Annual and inter-annual variability of the present climate in northern South America and southern Mesoamerica

Germa´n Poveda a,*, Peter R. Waylen b, Roger S. Pulwarty c

a Escuela de Geociencias y Medio Ambiente, Universidad Nacional de Colombia, Medellı´n, Colombia
b Department of Geography, University of Florida, Gainesville, FL, USA
c Cooperative Institute for Research in Environmental Sciences/Climate Diagnostics Center, University of Colorado, Boulder, CO, USA

Received 21 September 2004; accepted 20 October 2005

Abstract

Present climate of northwestern South America and the southern Isthmus is detailed in terms of major hydro-climatic controls, supported by evidence from station records, reanalysis data and satellite information. In this tropical region, precipitation is the principal hydro-climatological variable to display great variability. The primary objective is to view the controls that operate at intra-seasonal to inter-decadal time scales. This is a topographical complex region whose climate influences range in provenance from the South Atlantic to the Canadian Prairies, and from the North Atlantic to the Eastern Pacific. The situation is further complicated by interactions and feedbacks, in time and space, between these influences, which are interconnected over various scales. The greatest single control on the annual cycle is the meridional migration of the Inter-tropical Convergence Zone and its pattern of associated trade winds. Consideration of these alone and their interaction with the Cordilleras of the Andes and Central America produce a variety of unimodal and bimodal regimes. Regionally, two low level jet streams, the westerly Choco jet (5°N) and the easterly San Andrés jet (12–14°N), and their seasonal variability, have tremendous significance, as do mesoscale convective storms and mid-latitude cold fronts from both the northern ("nortes") and southern ("fríagems") hemispheres. There are many examples of hydro-climatological feedbacks within the region. Of these the most notable is the interaction between evaporation over the Amazon, precipitation onto the eastern Andes and streamflow from the headwaters of the Amazon. This is further compounded by the high percentages of recycled precipitation over large areas of the tropics and the potential impacts of anthropogenic modification of the land surface. The El Niño-Southern Oscillation phenomenon (ENSO) is the greatest single cause of interannual variability within the region, yet its effects are not universal in their timing, sign or magnitude. A set of regional physical connections to ENSO are established and their varying local manifestations are viewed in the context of the dominant precipitation generating mechanisms and feedbacks at that location. In addition, some potential impacts of longer run variations within the ocean-atmosphere system of the Atlantic are examined independently and in conjunction with ENSO. This review of the climatic controls and feedbacks in the region provides a spatial and temporal framework within which the highly complex set of factors and their interactions may be interpreted from the past.

© 2005 Elsevier B.V. All rights reserved.

Keywords: Climatology; El Niño-Southern Oscillation; Northern South America; Choco jet; San Andres jet

1. Introduction

This paper reviews the present climate of northern South America and southern Central America, mainly
focusing on precipitation and significant atmospheric circulation features and their variability at annual and interannual time scales. The region covers the northern tropical Americas, (equator—15°N, and 60°W–85°W) encompassing Costa Rica, Panama, Colombia and Venezuela. It abuts the tropical North Atlantic, Caribbean and eastern tropical Pacific. Regional atmospheric circulation and climate arise from interactions with the neighboring oceans, the Andes, the Cordillera of Central America, the Venezuelan–Colombian Llanos and the Amazon basin. The classic work of Snow (1976) provides an excellent initial understanding of regional climatology. It is the intent of this paper to detail the complexity and interaction of the factors governing climate, at various time and space scales, over the region through the use of recent data, including station records, climatic reanalysis and satellite data. Hence, one goal of this paper is to more fully analyze and describe the regional climates of the meso-Americas and their varying sensitivities to allow paleoclimate researchers to better reconcile modeling studies, proxy datasets and present-day observations.

The principle controlling factors will be shown to be the Inter-tropical convergence zones (ITCZ) of both the Atlantic and Eastern Pacific Oceans and the trade winds, the Hadley cell circulation, and interactions occurring between land surface and atmosphere at distinctive time scales. These in turn govern the strength and location of low level jets in the area and the propensity for the generation, path and penetration of such transient features as cold fronts and tropical storms. The excellent long paleo-sedimentary record available (14,000 yrs.) from the Cariaco Basin off the north coast of Venezuela has been the source of much speculation concerning variations in the position of the ITCZ, the strength of the trade winds, contrasts in Atlantic sea surface temperatures and rainfall on the contributing contiguous continental margin (e.g., Haug et al., 2001; Lea et al., 2003). El Niño-Southern Oscillation (ENSO) will also be shown to be an important part of interannual climatic variations over broad areas of the circum-Pacific region for millennia (e.g., Rodbell et al., 1999).

Section 2 reviews the annual climatological cycle, focusing mainly on precipitation patterns associated with the migration of the ITCZ, and discussing important physical mechanisms interacting within the ITCZ regionally. The roles of mesoscale convective systems (MCSs), important feedbacks between the Andes and the Amazon, and land surface–atmosphere interactions are also discussed. Section 3 deals with interannual variability, mostly associated with the extreme phases of ENSO, but also with ocean–atmosphere activity over the North Atlantic Ocean. The effects of ENSO are thoroughly discussed, and a summary of regional physical mechanisms associated with the occurrence of El Niño is presented. Final remarks are provided in Section 4.

2. Annual cycle

The annual hydro-climatic cycle is dominated by the meridional migration of the ITCZ, which in turn controls the various dynamics of the trade winds over oceans and land masses. However, the important role of the land–atmosphere system in shaping the climate of the region has been largely overlooked. Diverse regional and local atmospheric circulation patterns interact with the ITCZ and its migration to modify the annual cycle of precipitation. Among these, we identify two low-level jets, one easterly over the Caribbean around 10°–12°N (Poveda and Mesa, 1999; Magaña et al., 1999; Amador and Magaña, 1999), and the other, the Choco jet, westerly at about 5°N over the Pacific (Poveda and Mesa, 1999, 2000). The former is linked to two unique features of precipitation on the isthmus; a boreal winter maximum on the Caribbean flank and a “mid-summer drought” throughout the Pacific flank. The latter is associated with MCSs (Velasco and Frisch, 1987; Mapes et al., 2003; Zuluaga and Poveda, 2004). A third easterly jet is confined to the 600–700 hPa level over South America and the eastern equatorial Pacific (Hastenrath, 1999).

In deep tropical America, the activity of MCSs becomes a significant source of prolonged, intense precipitation events. The study of Velasco and Frisch (1987) pioneers the understanding of the dynamics and thermodynamics of MCSs over the region. Recent data from the Tropical Rainfall Measuring Mission (TRMM, Kummerow et al., 1998) permit a deeper understanding of the annual precipitation cycle. A considerable portion of the study region is occupied by the upper Amazon basin. The relevant climatic roles that Amazonia plays at both the planetary and regional scales are discussed, drawing attention to land surface–atmosphere interactions in the creation of annual and inter-annual variability in the study region.

This review purposely ignores other mechanisms and atmospheric circulation patterns associated with climatic variability over the region, including the dynamics of tropical easterly waves, squall lines over the Amazon basin, the South American low level jet (Byerle and Paegle, 2002; Campetella and Vera, 2002), and the intra-seasonal variability associated with the activity
of the 30–60 day oscillation. All of them need to be understood in terms of their modulation and coupling to the annual and inter-annual time scales.

2.1. Meridional migration of the ITCZ in the region

There is no single, universally held definition of the ITCZ. The classical view defines it as a region close to the equator of trade wind convergence, ascending air, low atmospheric pressure, deep convective clouds and heavy precipitation (Henderson-Sellers and Robinson, 1986). In turn, Hartmann (1994) defines it as the axis of the broad trade wind current of the tropics where the northerly and southerly trade winds meet, either in narrow bands or in the broader convergence zones over South America, Indonesia and Africa. Gu and Zhang (2001) describe it as a zonally elongated (i.e., parallel to the equator), latitudinally confined, rain or cloud band, composed mainly of non-propagating, random deep convective clouds and non-deep clouds. Hastenrath (2002) invokes a suite of variables including low surface pressures, negative values of divergence of horizontal wind, high rainfall rates and low values of Outgoing Longwave Radiation (OLR). Regional positions of the ITCZ and the concomitant location of maximum rainfall can be traced in the seasonal distribution of OLR. Assuming that disturbances in the dynamic fields are associated with deep convective activity (Gu and Zhang, 2001), Fig. 1 shows the extremes of the seasonal cycle of OLR, using data derived from the 40 year Reanalysis Project (Kalnay et al., 1996) of the National Center for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR). The quantity of OLR is inversely related to deep convective cloudiness, wave motions in the tropics and higher rainfall rates (Gruber, 1974; Nitta et al., 1985; Hendon and Liebmann, 1991; Wheeler and Kiladis, 1999).

The ITCZ has been explained in terms of synoptic-scale disturbances in the circulation, many of which propagate westward (Holton et al., 1971; Chang, 1973; Lindzen, 1974; Hess et al., 1993; Gu and Zhang, 2001, 2002), although other hypotheses have been put forward (e.g., Pike, 1971; Philander et al., 1996; Tomas and Webster, 1997; Ferreira and Schubert, 1997). The meridional oscillation of the ITCZ responds to the seasonal insolation cycle, lagging the zenithal position by approximately one month. This oscillation exerts a strong regional control on the annual hydroclimatological cycle. During the austral summer the continental ITCZ is located over the Amazon basin, southern Colombia and Ecuador, while those portions over the eastern equatorial Pacific migrate no further south than 3°N (Alpert, 1945, Hastenrath, 1966, 2002). Northward displacement occurs in the boreal summer, covering the upper Amazon and Orinoco basins, the Guianas, Colombia, and the Central America isthmus, as far north as the Costa Rica/Nicaragua border. Central Colombia and the western Andean cordilleras experience a bi-modal annual cycle with peaks during April–May and October–November, and less rain in December–February and June–August, mainly as the result of the double passage of the ITCZ (Figs. 2 and 3). A uni-modal cycle (May–October) is witnessed over the Caribbean coast of Colombia and the Pacific flank of the southern isthmus, reflecting the northernmost position of the ITCZ over the continent and eastern equatorial Pacific, respectively (Portig, 1965; Hastenrath, 1967, 1990, 1991, 2002; Waylen et al., 1996a). A slight reduction in July–August precipitation, known locally as the “Veranillos de San Juan”, “Canícula”, or mid-summer drought (Magaña et al., 1999) is present throughout the Pacific flank of Central America. Its causes are complex, but its effects have been shown to be modulated by the east Pacific “warm pool” bordering Central America, which attains maximum temperatures during this time, and to be associated with the development of a low-level jet at 15°N over the Caribbean (Pulwarty et al., 1998). A large-scale movement of deep convection way from northern South and Central America has been observed synchronous to the drought (Magaña et al., 1999). A single peak is also evident over the eastern piedmont of the Eastern Andes, mainly due to the advection of moisture from the Amazon which encounters the orographic barrier of the Andes, thus focusing and enhancing deep convection and rainfall on the eastern flank of the cordillera, with maximum rainfall occurring during June–August.

Unlike the major cordilleras of Costa Rica and the Andes, the mountains of Panama run from west to east. Their comparatively low relief and the narrowness of the southern isthmus, allow complex influences of moisture-laden winds from circulations above both the Caribbean and Pacific. The Caribbean slope exhibits an almost uniform precipitation regime, with minimum values of 100 mm/month. The Pacific slope experiences a rainy season from late April–early May to late November, and a dry season in December–April, during which 2 maxima are witnessed in June and October, as a response to the ITCZ migration.

Venezuela can be divided into three major climatic zones: (1) the north and west, comprising the Maracaibo basin, the east–west orientated Cordillera de la Costa (<1500 m high), and the prominent southwest–
northeast oriented Cordillera de Merida (up to 5000 m high); (2) the central area, known as the “Llanos”, a vast, low-lying basin; and (3) the Guyana Highlands in the southeast, with elevations between 400 and 1400 m (Pulwarty et al., 1992). Variability of precipitation in northern South America, and Venezuela, in particular, is made more complex by the geography of the region. The north–south trending cordillera and the presence of the major moisture source (Caribbean) to the north of the landmass, produce very different conditions from the classic boreal “monsoon” climates of South Asia and West Africa (Pulwarty et al., 1992). Nevertheless, a narrow equatorial trough can be identified in lower tropospheric wind shifts, zonal wind divergence, and pressure height deviations around 5°N. Bimodal rainfall distributions are widespread at stations west of 70°W, in sharp contrast to the July–August maximum of the Llanos. Further east, along the coastal zone, rainfall is more directly governed by the migration of the ITCZ. Remnants of old cyclonic systems and cold fronts (frigams) from Brazil and Argentina may yield heavy precipitation events over the Llanos (5°–10°N).
and occasionally into the Caribbean (Pulwarty et al., 1992).

Bimodal maxima (usually April–May and September–October) can be sub-divided into four subtypes. Coastal, low annual totals of about 430 mm and July August and November December maxima: Northwest lowlands, annual totals of 1000 mm with a larger peak in fall and a secondary one in spring; Venezuelan Andes, higher totals and both fall/spring peaks; and Eastern Guyana Highlands, 1250–1300 mm annually, a primary early boreal summer maximum and a secondary one in late fall. Orographic controls strongly amplify the July–August minimum in the Venezuelan Andes and sharpen the transition between unimodal and bimodal regimes (7.4–8.6°N, 71.5–72.5°W).

2.2. Regional mechanisms interacting with the ITCZ

2.2.1. The Choco low-level jet

Ocean–atmosphere–land interactions over the eastern Pacific have considerable influence on Colombia’s hydro-climatology. It has also been suggested that the Andes influence rainfall off Colombia’s Pacific coast through thermally driven gravity waves (Mapes et al., 2003). The Pacific lowlands are among the world’s rainiest regions, with mean annual precipitation in excess of 8000 mm and even beyond 12000 mm (e.g., López, 1966; Arnett and Steadman, 1970; Snow, 1976; Nieuwolt, 1977; Eslava, 1993; Poveda and Mesa, 1997, 2002; Mapes et al., 2003). These copious totals arise from the interaction of a low level tropical westerly, Choco jet from the Pacific, with the dynamics of mesoscale convective systems (Poveda and Mesa, 2000).

The Choco jet, thermally driven by the SST gradient between the Ecuador–Peru cold tongue and the Colombian Pacific, is further favored by the change of direction of the cross-equatorial austral trades which become westerly at this latitude (Poveda and Mesa, 2000; Hastenrath, 2002), and by the latent heat release of rainfall in MCSs over the region which favors surface convergence. The Choco jet is clearly visible in the seasonal cycle of airflow at 925 hPa (Fig. 4), transporting large quantities (mean rate of $3.78 \times 10^6$ kg s$^{-1}$ or $3774$ m$^3$ s$^{-1}$) of moisture inland where it contributes directly to the combined annual discharges (~5000 m$^3$ s$^{-1}$) of the Atrato and San Juan rivers, which exhibit among the world’s highest runoffs (Poveda and Mesa, 1999, 2000). The Choco jet is forced to ascend the topography of the western Andes and to interact with the
easterly trade winds, thereby further enhancing deep convection. Its seasonal strengthening (September–November) and weakening (February–March), partially explains why the October–November rainy season is more intense than that of April–May over central and western Colombia. At paleoclimatic time scales, there is evidence for the presence of La Niña-like conditions in the eastern equatorial Pacific and a stronger Choco jet in the northern Andes during the last glaciation (Martínez et al., 2003).

Fig. 5 depicts the October mean meridional–vertical cross-section of zonal winds at 80°W, between 5°S and 20°N. The Choco jet, is confined below 800 hPa and focused around 5°N at 950–900 hPa, with maximum wind speed of 7–8 ms⁻¹. Two easterly jets, the San Andrés (14°–16°N, and 900 hPa; Magaña et al., 1999; Poveda and Mesa, 1999), and the Equatorial Mid-tropospheric Easterly Jet (EMTEJ) at 600 hPa, centered on the equator (Hastenrath, 1998; 1999), show up as negative values and are discussed below.

2.2.2. San Andrés and the equatorial mid-tropospheric equatorial jets

The low level jet in the Caribbean trade winds (Poveda and Mesa, 1999; Amador and Magaña, 1999;
Fig. 4. Seasonal cycle of horizontal wind over the tropical Americas at 925 hPa. Note the southwesterly Choco jet moving from the Pacific inland at around 5°N. The easterly San Andrés jet (Section 2.2.2) can be found around 12°–14°N. Data source: NCEP-NCAR Reanalysis.
Gañana et al., 1999), acts as a link between the eastern Pacific and the Caribbean on interannual and shorter time scales. Its activity over the Caribbean has potential effects on tropical convective systems, such as easterly waves, and their organization into tropical cyclones. Core velocities (13°N–15°N, below 900 hPa) are high in July–August (12 m s\(^{-1}\)) and December–February (10 m s\(^{-1}\)), and least during September–November (6 m s\(^{-1}\)) and March–May (8 m s\(^{-1}\)).

The EMTEJ exhibits an annual cycle (Table 1) which is almost the inverse of the Choco and San Andrés jets: more intense during March–May and June–August, with core velocities larger than 10 m s\(^{-1}\), and weaker core velocities during September–November (6 m s\(^{-1}\)) and December–February (8 m s\(^{-1}\)). The EMTEJ experiences almost no meridional oscillation, being located over the Equator during March–May and over 2°S during September–November.

After crossing the isthmus, part of the San Andrés jet deviates to the southeast, feeding the Choco jet in conjunction with downstream portion of the cross-equatorial flow from the southern hemisphere. As a result, the San Andrés jet is weaker at 85°W than at 80°W. The physics of the San Andrés jet is still elusive, but probably it is associated with the dynamics of the tropical Western Hemisphere warm pool (WHWP), a region that extends over parts of the eastern North Pacific, the Gulf of Mexico, the Caribbean, and the tropical North Atlantic (Wang and Enfield, 2001, 2003; Wang, 2002, 2003).

### Table 1
Seasonal cycle of average wind velocities (m s\(^{-1}\)) at the core of the three jets discussed in the text

<table>
<thead>
<tr>
<th></th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
<th>DJF</th>
</tr>
</thead>
<tbody>
<tr>
<td>EMTEJ jet</td>
<td>−10 (0)</td>
<td>−10 (0)</td>
<td>−6 (0)</td>
<td>−8 (0)</td>
</tr>
<tr>
<td>San Andrés jet</td>
<td>−8 (1)</td>
<td>−12 (−4)</td>
<td>−6 (−2)</td>
<td>−10 (2)</td>
</tr>
<tr>
<td>Choco jet</td>
<td>2 (0)</td>
<td>4 (−1)</td>
<td>6 (−2)</td>
<td>4 (−2)</td>
</tr>
</tbody>
</table>

Negative figures indicate easterly wind direction. Figures in parentheses represent computed changes in mean core wind velocities warm phase years minus cold phase years of ENSO.

#### 2.2.3. Mesoscale convective systems (MCSs)

These systems (MCSs) deliver large quantities of moisture to the region (Fig. 6). Velasco and Frisch (1987) hypothesize that their formation and dynamics are related to the development of warm nucleus vortices of mesoscale size, forced by latent heat. Most are nocturnal and continental. The systems are defined in terms of black body infrared temperatures (−40 and −62 °C), and a required continental area in the order of 50,000 km\(^2\) (Maddox, 1980), which may drop to 30,000 km\(^2\) at low latitudes. The development and evolution of MCSs is closely linked to the activity of low-level jet streams and deep convection (Bonner, 1966; Raymond, 1978; Maddox, 1980; Lemaitre and Brovelli, 1990; Stensrud, 1996).

The role of the Choco jet in the development of numerous MCSs in the Panama Bight is paradigmatic (Velasco and Frisch, 1987; Poveda and Mesa, 2000), and both exhibit a strong diurnal cycle (Mejia and Poveda, 2005; Mapes et al., 2003; Poveda et al., 2005).
The annual cycle of MCSs and their atmospheric environments during 1998–2002 are studied by Zuluaga and Poveda (2004), using satellite information from the Tropical Rainfall Measuring Mission (TRMM, Kummerow et al., 1998). On average, a total of 44,709 rainfall events occurred, of which 19,875 (44.4%) and 24,834 (55.6%) were over land masses and oceans, respectively. Their monthly distribution and intensity for 1998 are shown in Fig. 6. A low number of systems actually develop into MCSs (4929), 2761 over oceans and 2168 over land, yet their contribution to total annual precipitation is disproportionate, with an average of 70% during the 1998–2002 period. The distribution of events developing MCSs is strongly controlled by the position of the ITCZ, yet during 1998 (warm phase ENSO) the most favorable region for MCSs was located over western Colombia and the adjoining Pacific. The Magdalena river valley, northern Antioquia, and the Amazon and Orinoco regions of Colombia also exhibit high frequencies of MCSs. Least favorable regions are northern Colombia, Venezuela, and western Peru. MCSs appear more frequently off the Pacific coast of the isthmus than the Caribbean, but there is a notable dearth of events in July and August (midsummer drought) in these Pacific waters, and a slight preponderance of the limited number of December events in Caribbean waters. Events are almost totally absent from the isthmus during January–April. The studies of Mapes et al. (2003) and Warner et al. (2003) provide diagnostics and a modeling framework for rainfall over western Colombia and off the Pacific coast.

Fig. 6. Monthly distribution of mesoscale convective systems (MCSs) over the tropical Americas, during 1998, according to criteria by Nesbitt et al. (2000). The size of the triangle is correlative with rainfall intensity within the MCS’s (Zuluaga and Poveda, 2004). Data source: Satellite sensors of the Tropical Rainfall Measuring Mission (TRMM).
atmosphere system, due to its large area (>6.4 x 10^6 km^2), tropical setting, and complex eco-hydro-climato-
logical dynamics which exert a global influence (e.g., Avissar and Nobre, 2002; Roberts et al., 2003). Mod-
eling results suggest changes in global, regional and local atmospheric circulation patterns associated with
deforestation or perturbations in the land surface–atmos-
sphere system (Shuttleworth, 1988; Zeng and Neelin, 1999). Zeng et al. (1996) propose that deforestation
would reduce the SST gradient over the tropical Atlant-
ic, thereby affecting global atmospheric circulation
through perturbations of Walker and Hadley cells, which
is confirmed by the studies of Zhang et al. (1996) and Werth and Avissar (2002). Relevant features
of the atmospheric circulation over the Amazon are studied by a number of authors (e.g., Kousky, 1985;
Kousky and Kagano, 1981, 1994). Major factors affect-
ing the Amazon basin include the meridional oscillation
of the ITCZ and the South Atlantic Convergence Zone
(SACZ, Kousky, 1988), which is defined as a persistent
cloud band oriented in the northwest–southeast direc-
tion, and encompassing southern Amazonia through to
the South-Central Atlantic, being most prevalent during
summer. The presence of the Andes affects consider-
dably the climate of Amazonia because they exert a
strong influence on air masses distributions, constitute
an orographic barrier for the easterly winds, and funnel
tropical maritime air east of the Andes towards the
southeast.

Land surface–atmosphere interactions constitute sig-
nificant drivers of the climate and weather variability of
Amazonia. The important role of precipitation in con-
trolling vegetation activity is reflected by the Normal-
ized Difference Vegetation Index, NDVI (Poveda and
Salazar, 2004). Evapotranspiration (ET in Fig. 7) from
vegetation is itself a key control on precipitation recycl-
ing (Pr in Fig. 7; Salati, 1985; Eltahir and Bras, 1994;
Trenberth, 1999). Over the Amazon, precipitation exhi-
bits a uni-modal annual cycle, with a wet season during
December–April and a dry season in June–October
(Obregon and Nobre, 1990; Zeng, 1999). According
to Horel et al. (1989) the establishment of the rainy
season over the Amazon basin occurs relatively fast,
taking one month or less, due to dynamic and thermo-
dynamic mechanisms. Conversely, the cessation of the
rainy season occurs more slowly.

The seasonal march of OLR depicted in Fig. 1
shows differences in the behavior of the ITCZ and
the SACZ over land and ocean regions, due to season-
ally varying land–atmosphere interactions and bound-
ary layer dynamics, which, by means of soil moisture
and evapotranspiration, control precipitation and weath-
er patterns over tropical land masses (Delworth and
Manabe, 1993; Elthair, 1998). Moist deep convection,
the predominant regional mechanism for intense
storms, is partially controlled by the dynamics of the
atmospheric boundary layer and land surface–atmo-
sphere interactions. Thus the seasonal cycle of rainfall
over the Amazon and adjoining regions reflects the
combined effects of large-scale forcing such as the mi-
gration of the ITCZ, moisture transport from the Atlan-
tic Ocean (Pa in Fig. 7), and land surface–atmosphere
interactions, including precipitation recycling.

Feedbacks are also present in the Andes–Amazon
bio-geophysical system (Fig. 7). The tropical Andes
constitute the Amazon headwaters, simultaneously
importing atmospheric water from the Amazon, and
exporting surface waters (Q), sediments (S), nutrients
(N), and constituents and pollutants (C) to it. This
feedback is most active during May–September, when

![Fig. 7](image-url)
the moisture laden southeast trades (Vq), and their orographic lifting over the eastern Andes, contribute to convective precipitation (P). The interactions and feedbacks are felt: (1) “locally” through precipitation recycling within the Amazon, (2) “regionally” by means of fluxes between the Amazon and Eastern Andes, and (3) “hemispherically” through the strengthening/weakening of Hadley cell circulation (Werth and Avissar, 2002). They also indicate the potentially important anthropogenic control on deforestation in both the Amazon and the Andes may exert on the functioning of the system.

Recent studies over the Amazon basin (Nesbitt et al., 2000; Petersen et al., 2002) indicate that wet-season convection exhibits a vertical structure that is intermediate to that observed over tropical oceans (less vertical development) and other continents (more lightning, larger vertical development). The convection also displays westerly and easterly modes (Cifelli et al., 2002), which are correlated to changes in 850–700 hPa zonal wind direction, and to the dynamics of the South American low level jet (Byerle and Paegle, 2002; Campetella and Vera, 2002). Each mode promotes different vertical structures and precipitation statistics (Petersen et al., 2002). The easterly one favors westward propagating precipitation as Amazonian squall lines (Greco et al., 1990; Cohen et al., 1995; Warner et al., 2003), exhibiting a strong diurnal cycle and cross the basin in 2–3 days. The rainbands appear to involve both moisture advection and ducted gravity wave dynamics in the presence of the EMTEJ (Poveda and Mesa, 1997; Warner et al., 2003). Also, during the easterly regime, MCSs and rain cells exhibit closely related propagations, mostly associated to the midlevel mean flow (Laurent et al., 2002), and enhancement of MCSs convective cells, and more intense in terms of kinematic and microphysical features, comparatively to the westerly regime (Carvalho et al., 2002; Cifelli et al., 2002). Instead, during this latter regime, the propagation of both MCSs and rain cells appear to be more disorganized, with apparently no clear relationship with the mean atmospheric flow at any level (Laurent et al., 2002).

3. Interannual variability

3.1. ENSO

ENSO is the main forcing mechanism of climatic variability in tropical South America from seasons to decades. El Niño refers to the unusual warming of sea-surface temperatures (SSTs) over the eastern and central tropical Pacific. Important components of this anomaly are the deepening of the oceanic thermocline in the eastern Pacific, and the weakening of the dominant surface easterly trade winds. During an El Niño event there is a shift in the center of convection from the western to the central Pacific. The accompanying Southern Oscillation, the “seesaw” of the atmospheric mass that produces a pressure gradient between the western and the eastern equatorial Pacific, is characterized by the Southern Oscillation Index (SOI). Negative values of the SOI are associated with warm events (El Niño), while positive values accompany cold events (La Niña) and a stronger Walker cell (west–east) circulation. ENSO is an aperiodic oscillation with a recurrence interval of between two and ten years, and an average of four (Trenberth, 1991), which appears to have become more frequent since the mid 1970’s (Trenberth and Hoar, 1996). El Niño events begin during the boreal spring and exhibit strong phase locking with the annual cycle (Tziperman et al., 1994; Chang et al., 1994; Webster, 1995; Neelin et al., 1998). Those events, which encompass two calendar years, are generally characterized by increasing SST anomalies during boreal spring and fall of the first year (year 0), which peak in winter of the following year (year +1). Anomalies then decline in spring and summer of year +1. Kiladis and Díaz (1989) recognize the following years of El Niño event during the past century: 1902, 1904, 1911, 1913, 1918, 1923, 1925, 1930, 1932, 1939, 1951, 1953, 1957–1958, 1963, 1969, 1972–1973, 1976–1977, 1982–1983, and 1986–1987. More recently, El Niño occurred in 1991–1992, 1994–1995, 1997–1998, 2002–2003, and 2004. La Niña events occurred in 1949, 1954, 1964, 1970, 1973, 1975, 1988, and 1998. Rossel (1997) provides a thorough review of the diverse classification schemes for phases of ENSO.

3.2. ENSO at paleoclimatic time scales

ENSO has been an important part of interannual climatic variations over broad areas of the circum-Pacific region for millennia (Rodbell et al., 1999). Several recent studies focus on the reconstruction of the paleoclimatic ENSO based on instrumental records (Whetton and Rutherford, 1994), tree-ring proxies (Jacoby and D’Arrigo, 1990; Stahle et al., 1998); isotopic analysis from coral reefs (Dunbar et al., 1994; Urban et al., 2000; Tudhope et al., 2001), ice cores (Thompson, 1992), and multiproxy records (Mann et al., 2000). Changes in the frequency, magnitude and duration of ENSO, as derived from proxy land-based data sets around the Pacific Basin, have been observed throughout the last millennium with significant decreases in
ENSO activity during the Medieval Warm Period 900–1300 AD (Diaz and Pulwarty, 1994). Haug et al. (2003) and Hodell et al. (1995) even extend these discussions of the effects of the ITCZ, trades and ENSO to the northern most limits of this area in discussing the collapse of Mayan civilization. As the following discussion will illustrate, there are large uncertainties in understanding of the interrelationship between ENSO characteristics and the background mean climate at the present.

ENSO tends to be phase-locked to the annual cycle so that the largest SST anomalies occur in the boreal winter. Model simulations of the annual cycle at various times in the past imply a somewhat shifted and greatly reduced annual cycle in equatorial SSTs (warmer conditions in September and cooler in March) 3000 to 12,000 years ago relative to today (Trenberth et al., 1998). Interannual SST variability has been shown to have decreased during the mid-Holocene and to have increased during the Late Glacial Maximum (LGM). There is no scientific consensus on the causes of the suppression of ENSO activity during the mid-Holocene, or on the causes of the strengthening of ENSO activity during the LGM. Clement et al. (2000) propose that ENSO may be highly sensitive to orbitally driven changes in the seasonal cycle of solar radiation in the tropics. Koutavas et al. (2002) reconstructed SSTs, using Mg/Ca ratios in foraminifera from seafloor sediments near the Galapagos Islands, from the LGM to the present. Cold-tongue SSTs varied coherently with precession-induced changes in seasonality during the past 30,000 years. SST descend only 1.2°C implying a relaxation of tropical temperature gradients, weakened Hadley and Walker circulation, a southward shift of the ITCZ, and a persistent El Niño-like pattern in the tropical Pacific (Koutavas et al., 2002). This is contrasted with mid-Holocene cooling suggestive of a La Niña-like pattern with enhanced SST gradients and strengthened trade winds. As noted by Rosenthal and Broccoli (2004), these observations are intriguing but raise questions as to whether they are representative of the broader tropical Pacific or of only local conditions. Enfield and Mayer (1997; Enfield and Alfaro, 1999; Waylen and Poveda, 1997; Poveda et al., 2001, 2003; Pulwarty et al., 1992; Diaz and Kiladis, 1996; Chu, 1991; Glantz et al., 1991; Lau and Sheu, 1991; Halpert and Ropelewsky, 1992; Diaz and Kiladis, 1992; Pulwarty et al., 1992; Diaz and Markgraf, 1993, 2000; Pulwarty and Diaz, 1993; Marengo, 1992; Marengo and Hastenrath, 1993; Waylen et al., 1996a, b; Enfield and Mayer, 1997; Enfield and Alfaro, 1999; Poveda and Mesa, 1997; Poveda et al., 2001, 2003; Waylen and Poveda, 2002).

3.3. Present effects of ENSO in tropical America

In general, there is a coherent pattern of climatic and hydrological anomalies over the region during extreme phases of ENSO. Negative anomalies in rainfall, soil moisture and river flows, as well as warmer air temperatures, occur during El Niño. The reverse is true for the cold phase (La Niña), although there are regional differences in timing and amplitude. The climatic and hydrological effects of ENSO in the tropical Americas have been the subject of intense research (e.g., Hastenrath, 1976, 1990; Hastenrath et al., 1987; Waylen and Caviedes, 1986; Aceituno, 1988, 1989; Rogers, 1988; Kiladis and Diaz, 1989; Ropelewski and Halpert, 1987, 1996; Chu, 1991; Glantz et al., 1991; Lau and Sheu, 1991; Halpert and Ropelewsky, 1992; Diaz and Kiladis, 1992; Pulwarty et al., 1992; Diaz and Markgraf, 1993, 2000; Pulwarty and Diaz, 1993; Marengo, 1992; Marengo and Hastenrath, 1993; Waylen et al., 1996a, b; Enfield and Mayer, 1997; Enfield and Alfaro, 1999; Poveda and Mesa, 1997; Poveda et al., 2001, 2003; Waylen and Poveda, 2002).

3.3.1. Summary of regional physical mechanisms associated with ENSO

ENSO has affected the hydro-climatology of northern South America for centuries. For instance, Shongart et al. (2004) using dendrochronology reconstruction identified an increase in El Niño severity on Amazonia during the last two centuries. The present hydro-climatological anomalies associated with ENSO in northwestern South America can be diagnosed as a combination of mechanisms as follows (Fig. 8):

1. The reduction of the SST gradient in the eastern Pacific weakens the Choco jet, decreases mois-
ture transport inland (Poveda and Mesa, 2000; Poveda et al., 2001), particularly through the associated reduction in the intensity and number of mesoscale convective complexes (Velasco and Frisch, 1987). Fig. 9 reflects this reduction during El Niño (and intensification in La Niña) in the annual cycle of moisture transport by the Choco jet at 925 hPa. Such anomalies in moisture transported by the Choco jet are important contributors to the rainfall anomalies reported over central and western Colombia.


(3) Changes in pressures over tropical South America (Fig. 8) contribute to:

(4) the displacement of the centers of convection within the ITCZ in the eastern Equatorial Pacific towards the west and south of their normal positions (Pulwarty and Diaz, 1993).

(5) The reduction in the feedback between precipitation and surface convergence in tropical South America associated with Hadley cell circulation (Numaguti, 1993) and Caribbean trade winds.

(6) The disruption of land–atmosphere interactions, and the anomalies eastward travel with time, due to the regional coupling between anomalies of precipitation, soil moisture, vegetation and evapotranspiration (Nepstad et al., 1994; Jipp et al.,

Fig. 8. A summary of hydro-climatological mechanisms which combine to explain drought conditions in tropical South America during El Niño. Numbers and symbols correspond to mechanisms and processes listed in text.

Fig. 9. Long-term annual cycle of average moisture transport by the zonal winds of the Choco jet at 925 hPa, in units of \( \text{ms}^{-1} \ \text{g kg}^{-1} \), during El Niño years (triangles), and La Niña years (squares).
A reduction in evapotranspiration also contributes to decreased precipitation from evaporation recycling. Diminished cloudiness promotes increased solar irradiance and surface temperatures, thus reinforcing dry conditions. Even in wet tropical climates, water limitation (as during El Niño events) imposes conditions of water stress on forests (Oren et al., 1996). Land surface hydrological processes must be playing a major role in determining the velocity at which the anomaly propagates over the continent through positive land–atmosphere feedbacks (Fig. 10).

Initially, during the boreal summer of Year 0, the northeast trade wind fields intensify (weaken) during El Niño (La Niña), however in concordance with the changes in surface pressures (3) in the boreal winter, the fields reverse in Year+1, triggering a change in SSTs over the Caribbean and the tropical North Atlantic, physically explained by Curtis and Hastenrath (1995).

SSTs and the strength of the trades are two important controls on the intensity of tropical easterly waves and tropical storms over the North Atlantic and Caribbean (Frank and Hebert, 1974; Gray and Sheaffer, 1991).

3.3.2. Costa Rica and Panama

Fig. 11 shows the regionally dominant positive association between the SOI and annual precipitation (Estoque et al., 1985; Ropelewski and Halpert, 1987).

However, significant associations are geographically restricted to the Pacific slope. Negative associations implied along the Caribbean slope of Costa Rica, continue into the Boca del Toro region of northwestern Panama. The strongly unimodal regime of the Pacific slope (Figs. 2 and 3) is closely associated with the position of the ITCZ and cross-equatorilal westerlies (Hastenrath, 2002). Their southern and western displacement during warm phases of ENSO brings about a general reduction in rainfall, which is most apparent (Fig. 12) during the veranillos of July and August (Waylen et al., 1996b). A compounding factor may be the synchronous strengthening of the San Andrés jet (Table 1), enhancing the rain-shadow effect on the Pacific slope, jetting through topographic gaps (Legeckis, 1988; McCreary et al., 1989), thereby reducing SST (Magaña et al., 1999) and convective activity over the eastern Pacific warm pool. This is clear in analyses of seasonal streamflow (George et al., 1998) and flooding (Waylen and Laporte, 1999; Quesada and Waylen, 2004) in the region, where even the slightly higher risk of flooding in portions of the rainy season preceding the veranillos, correspond to the reduction in the San Andrés jet during MAM.

By contrast, the Caribbean regime is uniform (Figs. 2 and 3). Deviations found in association with warm phase ENSO (Fernández and Ramírez, 1991; Waylen et al., 1996b) are shown in Fig. 11 and Table 1, reflecting changes in the strength of the San Andrés jet: a slight reduction in MAM, marked increases in June–August, and a potential reduction in December–February. The offsetting seasonal changes in precipitation are responsible for the apparent lack of association of annual

Fig. 10. Schematics of the land surface–atmosphere interactions associated with extended dryness during El Niño in tropical South America. ET and Rn represent evapotranspiration and net radiation. Adapted from Fig. 14.3 of Sellers (1992). Reprinted with permission from ****.
precipitation totals with ENSO (Fig. 8). Rivers on the Caribbean slope (George et al., 1998; Quesada and Waylen, 2004) similarly show an increase in June–August and a decline in December–February, although this signal may be limited to elevations of less than 1000m. Above that altitude rainfall records indicate a response to ENSO more characteristic of the Pacific slope. It may be noteworthy that the greatest effect of ENSO on the San Andrés jet appears to be felt below 900 hPa.

The relationship between boreal winter precipitation (Figs. 3 and 4), which is generally associated with the passage of fronts, or nortes, originating over North America (Hastenrath, 1967; Klaus, 1973; Schultz et al., 1997, 1998), and ENSO, is unclear. However, Fig. 12 and analysis of flood timings (Quesada and Waylen, 2004) suggest that the earlier norte season (November–December) experiences a reduction in rainfall during warm phases of ENSO, while the latter period (January–March) is wetter. Schultz et al. (1998) provide some physical basis for this possibility in terms of the origins of the anticyclones and the positions of the mid-latitude and sub-tropical jets. During cold phases of ENSO these patterns are generally reversed as implied by observations of year +1 and +2 in Fig. 12.

The regional response of precipitation to ENSO in Panama is not as clear as in neighboring countries. The general tendency (with the exception of the extreme northwest) is for drought during El Niño (Estoque et al., 1985; Hastenrath, 1991), and excessive rainfall during La Niña. The interaction of topography and the regional jets appears to play a major role in controlling interannual variability in Costa Rica and Colombia. However, the latitude of Panama (7–9°N) falls between the cores of the Choco (4–6°N) and San Andrés (11–18°N) jets and its topography runs parallel to them.

3.3.3. Colombia

Both ENSO phases are associated with hydro-climatic anomalies in Colombia which amplify the hydrologic regime through complex and non-linear interactions. Seasonal cross-correlation analyses confirm that El Niño produces drier and more prolonged dry seasons, and La Niña wetter rainy seasons. Fig. 13 maps seasonal cross-correlations between the Multivariate ENSO Index (MEI), which synthesizes the ocean–atmosphere conditions of the tropical Pacific in both phases of ENSO (K. Wolter, NOAA http://www.cdc.noaa.gov/~kew/MEI/), and discharge records at the marked stations. Correlations are significantly positive and high for MEI during September–November and December–February with river discharges during De-

![Fig. 11. Distribution of simple Pearson product moment correlation coefficients of annual precipitation and the Southern Oscillation Index in Costa Rica. Red figures indicate positive correlations and blue are negative. Solid symbols are statistically significant at the 0.05 level.](image-url)
Correlations decay during March–May and beyond.

Streamflow and precipitation from western and central Colombia indicate significant positive correlations with MEI (~0.7 to 0.8) during December–February (Poveda et al., 2001), the season of ENSO’s greatest strength, and thus of greatest teleconnections and direct impact on Colombia. Forecasts of December–February discharges are possible as a result of the high correlations with the MEI of the preceding September–November and June–August (Poveda and Penland, 1994; Carvajal et al., 1998; Poveda et al., 2002, 2003). The skill of hydrological forecasting based upon the climate of the tropical Pacific is reduced during the boreal spring, as witnessed by the very low correlations with the MEI in March–May.

Analyses of diverse atmospheric and terrestrial fields of the NCEP/NCAR Reanalysis and hydrologic records lead to the conclusion that the hydrology is most affected by ENSO during December–February (year +1), September–November (year 0) and June–August (year 0), while MAM (years 0 and +1) are the least affected (Poveda et al., 2001).

Soil moisture is considerably depleted during El Niño, but the exact extent and severity is dependent on land cover (Poveda and Jaramillo, 2000). Coupling between the vegetation–soil system and land cover modulates hydrological (space-) time variability in the tropics, and the effects of El Niño-induced dry spells may potentially be modified. Similar strong regional patterns are apparent in correlations of vegetation cover (NDVI) with ENSO and with SOI (Myneni et al., 1996; Asner et al., 2000; Poveda et al., 2001; Poveda and Salazar, 2004). Decreased plant activity during El Niño is entirely consistent with negative anomalies in precipitation, streamflow, soil moisture, and actual evapotranspiration (Poveda et al., 2001).

Streamflow anomalies are more pronounced in December–February (year +1) and June–August (year 0) as a result of the combined anomalies in precipitation, soil moisture, and evapotranspiration. During September–November (year 0), effects are felt but ameliorated by smaller soil moisture and precipitation anomalies. Large-scale forcing and land surface hydrology play a key role in the dynamics of hydro-climatic effects of ENSO over the region (Marengo and Hastenrath,
1993) and in feedbacks within the regional land-surface–atmosphere system (Poveda and Mesa, 1997). The ENSO signal propagates to the east in northern South America, leading hydrological anomalies by one month over western Colombia (Poveda and Mesa, 1997) and by six-to-ten months in the Amazon River basin (Richey et al., 1989; Chu, 1991; Eagleson, 1994). Diagnostics of the physical mechanisms associated with these hydro-climatological anomalies are discussed in Section 3.2.1.

3.3.4. Venezuela

Land-sea temperature contrasts in the tropics strongly affect the evolution of monsoons. In most other monsoonal regions this gradient is driven by land temperature changes and therefore primarily by insolation. During ENSO events, the contrast in northern South America is also created by the large SST anomalies in the eastern Pacific. Pulwarty and Diaz (1993) observed that differences of up to 2 °C modulate the migration of March–May convection, and that deep convection moved westward during warm ENSO events, in contrast to north and eastward shift during cold events. This March–May convection in southern Venezuela and Colombia may result not only from warm SSTs off Peru but also from the westward migration of convection associated with the South American monsoon. Some portions of central South America may therefore experience diminished seasonal rainfall through redistribution of convection rather than total regional suppression. A corresponding eastward displacement was observed during the strong La Nina event of 1988–89. Further study of this east–west land–sea temperature contrast and its modulation during ENSO, especially by remote heating along the Peruvian coast, are needed (Yu and Mechos, 1999).

As in the remainder of the study area, the response to ENSO depends on the stage of the event’s life cycle (Giannini et al., 2000). The strengthened trade winds inhibit convective activity over the southern Caribbean during July and August (Year 0) of a warm event. But precipitation increases the following year (+1) is associated with weakened trades and positive SST anomalies in the tropical North Atlantic (Pulwarty, 1994; Enfield and Mayer, 1997; Poveda and Mesa, 1997; Giannini et al., 2000). Large-scale controls on interannual variability of Venezuelan climate may be summarized as follows. Wetter than average rainy seasons require; a persistent positive SOI, below normal SSTs in the eastern Pacific, above normal SSTs in the Caribbean.

![Fig. 13. Maps of seasonal cross-correlations between the Multivariate ENSO Index (MEI) and river discharge anomalies (Q) gauged at the marked points, during September–November (SON, top), and December–February (DJF, bottom), with following seasons. The diameter of the circle indicates the value of correlation, scaled as shown in the bottom right, with filled symbols representing significant correlations at the 0.05 level.](image-url)
bean in Year 0, weaker easterlies between 10°N and 20°N at 200 mb, below normal SSTs in the tropical South Atlantic, and the northward displacement of the South Atlantic High reaching into Brazil (Pulwarty et al., 1992). Regional variations in climatological controls are also manifest in relationships with ENSO, through both direct atmospheric links and the lagged influence of eastern Pacific SSTs. Rainfall of the Maracaibo region is significantly modulated by the eastern Pacific, as it is in the Andes region and the northern Cordillera de la Costa.

Coastal regions of Venezuela and Colombia are also subject to the influence of "nor tes". In December 1999, during a particularly heavy event (512 mm in two days i.e. more than the annual total for the north coast), convection was pushed onto the Cordillera de la Costa by an extratropical cold front moving off North America, which penetrated the tropics (Lyon, 2003). Over 15,000 people died, a toll greater than any previously associated with tropical storms during the past century.

3.4. North Atlantic

The North Atlantic is a potentially important regional source of moisture and control on the strength of the trade winds, propagation of easterly waves and tropical storms. The North Atlantic Oscillation (NAO) is one macro-climatic indicator of the state of the atmosphere–ocean system over the maritime area. First postulated by Sir Gilbert Walker, it is analogous to the Southern Oscillation (SO), reflecting differences between the subtropical and sub-polar North Atlantic at monthly and seasonal time scales. The NAO manifests itself in; westerly mid-latitude winds, SSTs, and the climate on the adjacent continental regions. Similar to the SO, the NAO is defined in terms of pressures under the subtropical anticyclone (e.g., Azores), and sub-polar surface cyclone (e.g., Iceland). A positive mode of the NAO arises when the subtropical high is anomalously strong and the Icelandic low is very deep. Westerly winds of the North Atlantic then become stronger. During negative NAO both centers are anomalously weak. Temporal variations in the NAO are linked to weather disturbances over Europe, the eastern United States, and the Mediterranean basin. Rogers (1988) suggests that precipitation over the Caribbean and the tropical Americas vary during extremes of the NAO, due to its influence over the strength of the North Atlantic trade winds. Hurrell et al. (2003) provide a detailed and updated discussion on the NAO.

Seasonal cross correlations of NAO and streamflow in Colombia (Poveda and Mesa, 1996; Poveda et al., 1998) indicate significant correlations, declining from north to south. Most exist between the NAO in MAM and discharges in DJF, and NAO in DJF with flows in JJA and SON. Fig. 14 depicts results from 10 rivers, with the correlation coefficient between the NAO index during the quarter shown on the abscissa, and river discharges during the seasons (quarters) marked by the different lines. These observations may be explained by the intensity of the trade winds over the tropical North Atlantic and the Caribbean, as influenced by the position and strength of the North Atlantic Anticyclone. The exact nature of this relationship and its interaction with the annual cycle of the trades, more intense in December–February less intense during June–August, warrants further research.

As noted previously, the strength of the trade winds is closely related to the veranillos of the Pacific slope and increased rainfall along the Caribbean, both of which show marked association with ENSO. Vargas and Trejos (1994) propose that changes in pressures in the North Atlantic are responsible for many low frequency changes observed in precipitation records within Costa Rica. An analysis of seasonal precipitation totals throughout Costa Rica (Waylen and Quesada, 2002) supports earlier conclusions (Enfield and Alfaro, 1999; Giannini et al., 2000), that regional precipitation may be sensitive to SST of the tropical North Atlantic. Monthly SSTs since 1950 over the Tropical North Atlantic and Niño 3.4 (www.nmic.noaa.gov/data/indices/index.html) are used to place each month into one of four classes depending on whether oceanic temperatures are above or below long run median. In Fig. 15, the pair of maps beneath the heading "Cold Pacific" can be viewed as being roughly analogous to the cold phase of ENSO and those under "Warm Pacific" are suggestive of warm phase, although this is far from a one to one correspondence. The horizontal divisions indicate the state of the tropical North Atlantic. Each chloropleth map depicts the class’ mean veranillos (JA) precipitation, expressed as a standard normal deviate from the undifferentiated station mean, under each set of combined oceanic temperature conditions.

The seasonal drought on the Pacific slope and excess rains on the Caribbean, associated with a warmer Pacific, are particularly marked during periods of a cooler Atlantic, with the regional cordillera clearly delineating the two effects. If the Atlantic is warmer, the drought is more ubiquitous but not as severe. A cooler tropical North Atlantic implies higher atmospheric pressures, a strengthening of the trades and a more pronounced windward/leeward effect. By contrast, a cooler Pacific
Fig. 14. Cross-correlation coefficients between the North Atlantic Oscillation (NAO) and ten river discharge stations in Colombia. Correlations are shown for the NAO during the season on the abscissa and river discharges during the season denoted by the symbols.
is associated with a greater onshore flow to the Pacific slope, and reduction of seasonal rainfall along the Caribbean. A warmer than normal tropical North Atlantic further encourages this onshore flow, particularly in northwestern areas of the country, where it may accompany a reduction in local upwelling of cooler oceanic waters in the Gulf of Papagayo.

Enfield and Mayer (1997) report that rainfall is enhanced over the Caribbean and Central America west of the cordillera, when cool SST anomalies are found in the eastern Pacific and warm ones in the tropical North Atlantic oceans. Such conditions weaken the trade winds, allowing enhanced convection over the Caribbean Islands and greater penetration of cross-equatorial westerlies along the Pacific coast. A cold Pacific and warm tropical Atlantic are robust features for June–August composite SST anomaly differences and often occur during rapid transformation from warm to cold ENSO phases, whereas a positive relationships between Atlantic SSTs and rainfall in the Llanos is clearly defined, especially a few months in advance.

4. Final remarks

We have reviewed the main features of the present climate of northern South America and Southern Central America, focusing on Costa Rica, Panama, Colombia and Venezuela, and its variability at annual and interannual time scales. Due to the specific time scales of our review, we have explicitly ignored important physical mechanisms controlling intra-seasonal climate variability in the study region. Among those we did not mention the activity of the 30–60 day or Madden–Julian Oscillation (Madden and Julian, 1972), which is known as the principal mode of climatic fluctuation in the intra-seasonal band over tropical South America (Mo and Kousky, 1993). The diurnal cycle of precipitation in the tropical Americas is another important time scale omitted from our study, but it certainly is associated with annual and inter-annual time scales in nontrivial fashion (Poveda et al., 2005). Likewise tropical easterly waves over the tropical North Atlantic, Caribbean and northern South America are responsible for the formation of many severe tropical weather systems on the Caribbean, ranging from severe storms to hurricanes. (Riehl, 1945; Riehl and Malkus, 1958; Chang, 1970). The strong altitudinal gradients within tropical mountain ranges particularly in the Andes of Colombia and Venezuela, and the Cordillera of Costa Rica are important in controlling climatic variability over space.
This synthesis of the principal mechanisms and phenomena which control climate variability over the study region at annual and inter-annual timescales provides important insights into variability at longer (paleoclimatic) timescales, and permits the evaluation of climatic evidence of climate and global change (IPCC, 2001), including the accelerated retreat of tropical glaciers (Kaser and Osmaston, 2002), outbreaks of tropical diseases (Patz, 2002), and the most rapid loss of biodiversity in the world (Myers et al., 2000). Besides, the region possesses urgent basic and applied research needs prompted by large scale deforestation, erosion and land degradation, vulnerability and risk of human populations and settlements, and water pollution.

Acknowledgements

Valuable discussions with colleagues throughout the years have contributed to shape these ideas. Among them we thank H. Diaz, S. Hastenrath, O. Mesa, P. Aceituno, C. Nobre, C. Penland, C. Caviedes, and H. Riehl. TRMM data set was kindly provided by NASA Goddard Space Flight Center, Data Archive and Distribution Center (DAAC), and data set from the NCEP/NCAR Reanalysis by E. Kalnay. Research referred to in this paper has been variously supported by; CIIRES University of Colorado, Boulder) (GP), Colciencias (GP), DIME (Universidad Nacional de Colombia at Medellin) (GP), Inter-American Institute for Global Change Research (GP and PW), NOAA (PW), National Science Foundation (PW). M. D. Zuluaga and P. A. Arias helped to prepare Figs. 6 and 9, respectively.

References


