure to main atmospheric flow. Similar observations have been made by Buytaert *et al.* (2006) and Rollenbeck and Bendix (2011) when studying rainfall distribution in the Andes of southern Ecuador. In the region under investigation, the Pacific realm is topographically characterized by the steep valley morphology of the river Jequetepeque, whereas the Atlantic realm is characterized by the smoother morphology of the inter-Andean valley of Cajamarca. In the Pacific realm, correlation coefficients of daily and monthly rainfall are higher than in the Atlantic realm, thus implying that west of the CWD the far distant localities underlie similar forcing, whereas localities east of the CWD are substantially influenced by small scale local weather anomalies generated by thermal overheating and local wind systems.

A very prominent feature of rainfall variability within the Pacific realm is rooted in the diurnal march of the valley breeze system. During day-time, thermally induced up-slope breezes develop, whereas during night-time easterly down-slope breezes prevail owing to nocturnal radiative cooling. At higher levels, winds transport air from the east across the CWD towards the Pacific, thereby strengthening down-slope breezes (Howell, 1953). However, rainfall is mainly associated with up-slope breezes that evolve during the forenoon and finally meet, oppose and gradually overcome the easterly winds near the CWD. If these westerly winds encounter easterlies that are stable and warm, stratocumulus and cumulus cloudiness may develop, but rainfall of more than drizzle intensity does not occur (Howell, 1953). If the easterlies are unstable, however, or cool enough in comparison with the Pacific air, atmospheric instability causes heavy rainfall to occur. More recently, Rollenbeck and Bendix (2011), when studying rainfall distribution in southern Ecuador, showed that torrential rainfall occurs if cloud systems originating in the west converge with easterly flow.

Although the diurnal cycle of this motion is evident, not all areas within the Jequetepeque valley are affected in the same way and some kind of decoupling is observed. This is related to the fact that cooling is sufficiently strong to cause down-slope winds only during the latter part of the night and often only at higher elevations where net outgoing radiation is greater (Howell, 1953). These descending air masses do not reach the Pacific coast but override the body of air that has been brought in from the Pacific the previous day and has not been subject to strong down-slope movement. Howell (1953), who studied local weather in the catchment of Chicama, a westward draining river in northern Peru, named this low-level air body the Pacific intermediate layer. This layer is bounded by westerly air masses of Pacific origin at the bottom and by easterly air masses that descend from higher altitudes at the top. Howell (1953) found the boundaries to be located at altitudes between 750 and 900 m asl. In consequence, the meteorological data of Huaquillas rain gauge (1034 m asl), which is roughly located at the altitude of the lower boundary, differ significantly from the data of higher elevation rain gauges.

As shown in this study, rainfall variability within the Atlantic realm is rooted in a complex interrelation of easterly air flow dynamics, thermally induced diurnal wind systems and frequent spillover of Pacific air masses, resulting in a reorganization of local wind fields and an enforcement of convective activity. The CWD, defined as a hydrological barrier, acts as a climatological barrage. On the one hand, rainfall totals within the Atlantic realm are significantly lower owing to lee effects caused by summits higher than 4000 m asl, which encircle the inter-Andean basin of Cajamarca and hinder moist air masses from spreading and precipitating large quantities of rainfall. On the other, the CWD plays a major role as a mixing zone characterized by overlapping western and eastern disturbances. Overriding westerly winds of Pacific origin cause a reversal of the rain-bearing, thermally induced up-slope winds over the western flanks of the inter-Andean valley of Cajamarca. We therefore conclude that the complex spatial and temporal pattern of rainfall variability in the Atlantic realm is due to local scale anomalies in wind fields, triggered by westerly low-level flow originating in the Pacific realm.

ENSO, with its extreme phases of EN and LN, is superimposed on rainfall variability due to the regional forcing of the valley breeze system and the local forcing of wind reversal or locally enhanced convection. Although direct effects of ENSO on high Andean rainfall variability are in debate, it seems that several fingerprints of EN can be found in the region under investigation.

Most rain gauges used in this study reveal lower than normal rainfall during EN episodes and normal to above-normal rainfall during LN. The LN/wet and EN/dry signature has been noted before by other authors when studying rainfall variability over the Altiplano, the eastern Andean slopes or the nearby La Encañada catchment (Tapley & Waylen, 1990; Vuille, 1999; Garreaud & Aceituno, 2001; Ronchail & Gallaire, 2006; Romero *et al.*, 2007). Vuille (1999) and Garreaud and Aceituno (2001) proposed as a mechanism for the rainfall anomaly that low rainfall is related to a relative increment in westerlies, hindering advection of moist air from the Amazonian region. This might be true for the region of Cajamarca as well, but as the wind field analyses we have conducted are restricted to low-level winds and are biased by local scale wind anomalies, no conclusions regarding the high-level flow can be drawn. However, we found that within the area under investigation two rain gauges (Alto Chetilla, 3422 m asl, and Granja Porcón, 3180 m asl) contradict the EN/dry scheme. These rain gauges are both located in the Pacific realm and at altitudes higher than 3000 m asl. Our findings suggest that these high-elevation rain gauges benefit from large scale instabilities, which originate east of the CWD, and that abundant nocturnal rainfall partly compensates for the rainfall deficit. However, the spatial extent of these wet pulses seems to be limited to areas located above 3000 m asl. Zolman *et al.* (2000) compared the number, size and intensity of tropical MCSs for the EN period of 1993 and the LN period of 1995. They found that during EN the number of MCSs and the area covered by them is higher than during LN. However, their data solely for the DJF period are less conclusive, so that higher numbers of MCSs are observed over the eastern Pacific, whereas slightly lower numbers are observed over the tropical South American region. We therefore propose that the EN/dry and LN/wet scheme is valid for the region of Cajamarca, but that deviation from that scheme must be taken into account. Especially high-elevation regions on the western escarpment may benefit from atmospheric instabilities brought in by easterlies which result in abundant nocturnal rainfall and thus substantially modify the seasonal rainfall budget.

Another finding is that a modulation of the regional precipitation pattern occurs during EN, so that the secondary rainfall minimum is delayed for one or two months, irrespective of elevation. It is fairly well known that bimodal rainfall distribution in the tropics is connected with the seasonal displacement of the ITZC (Bendix and Lauer, 1992). Caviedes and Endlicher (1989) argue that during EN the ITCZ does not migrate uniformly but periodically, as seen in the occurrence of pulses of warm water, cloudiness or rainfall. The atmospheric instability that results from increased SST off the western coast of tropical South America and the pulsating march of the ITCZ during EN phases might be the reason for the observed skewed bimodal rainfall distribution and the delay of the secondary rainfall minimum. However, to propose a mechanism for this observed anomaly is beyond the scope of this study.

A striking feature of the 2009/2010 EN period is the positive temperature anomaly of 0.9°–1.5°C compared to the preceding rainy season. Positive temperature anomalies in the Andean highlands due to EN have been noticed before by several authors (Aceituno, 1988; Vuille *et al.*, 2000; Francou *et al.*, 2004). We found that, these temperature anomalies enforce atmospheric turbulence and cause heightened wind speeds. As outlined in this study, wind activity and convective activity are interrelated by the penetration of low-level westerly winds into the valley of Cajamarca.

However, the thermally induced valley breezes on both sides of the CWD are affected by higher temperatures. Up-slope winds within the Pacific realm may cross the ridges of the CWD and deflect the up-slope breeze within the Atlantic realm. By doing so, they enhance convection over the valley floor of Cajamarca. Convection results in the buildup of cloud systems, which in turn reduce thermal heating by reflecting incoming solar radiation. As a result the maximum of convective activity, as seen in rain abundance, occurred in the early afternoon in contrast to the late afternoon maximum observed during LN 2008/2009. This mechanism–triggered by temperature anomalies, strengthened by wind anomalies and finally resulting in earlier convection–may be partly responsible for lower rainfall during EN, as earlier convection in combination with cloudiness may diminish overall thermal overheating of the valley of Cajamarca and thus ultimately diminish total rainfall.

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