The rainfall over tropical South America generated by multiple scale processes

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The rainfall over tropical South America generated by multiple scale processes

by

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Program of Study Committee:
Tsing-Chang Chen, Major Professor
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This thesis is dedicated to the memory of my young brother Armando Carrillo who taught me to enjoy every second of this life, and who will be in my heart forever.
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ABSTRACT

The rainfall regime over Central America and tropical South America is the most important hydro-meteorological event in the tropics after the Asian-Australian monsoon system. Therefore, in this study we analyzed how the tropical rainfall over tropical South America is developed using a multiple time and space scale approach.

It was found that the rainfall regime is affected mainly by the ENSO, showing two opposite phases not only in the precipitation, but also in the dynamics terms. Although, there is an advance understanding of the role of the intraseasonal mode of the global tropical precipitation, documentation of the role of the 12-24 day mode over this area of the world is not found. We were able to link rainfall perturbation systems with the 12-24 day mode, which seems to be the dominant intraseasonal but it is embedded and modulated by the 30-60 day mode. Thus, the intraseasonal (12-24 and 30-60 day modes) acts in a harmonic way to portray a quasi-periodic behavior of the rainfall and dynamics elements over the tropics in South America.

Finally, a description of the weather systems embedded in these scenarios mainly affected by the warm and cold ENSO phase are presented. There it was found synoptic (eastward) and high-frequency (westward) moving disturbances interaction can help to understand the rainfall pattern under the influence of the ENSO in tropical South America. Thus, during the cold ENSO the positive anomalous precipitation is explained because the constructive interference of these two perturbations and during the warm ENSO the synoptic disturbance is suppressed over the tropics.
CHAPTER 1. GENERAL INTRODUCTION

Understanding the variability of the rainfall regime over tropical South America is important because of the strong dependence on water supply in hydro-energy, agriculture, and transportation. For example, 60% of the energy consumed in Peru is generated by hydro electrical plants. Also, Brazilian agriculture is mainly supported by natural irrigation. Moreover, forest countries like Brazil and Peru depend on the rivers’ navigability to provide an affordable means of transportation. Therefore, the importance of understanding the mechanism to produce water through rainfall variability in tropical South America has not only meteorological implications, but also practical applications to social issues within the region.

Eastward and westward rainfall perturbations have been identified in tropical South America. Kousky (1980) claims that the westward perturbation is generated by the local land sea-breeze. Cohen et al. (1995) identified that in some cases those westward perturbations can penetrate inland. Recently, Liebmann et al. (2009) found some types of eastward rainfall propagations which can generate west in tropical South America over the continent or offshore over the ocean. Also, Petersen et al. (2002) have classified the rainfall pattern over tropical South America using the direction of the trade winds over the continent (westerly and easterly). However, their classification only works in the central continent and south of the equatorial line (~60°W, 10°S). Thus, the idea of westward and eastward-propagating disturbances seems to be very important in the role of rainfall production over tropical South America. Nevertheless, those westward/eastward perturbations might not be isolated
processes. Indeed, they could be embedded in processes of superior order which might control and modulate their development. Thus, the first question which emerges is: what are the processes which rule the synoptic scale behavior in tropical South America?

The rainfall in the tropics has different dynamics compared with middle latitudes (Charney 1963), and the mid-latitude approach like the quasi-geostrophic theory is not applicable. However, over time, other theories have risen to explain the dynamics in the tropics. For example, Halley (1686) proposed that as a consequence of the heating in the tropics, the air is less dense below and because of this, it lifts easier. Thus, he proposed an early theory explaining tropical circulation. Later, Hadley (1735), using the principle of angular momentum conservation, gives solid arguments to explain the easterly component of the trade winds. In the past century, Bjerknes (1969) drew attention to the east-west circulation (Walker Circulation) which is linked to heating areas in the tropics, with the ascendant branch in the west Pacific and the descendant branch in the east Pacific. What can be the importance of the ascendant/descendant branch to the weather in tropical South America?

The surface heating seems to be an important forcing, which can be corroborated by the three major areas of rainfall observed in the tropics, and are associated with the global distribution of diabatic heating (Chen et al. 1986). The most important area of rainfall production is located in the Maritime Continent (Ramage 1968), where the Asian and Australian monsoon systems play an important role. The other two areas are located over Central and tropical South America and tropical Africa (Chen 2005). Thus, the hydrological cycle of the earth can be seen as a sink area over the Asian-Australian monsoon region and a source area in the complementary region (Chen et al. 1995b). How do the perturbations over
the Asian-Australian monsoon area affect the weather and climate behavior over tropical South America?

There is an east-west hemispheric interchange/interbalance which is not only seen in the hydrological cycle, but also in dynamic elements as velocity or its derivations. For example, using the potential function of water vapor transport \((\chi_Q)\), Chen et al. (1995b) showed that the planetary scale sink area (high precipitation) is maintained by the convergence of water vapor transport, and is also forced by the potential function velocity \((\chi)\) which is paired together in lower and upper level. Can we find an analogous representation to portray the interaction of processes over tropical South America? Can we use that representation to understand the function of individual terms in a multiple scale processes? Before answering those questions, a brief description of the precipitation regime and the diversity of time scales are described.

**Background review of the climate variability**

**Precipitation**

South America is a source of water vapor during the wet phase of the Asian-Australian monsoon system and the sink of water vapor during its dry phase. Therefore, this wet/dry process oscillates and is ruled by the annual cycle of the global circulation as a first order of approximation. The global rainfall pattern has a quasi-periodic behavior like the annual and diurnal cycles (Chen et al. 1995b), and South America experiences the wet phase of the annual rainy cycle during the austral summer.

The variance of monthly mean precipitation departure can help us to identify areas of maximum perturbation on Central America and tropical South America. In Fig 1.1b, three
areas with maximum rainfall variance can be observed, which are denoted by 1, 2, and 3. Area 1 (A1) corresponds to the North American Monsoon System (Garreaud and Wallace 1997), which has a maximum peak during June-July-August (JJA). Area 3 (A3) coincides with the tropical Atlantic Inter Tropical Convergence Zone (ITCZ), which has a clear signal during Jun-Nov (Gu and Zhang 2001). Finally, area 2 (A2) is located over the South American Monsoon System (SAMS) (Vera et al. 2006).

Thus, over South America, the maximum variance, A2, is located between 10°S and 0° in the central and east part of the continent. The rainfall variability of the area A2 (Fig. 1.1c) portrays a clear annual mode with a maximum peak in March and a minimum in September, so the wet season can be observed from December to May. Although the rainy season starts in December, an abrupt increase can be noted in February, then an abrupt decrease after April. The peak during February-March-April (FMA) is known as the fall annual mode, and it was identified in other studies (e.g. Kousky 1980). Also, the location of the maximum variance is confirmed when the variance of the annual and semi-annual Fourier harmonic is computed (Fig. 1.1e). Therefore, the time period for this study is focused over the fall annual mode (FMA), because it seems that FMA is the period where the rainfall perturbation is more active.

The climate variability

The rainfall during the FMA season is affected by diverse processes at different time scales. For example, Vera et al. (2006) summarized the current understanding of the variability of the rainfall over South America from interdecadal to synoptic scales in the way to have a unified view of the South American monsoon system. Also, Ropelewski and
Halpert (1987, 1989) pointed out the effect of the Southern Oscillation of the interannual variation of the rainfall over South America. They found, using a time series analysis of approximately 1700 globally located stations, that a large scale regional precipitation pattern in northeastern Brazil is affected by a dry (wet) rainy season during cold (warm) years of the El Niño/Southern Oscillation (ENSO).
Also, Grimm and Zilli (2009) and Paegle et al. (2002), using Empirical Orthogonal Functions (EOF), identified areas of correlation between rainfall and Sea Surface Temperature (SST). Most of the areas were in agreement with Ropeleski and Halpert's results. In Peagle et al.'s research, they not only studied the effect of the ENSO over South America, but they also included the effect of the North Atlantic Oscillation (NAO). Using Rotated Empirical Orthogonal Functions (REOF), they found that the leading mode of the REOF analysis explained 12% of the total variance. The results are in agreement with previous studies, and they found that the NAO affects the area in northeastern Brazil, but the NAO contribution is limited when compared with the ENSO contribution.

Using rawinsonde data over Canton Island (3°S, 172°W), Madden and Julian (1971) found the existence of a pronounced peak (41-53 day) in the spectrum of zonal wind in upper (150 mb) and lower (850 mb) levels. They concluded that the revealed oscillation is a manifestation of the large scale circulation, which is located near the equator. Also, they claimed that the oscillation can not be a Kelvin wave type, because the meridional wind does not show an intense peak in the power spectrum. In South America, the intraseasonal variation, known as the Madden-Julian oscillation, was found to have an important impact on the rainfall pattern although most attention was on the South Atlantic Convergence Zone (SACZ). Thus, Nogues-Paegle and Mo (1997) claimed the existence of a dipolar zone between the SACZ and subtropical area. Also, a wave train in sub-tropical areas was proposed as a forcing mechanism to explain the dipole system (Grimm and Silva Dias 1995). However, few efforts have been made to investigate the relevant importance of the Madden and Julian Oscillation (MJO) in tropical South America, and no report of the impact of the 12-24 mode has been found to explain the mechanism of the rainfall formation over this area.
Although the 12-24 day mode was identified to have a global impact (Chen et. al 1995b), researchers are still analyzing the intraseasonal spectral band as a 10-90 day mode (Nogues-Paegle et al. 1998).

As was pointed out by Vera et al. (2006), studies of synoptic perturbation on tropical South America are limited. The effect of the incursion of mid-latitude systems in tropical areas (Garreaud and Wallace 1998) is only mentioned a few times. In those cases, frontal systems are the main mechanisms that alter the meteorological conditions over the Amazon basin, and they can occur at any time during the entire year. During the dry season (winter-JJA) an extreme descent of temperature is observed, and during the wet season (summer-DJF) the cold front triggers cumulus convection such as mid-latitude front systems (Marengo et al. 1997). Garreaud and Wallace (1998) identified that cold surges in the southern region of South America can incur into the tropics during summer time producing rainfall ahead of the cold front. Also, Kousky (1979) found that the rainfall over northern Brazil is affected by those cold surges. On the other hand, using EOF and composite analysis of meridional wind and water vapor, Kayano (2003) identified synoptic disturbances propagating westward in the tropical Atlantic ocean through South America. He also found three different ways of synoptic systems’ perturbation: 1) equatorial trade, 2) westward disturbance, and 3) mid-latitude synoptic wave systems. In his EOF analysis of 925 meridional wind data, the first eigenvector has a dipolar structure, and the second eigenvector has a monopolar horizontal structure. The first and second components are in quadrature, so a propagation structure can be constructed and used as a proof of synoptic perturbation propagation; however, unclear evidence of synoptic disturbances using weather charts has been shown to prove the dynamic forcing of the rainfall pattern. For instance, divergence fields are used to prove synoptic
disturbance propagation (Cohen et al. 1995). It was also proven by Charney (1963) using scale analysis that in the tropics the divergence field is very noisy because the importance of vertical motion. Therefore, a deep analysis of the role of synoptic disturbances in rainfall generation is needed for tropical South America.

Motivation and thesis organization

Although many efforts have been made to understand the rainfall regime over tropical South America in its different time scales, few investigations have addressed the problem as an interaction of processes from lower to higher frequencies. Also, the role of the 12-24 day mode perturbation on rainfall generation is not well understood, even though it is very well known that there exists a direct effect on the rainfall regime over tropical South America as a consequence of the ENSO.

Moreover, there is no documentation of how the systems at synoptic scales are affected, and how these changes are reflected on the convective processes responsible to generate rainfall. Therefore, the main question in this research is: how does the interaction of synoptic disturbances propagating systems help to clarify the enhancement (suppression) of rainfall in tropical South America? Thus, this investigation pretends to give indications of possible mechanism which generate the rainfall regime during the ENSO phases.

This study is organized as follow: in chapter 2, the data sets used are introduced and the methodology is explained. An overview of the role of the ENSO in the interannual rainfall variation, and the impact of the east-west Walker circulation during ENSO phases is presented in chapter 3. A discussion about the modulation of the rainfall by the intraseasonal
mode on tropical South America, and a description of the role of westward and eastward disturbances in the enhanced/suppressed rainfall embedded in processes of multiples scales, with a tentative explanation of the function of the horizontal synoptic configuration is presented in chapter 4. Finally, in chapter 5, general conclusions are provided and tentative future research work is discussed.
CHAPTER 2. DATA SOURCES AND METHODOLOGY

It is well known that the convective activity over tropical South America is controlled by multiple time scale processes. These time scale processes are in the order of interannual, intraseasonal, synoptical, and diurnal. Chen et al. (1995a) and Chen et al. (1995b) showed that the intraseasonal 12-24 and 30-60 day modes have an important impact in the global rainfall regime. However, little effort has been made to investigate the impact of the higher frequency mode connected with the low frequency mode on the modulation of the convective activity over tropical South America. Therefore, in order to prove the existence of multiple time scale interactions which rule the rainfall activity over tropical South America, different time and spatial scale data sets were employed. The data sets used to evaluate the multiple scale processes are listed in table 1 and described in the following sections.

2.1 Data sources

2.1.1 Precipitation data

Three types of precipitation data sets were used:

CMAP

In this research, The Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) data set was used to explore interannual and intraseasonal patterns. The CMAP data set proved to be a reliable data source for climate studies (Xie and Arkin
1997). Briefly, the technique used to estimate precipitation by CPC is through merging observation from rain gauges, rain estimation from several satellites (infrared [IR], outgoing longwave radiation [OLR], microwave sounding unit [MSU], and microwave [MW] scattering and emission), and precipitation forecasted by numerical models. The merged analysis consists of two steps: the first step is reducing the random error by using the maximum likelihood estimation method to calculate the shape or distribution of the precipitation field, and the second step is to constrain the amplitude of the precipitation by using rain gauge observation with a blending method proposed by Reynolds (1988). The time resolution data set consists of pentad and monthly analysis of the gridded field of global precipitation. Our analysis covers the period from 1979 to 2008 for the monthly data and from 1979 to 2007 for the pentad data. The horizontal resolution of the data set is 2.5° (longitude) by 2.5° (latitude) degrees for both cases.

**TRMM**

High horizontal and temporal resolution precipitation data sets were used in this research to analyze the importance of high frequency modes as well as the diurnal cycle on the rainfall over the Amazon Basin. To accomplish this goal, the recently released Tropical Rainfall Measuring Mission (TRMM) Multisatellite Precipitation Analysis was used. TRMM provides a blending of multiple satellites and rain gauge observation. The input data sets and algorithms to generate the TRMM rainfall are described in detail by Huffman et al. (2007). The horizontal resolution of TRMM is 0.25° (longitude) by 0.25° (latitude) degrees, and its temporal resolution is 3 hours. Its geographical extension is global, but it is limited in latitude from 50°S to 50°N.
OLR

Although there is good quality and easily accessible high resolution data after 1998, it is not the case for previous years. However, it was important to analyze the interaction of different scales before 1998. Therefore, interpolated Outgoing Longwave Radiation (OLR), from the National Oceanic and Atmospheric Administration (NOAA), was used as a good proxy for deep tropical convection activity. Arkin and Ardany (1989) pointed out that $\Delta \text{OLR} = 235 \text{ W.m}^{-2} - \text{OLR}$ can be a useful index for convective activity, especially in the tropics. The advantage of using the OLR data set is that it covers the gaps before 1998 where there is not daily data available. This interpolated version from NOAA is advantageous because it covers most of the missing values or incomplete global coverage due to satellite problems (Leibmann et al. 1996). The horizontal resolution of OLR is $2.5^\circ$ (longitude) by $2.5^\circ$ (latitude) degrees, with a daily time resolution scale. The data set used in this research corresponds to the time period from 1979 to 2008.

2.1.2 SST

Correlation with the Niño 3.4 index and other variables such as rainfall were computed to investigate possible teleconnections over the continental area. The Niño 3.4 index is defined as the areal average of SST over a box limited by $5^\circ$S - $5^\circ$N and $120^\circ$W – $120^\circ$W. This Sea Surface Temperature (SST) data set was provided by CPC (Climate Prediction Center).

2.1.3 Wind

Surface and upper level charts were analyzed in order to illustrate how the synoptic
systems and climatology respond under a particular event like the ENSO. Those surface and upper level winds are used from three different sources: Reanalysis 1, Reanalysis 2, and Merra.

Reanalysis 1 and Reanalysis 2 have the same spatial resolution. Although, Reanalysis 1 has been available since 1948 and Reanalysis 2 has been available since 1979, this research analysis is restricted to the years between 1979 - 2008. Therefore, there is no distinction in time from R-1 and R-2. Also, the time resolution for both R-1 and R-2 is 6 hours, and the spatial resolution is 2.5° (longitude) by 2.5° (latitude) degrees, as well as the vertical resolution (28 levels).

NCEP/NCAR Reanalysis 1 (R-1) uses a state of the art global data assimilation system, which was fed with the major observational data sets that existed at that time. Also, it worked with the same Data Assimilation System (DAS) during the whole simulation period, with the idea of eliminating possible jumps in time series observation. To perform that, an advanced quality control was designed (Kalnay et al. 1996). All these characteristics make R-1 a reliable source of data.

However, the number of multiple-step processes involved in the generation of the R-1 data set makes it prone to many human errors as it was documented by Kanamitsu et al. (2002). Therefore, a second version of reanalysis was released, the NCEP/DOE Reanalysis 2 (R-2). Even though this data set is not a new generation of reanalysis, it can be considered a post version of R-1 where the human error detected after R-1 was released, were fixed. In R-2 some additional observational data sets were added after 1993 because the data processing system implemented in R-1 was improved, and observed soil moisture forcing was also added.
Spatial high resolution interactions were analyzed using the Modern Era Retrospective-Analysis for Research and Applications (MERRA). It is a NASA reanalysis data set for the new satellite era, and its main characteristic is the use of the sophisticated Goddard Earth Observing System (GEOS)-5 DAS (Rienecker et al. 2008). The period of the data is from 1979 to 2007, and the horizontal resolution used in this research is 0.66° (longitude) by 0.50° (latitude) degrees. Only vertical levels at 850 and 200 mb were employed.

2.1.4 Observational synoptic charts

Synoptic charts were obtained from National Climatic Data Center (NCDC) and CPTEC (Centro de Previsão de Tempo e Estudos Climáticos). Those charts were used to confirm that the synoptic perturbation observed in the R-1 or R-2 reanalysis data during specific events was also registered at the surface. This helps us to validate possible errors with the reanalysis data set. The data was accessed via the Service Records Retention System (SRRS) at the NCDC of NOAA. They provide analysis and forecast charts for several regions of the world, and the corresponding charts for tropical analysis were used in this research. In most cases, the information is available every 6 hours.

2.2 Methodology

Our strategy is splitting the data sets in several time scales: interannual, intraseasonal, synoptic, and diurnal. Fourier analysis and variations of spectral techniques were used for this purpose. The Hovmöller diagram is employed to portray the interaction of multiple scales. Also, the maintaining of precipitation is analyzed through the water vapor budget
equations. In addition, the circulation is divided in its rotational and divergent circulation using the Helmholtz’s theorem. Other tools were used to link all the different scales from the interannual variation through the diurnal cycle. In the following sections these procedures are described.

### 2.2.1 x-t diagram

Since Hovmöller (1949), the tracking of atmospheric perturbations has been analyzed using the x-t diagram. In this research, we want to investigate how diverse atmospheric disturbances like the Madden and Julian Oscillation (MJO) or the 12-24 day mode are embedded in a large scale. Therefore, several x-t diagrams are plotted to portray rainfall propagation. This is used in a similar manner as done by Hovmöller (1949). However, depending on the time scale, diverse features of the dynamics can be observed, which are going to be clarified in this study. For example, Petterssen (1956) illustrated the use of the Hovmöller diagram to observe propagation of an individual trough or ridge embedded in the amplification of these troughs and ridges, which can be observed because of the difference in speed velocity. In this particular research, it is possible to observe different types of perturbation because of the used of the x-t diagram approach.

### 2.2.2 Divergent wind

As it was established by several studies (e.g. Chen and Baker 1986), the east-west Walker Circulation is the main large scale mechanism that suppresses/enhances the rainfall activity over tropical South America. Therefore, the east-west circulation is plotted to portray the interannual variation of the suppression/enhancement of the convective activity.
Following Chen's approach (Chen 1985) the divergent wind component was computed for this purpose.

The Helmholtz's theorem establishes that a wind field can be separated into rotational and divergent components as in equation (1a).

\[ \vec{V} = \vec{V}_r + \vec{V}_d \]  

or

\[ \vec{V} = \hat{k} \times \nabla \psi + \nabla \chi \]  

(1b)

Where \( \vec{V} \) is the total wind, \( \vec{V}_r \) is the rotational wind component, \( \vec{V}_d \) is the divergent wind component, \( \psi \) is the streamfunction, \( \chi \) is the potential function, and \( \hat{k} \) is the unity vector in the z direction. If the divergence is applied to the equation (1b), the first term of the right side will be zero because a pure rotational field does not have divergence. Hence, the resultant equation will be a Poisson type equation such as (2).

\[ \nabla \cdot \vec{V} = \nabla^2 \psi \]  

(2)

Therefore, \( \chi \) can be found solving (2), and it can be used to construct the divergent wind field as in equation (1a). In this research, the approach employed by Chen (1985) was used.

**2.2.3 The water-vapor budget**

The hydrological cycle can be depicted by the water vapor budget equation (Chen, 2006).

\[ \frac{\partial W}{\partial t} + \nabla \cdot \vec{Q} = E - P \]  

(3)
Where, E=Evaporation, P=Precipitation, $\vec{Q} = \frac{1}{g} \int_{0}^{p_f} VqdP$ = water vapor flux, q=specific humidity, p=pressure, g=gravity, V=velocity vector, and $W = \frac{1}{g} \int_{0}^{p_f} qdp$ = precipitable water,

In the stationary regime $\frac{\partial W}{\partial t} = 0$ and the (E-P) is maintained by the imbalance of $\nabla \cdot \vec{Q}$. As water vapor is confined in low levels, $\nabla \cdot \vec{Q}$ should be tied to the (E-P) behavior.

Using again the Helmholtz’s theorem, the water vapor flux can be expressed as follows:

$$\vec{Q} = \vec{Q}_D + \vec{Q}_R$$  \hspace{1cm} (4)

or

$$\vec{Q} = \vec{k} \times \nabla \psi_{\chi} + \nabla \chi_{\nabla}$$  \hspace{1cm} (5)

And applying the diverge operator, $\nabla \cdot$, to (7)

$$\nabla \cdot \vec{Q} = \nabla^2 \chi_{\nabla}$$  \hspace{1cm} (6)

$\frac{\partial W}{\partial t} + \nabla^2 \chi_{\nabla} = E - P$  \hspace{1cm} (7)

Therefore, $\chi_{\nabla}$ can be seen as the tool to identify source/sink areas used by the climate system to produce rainfall.

### 2.2.4 Power spectral and band-pass filtering

A spectrum analysis is performed to identify the predominant time scale perturbation from the precipitation time series, and the Cooley-Tukey fast fourier method was used for this purpose (Cooley and Tukey 1965). The same technique was employed by Murakami (1976) to identify disturbances over the tropical Atlantic during the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE).
Moreover, the band-pass filter method is similar to that used by Murakami (1979), he applied a method called recursive technique for the band-pass filter. There the main idea was to create a variable transformation from the original time series in order to have another manner to compute the filtering algorithm, so a low order Butterworth function can be used to assure a rapid convergence. The advantage of this technique is to avoid the cutoff of the result when the time series length is not comparable to the periodicity of the perturbation.

Table 1. Different datasets used to analyze the multiple scale processes.

<table>
<thead>
<tr>
<th>Data</th>
<th>Time resolution</th>
<th>Horizontal Resolution (longitude, latitude)</th>
<th>Study</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMAP</td>
<td>pentad/monthly</td>
<td>2.5° x 2.5°</td>
<td>Xie and Arkin (1997)</td>
</tr>
<tr>
<td>OLR</td>
<td>daily</td>
<td>2.5° x 2.5°</td>
<td>Liebmann et al. (1996)</td>
</tr>
<tr>
<td>TRMM</td>
<td>3-hour</td>
<td>0.25° x 0.25°</td>
<td>Huffmann et al. (2007)</td>
</tr>
<tr>
<td>R-1</td>
<td>6-hour</td>
<td>2.5° x 2.5°</td>
<td>Kalnay et al. (1996)</td>
</tr>
<tr>
<td>R-2</td>
<td>6-hour</td>
<td>2.5° x 2.5°</td>
<td>Kanamitsu et al. (2000)</td>
</tr>
<tr>
<td>MERRA</td>
<td>6-hour</td>
<td>0.66° x 0.50°</td>
<td>Rienecker et al. (2008)</td>
</tr>
</tbody>
</table>
CHAPTER 3. THE ROLE OF THE ENSO IN THE INTERANNUAL TIME SCALE
OF THE RAINFALL OVER TROPICAL SOUTH AMERICA

3.1 Introduction

Understanding the variability of the interannual rainfall over tropical South America is important because the high vulnerability of the infrastructure in the area. For example, during 1998, cities in northern Peru (Tumbes and Piura) were hit by an unusual sequence of storms. That affected the entire economy because the urban infrastructure was not designed to support tropical rainfall, and because Piura city is almost a desert. The damages were estimated in the order of US $ 3480 million\(^1\). Recently, in May 2009, 40 people died and about 300 000 were homeless in northern Brazil as consequence of a extraordinary flooding\(^2\). Therefore, understanding the interannual variation of the rainfall in Tropical South America has practical applications to solve social issues.

Several studies focused their attention on the impact of the El Niño/Southern Oscillation (ENSO) over the interannual variation. For example, Ropelewski and Halpert (1987, 1989) concluded that during the warm phase of the ENSO, a negative rainfall anomaly is observed over continental South America. Moreover, other studies analyzed the influence of the Atlantic SST. Thus, Marengo (1992) observed that an excess of rainfall can be a consequence of an intensification of the trade winds which carry moist air from the Atlantic

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\(^1\) Ministry of Economy and Finances of Peru, [www.mef.gob.pe](http://www.mef.gob.pe)

\(^2\) VOA NEWS, May 11, 2009, [www.voanews.com](http://www.voanews.com)
ocean to the central continent. Mora and Shukla (1981) and Nobre and Shukla (1996) claimed that the trade winds intensification may be due to a southern migration of the Inter Tropical Convergence Zone (ITCZ) possibly controlled by perturbations in the Atlantic Ocean. However, numerical simulations performed by Law and Nath (1994) established that perturbations in the Atlantic Ocean could be driven by the Pacific Ocean. Therefore, the ENSO may play a primary role in the interannual variation of the rainfall regime of South America. This idea was corroborated by Grimm and Zilli (2009) and Paegle and Mo (2002) analyzing Sea Surface Temperature (SST) and precipitation using empirical orthogonal function.

Therefore, the purpose of this chapter is to analyze how the ENSO impacts the rainfall regime of tropical South America in an interannual time scale and to also identify how the atmospheric circulation responds and links this interannual perturbation with processes over the continent.

3.2 Interannual variation of the SST

Several authors choose DJF as the season to study the rainfall interannual variation over South America. This happens because 60% of the rain is produced during the mature monsoon life cycle (Garreaud and Aceituno 2001), but there is something that has not been explored much beyond the scope of the DJF season in tropical South America. Thus, this study observed a good rainfall signal during FMA as was indicated in the previous chapter, and the FMA season can be named as the annual-fall mode. Kousky (1980) found a contrast in the rainfall regime analyzing the diurnal cycle of rainfall in northern Brazil. He pointed out that eastern tropical South America has its maximum rainfall during MAM, which
corresponds to the austral fall, and in the southern region the maximum rainfall occurs during DJF, the summer austral annual mode. As in other studies done in other parts of the world, the spring/fall mode has not been very well explored (Wang 2004). Wang (2004), using Fourier harmonic analysis and empirical orthogonal functions (EOF), isolated the spring mode for North America and found that the spring mode is a good representation of the precipitation occurrence during the spring season in the Northern Hemisphere. Using the same procedure, we found that the contribution is small (<20%) for the mature phase of the SAMS, though over the tropical continent it demonstrated an important contribution. This is also confirmed in the variance and annual variation of rainfall (Figs. 1.1b and 1.1d).

Two major features can be observed when the long term mean seasonal average of FMA rainfall is plotted (Fig. 2.1a): 1) a maximum over northeastern South America is observed, which occurs on the mouth of the Amazon River. This area of important rainfall has been studied intensively. Early, Kousky (1980) found the existence of a strong diurnal cycle associated with the differential heating between the land and the ocean. He proposed that a large scale continental land breeze offshore which converges in low levels with the onshore trade wind is a possible mechanism to explain this phenomenon. Although, there are not more details about the role of the interannual variation, it can be inferred that there is a special season, January to May, for the occurrence of this continental scale land-sea breeze; 2) The other maximum precipitation occurs around 70°W, even though its intensity is less than the maximum area in the eastern coast. The possible mechanisms for the genesis of this convective activity are: a) diurnal diabatic heating (Nguyen et al. 2008), and b) mid-latitude interactions such as cold fronts (Liebmann et al. 2009). In addition, the same plot using GPCP (Fig. 3.1b) confirms the reliable estimation of rainfall by CMAP.
On the other hand, when the variance of the interannual variation (FMA) of precipitation departure (ΔP) is computed, only one of the two previous described areas shows a maximum perturbation in an interannual time scale, and another area appears. Those areas are labeled as 1, A1, and 2, A2, in Fig. 3.1c. The former is located in the tropical west coast and the latter in the east coast. Both A1 and A2 show a similar amplitude perturbation. A2 was first noted by Ropelewski and Halpert (1987), they used the Southern Oscillation (SO) index to relate rainfall around the world with global circulation. A1 drew attention after the ENSO 1983 (Philander 1983) for the international community, but local researchers from
Peru documented this event prior to 1900 (Carrillo 1892). Later, Horel and Cornejo-Garrido (1986) showed that the rainfall in northern Peru, A1, may be due to the local land-sea breeze process. In area A1, the SST index Niño 1+2 shows a correlation over 0.8, which is little higher than the correlation with the SST index Niño 3.4. Therefore, this pattern is affected by the large scale circulation which is locally amplified by the response of the warm/cold ENSO phase.

Figure 3.2  a) The histogram of the interannual variation of precipitation departure over area 1 superimposed with the interannual variation of SST index Niño 3.4, and b) similar as a) except for the area 2.
The interannual variation of precipitation over A1 and A2 is plotted in Fig. 3.2, with the Niño 3.4 index (solid line) superimposed. It can be noted that extreme rainfall in area 1 is very closely related to a strong SST positive anomaly. Although the correlation between ΔP and ΔSST (Niño 3.4) is not high, higher correlation can be obtained if the index Niño 1+2 is used instead of the Niño 3.4 (not shown). The increase in correlation may be an indication of the role played by coastal processes in the rainfall over this area (Horel and Cornejo-Garrido 1986).

A seesaw relationship between the rainfall in A1 and A2 can be observed in Figs. 3.2a and 3.2b. It can be noted that this seesaw relationship is better defined during warm years. In this research, warm years are defined as years which SST departure is 0.5 °C greater than its long term mean, and cold years which SST departure is lower than -0.5 °C. Thus, in 1983 and 1998 it can be found that the precipitation is higher in A1 and lower in A2, which coincides with the highest value of SST during 1983 and 1987. The seesaw relationship during extreme warm years seemed to be related to the east-west Walker circulation as it will be discussed later.

On the other hand, during cold years the response of rainfall against SST is not as linear as in warm years. In A1, extreme cold [SST] does not mean extreme dry [P], and in A2, extreme cold [SST] does not mean extreme wet [P]. This happens because the rainfall over tropical South America is a combination of processes at different time scales which interact between them using propagation properties to produce the observed rainfall pattern. The difference between one stage (cold/warm) is how the perturbation is triggered or suppressed in relation to the SST modulation. For example, in A1, the strongest cold SST (Niño 3.4 index) value does not mean a very dry season as can be seen in 1989 and 1999.
Also, A2 shows two extreme wet years in 1985 and 2000, but 1989 is not as wet as can be expected with a lower SST value. Therefore, we can infer that although a cold SST year means more rainfall over the tropical continent, this process is not as 'linear' as it might be observed for warm years in A1. This immediately raises the question: what mechanism makes response A1 more linear during warm years than A2 during cold years. If A2 shows more complex behavior, is this behavior caused because more processes are involved in it?

Figure 3.3  
a) The correlation of both precipitation and SST index Niño 3.4 departure for the interannual variation (FMA), b) the composite precipitation departure for cold ENSO years, and c) the composite precipitation departure for warm ENSO years.

Since the previous analysis, it can be concluded that the SST is an important climate component in regulating and controlling the interannual variation of the rainfall over tropical
South America. This can be noted in Fig. 3.3a where the correlation of SST (Niño 3.4) is shown against precipitation. The correlation is drawn only from cold and warm years, so neutral years are removed. A similar result is found when a complete SST time series is used, but using only composite ENSO years (1983, 85, 87, 89, 92, 96, 98, 99, 00, 06, 08) the correlation is enhanced. Thus, the correlations in A1 and A2 are 0.8. Therefore, SST plays an important role in the convective activity over tropical South America.

Moreover, a composite plot of precipitation departure for cold cases (Fig 3.3b) and warm cases (Fig 3.3c) can be useful to visualize the horizontal extension of the seesaw pattern which is shown in Fig. 3.1a too. The signal is quite strong in the western and eastern continent, where the local effect of the SST may play a secondary role. In addition, in the continental tropical zone there is a reverse signal between rainfall and SST: wet seasons during cold years and dry seasons during warm years. The signal is stronger during the warm years than the cold years. Later we are going to discuss what possible mechanisms make this reversal behavior possible and explain why the signal during cold years is weak when compared with warm years.

3.3 The east-west Walker circulation impact on South America

The influence of the ENSO on the interannual variability of the rainfall was clearly shown in the previous section. The next question which can be raised is what causes it? As it was proposed, the rainfall regimen over this area is a consequence of multiple time scale processes. Therefore, in this section we are going to analyze the effect of large scale circulation on tropical rainfall. It is very widely known that the ENSO has a global impact (e.g., Kidalis et al. 1989). Ropelewski and Halpert (1987) analyzed the influence of the
Southern Oscillation over global precipitation. Also, Chen and Baker (1986), and Chen (1985) studied the influence of the east-west Walker circulation on the global climate. They claimed that the descendant branch may suppress the convective activity and the ascendant branch may enhance the convective activity. In tropical South America, this can be observed in an east-west Walker circulation plot shown in Figs. 3.4a and 3.4b which portray this fact. Fig. 3.4a shows the composite analysis for warm cases, and Fig. 3.4b shows a composite analysis for cold cases. During warm years a descendant branch is observed which may be responsible for the rainfall inhibition. Also, the corresponding ascending branch is observed during cold years. We want to note that the ascending and descending branches are in reverse relationship, especially at 40°W, near the maximum variance of precipitation in the eastern continent. This is in agreement with the reverse behavior of the rainfall during the cold/warm phases.

Thus, it can be concluded that a possible mechanism to explain the interannual variation of the rainfall regime over tropical South America is the interannual change of the divergence circulation as a response of the SST (Niño 3.4 index) which is portrayed using the east-west Walker circulation. Chen and Yoon (2000) found a similar mechanism to explain the interannual variation of the rainfall regime over Indochina. In their plots (reproduced here, Fig 3.5), potential function of water vapor transport (χQ) superimposed with precipitation (P) is used to illustrate this situation. Water vapor is concentrated in the lower levels, and its transport can help to maintain precipitation. Also, the divergent component of χQ indicates the 'pure' convergence of water vapor. In Fig. 3.5, it can be observed that convergence of χQ coincides with positive anomalous precipitation over the Amazon Basin of tropical South America. A reverse signal, but not clear, is observed for the warm years’
Therefore, the divergent circulation is an important mechanism in regulating the interannual variation of the rainfall over tropical South America, as it was shown using two quasi-independent approaches: the east-west global circulation, and the convergence/divergence of the water vapor transport. In both cases the enhancement/suppression of the rainfall was observed.
3.4 Upper and low level climate configuration

As global divergent circulation can be a mechanism to explain the interannual variation of rainfall, similar results should be noted using weather map analysis, like streamlines. Chen (2003) pointed out that a reverse pattern circulation over monsoon regions exists which was portrayed using the potential function ($\chi$) in 850 and 200 mb, and these patterns are in agreement with the Sverdrup dynamics over tropical areas (Chen 2005). The Sverdrup dynamics approach was first dealt with by Charney (1963) who proved it using the scale analysis method. Later, Chen (2005) portrayed a complete global picture using observational data, like reanalysis. Thus, Chen (2003) and Chen (2005) were able to explain the monsoon dynamics using the Sverdrup’s approach. In this research instead of using $\chi$, the streamline method was employed. Fig. 3.6 shows the streamline at 850 mb for total wind (Fig. 3.6a), wind departure (Fig. 3.6b) for the cold years’ composite, and wind departure for

![Figure 3.5](image_url) Composite global charts of both potential function of water vapor transport, $\Delta \chi_{\text{Q}}$, and precipitation, $\Delta P$, from their long term mean departure for cold a) and warm b) ENSO phase.

As global divergent circulation can be a mechanism to explain the interannual variation of rainfall, similar results should be noted using weather map analysis, like streamlines. Chen (2003) pointed out that a reverse pattern circulation over monsoon regions exists which was portrayed using the potential function ($\chi$) in 850 and 200 mb, and these patterns are in agreement with the Sverdrup dynamics over tropical areas (Chen 2005). The Sverdrup dynamics approach was first dealt with by Charney (1963) who proved it using the scale analysis method. Later, Chen (2005) portrayed a complete global picture using observational data, like reanalysis. Thus, Chen (2003) and Chen (2005) were able to explain the monsoon dynamics using the Sverdrup’s approach. In this research instead of using $\chi$, the streamline method was employed. Fig. 3.6 shows the streamline at 850 mb for total wind (Fig. 3.6a), wind departure (Fig. 3.6b) for the cold years’ composite, and wind departure for
the warm years’ composite (Fig. 3.6c). In each case the corresponding precipitation (total, cold composite departure, and warm composite departure) is superimposed.

In the Southern Hemisphere two subtropical anticyclone systems can be noted, the South Pacific Anticyclone (SPAC) and the South Atlantic Anticyclone (SAAC). During the cold years, the SPAC is intensified as can be observed in Fig 3.6b. This intensification also happens in the Northern Hemisphere with the North Pacific and North Atlantic anticyclone
systems. It seems that those systems work together to maintain a steady and strong easterly wind over the eastern Pacific (Riehl 1954). This type of climate configuration inhibits the development of synoptic perturbation propagating eastward or westward over the eastern Pacific. This fact is discussed in chapter 4. On the east side of the continent, over the Atlantic Ocean, one can note the weakening of the SAAC. This may possibly generate a cyclonic system over the continent due to diabatic heating, and at the same time this low system may induce a low level convergence over the continent.

![Figure 3.7](image)

**Figure 3.7** Same as figure 3.6 except for 200 mb.

During warm years the scenario is quasi opposite. The subtropical anticyclones in the
Pacific Ocean and the anticyclones in the north Atlantic are weaker, so this reduces the steadiness and intensity of the trade winds over the eastern Pacific. It can be observed that a type of westerly departure wind is overlapping with the positive rainfall anomaly. It will be discussed later how climate configuration allows eastward and westward disturbance propagation over the eastern Pacific. In the east side of the continent, the subtropical South Atlantic anticyclone is intensified, and appears as an extension of this anticyclone over the continent. This map may be the surface representation of the mechanism illustrating why the rainfall is suppressed. We had argued about a global divergence, but when a full composition (rotational and divergent) of wind is portrayed, it seems like this anticyclone (and the others) plays an important role. In addition, over the continent around 5°S, 70°W a diffluence wind east of the Andes Mountains is observed. This configuration may play two roles: first, acting as the divergent part of the subsidence branch of the large scale east-west circulation, and second, activating weak convective activity using the steep Andes Mountains as a barrier to generate low level convergence. We believe that this is the reason why the variance, showed in Fig. 3.1c, does not show big amplitude over this area. Because no matter the cold/warm ENSO phase, convective activity can always be formed on this side of the continent.

Moreover, a brief description of the climatological configuration of the upper level wind, using composite analysis is included. Fig. 3.7 shows the streamlines of upper level winds (200 mb) for the total component (Fig. 3.7a), wind departure for cold cases composite (Fig. 3.7b), and wind departure for warm cases composite (Fig. 3.7c). The corresponding precipitation (total, cold departure, and warm departure) is superimposed. In Fig. 3.7a, the Bolivian High and its associated Tropical Upper Tropospheric Trough (TUTT) appears as a main feature. Many researchers attribute the origin of the convective activity to the Bolivian
High (e.g., Lenters et al. 1997), and also numerical model simulation has been used to investigate that the origin of the Bolivian High is due to diabatic heating during the summer season (Nilo et al. 1999). Other researchers attribute the Andes Mountains as a possible force for the Bolivian High formation. Also, Chen et al. (1999) found that not only the diabatic heating over South America, but also the diabatic heating over the African continent are responsible for the Bolivian High formation.

This seesaw pattern in the eastern and western tropical continent should be noted not only in the global divergent circulation, but also in the wind vector (streamlines method). Using wind departure from its long term mean climatology, a diffuence zone can be observed over the eastern tropical continent (Fig. 3.7b). In the same location an anomalous positive precipitation reflects the effect of this diffuence wind. This diffuence wind is sustained by two anticyclone systems, one northern system near the west coast of Africa, and another southern system near the southeast coast of South America. On the contrary, a reverse pattern is observed in the same area for the warm composite (Fig. 3.7), in this case the wind forms a confluence zone which inhibits the cumulus convection as it is observed in the superimposed negative precipitation anomaly. This confluence zone is sustained by two upper level cyclones, in the opposite way it was observed in the cold case.

During cold years, a cyclonic system appears in western tropical continent. In upper levels this cyclonic system acts as a convergence center, so an area of subsidence is formed. This is in agreement with what was shown using the divergent wind and vertical velocity. Therefore, this subsidence wind inhibits upward ascent which can be verified with the observed negative anomaly precipitation (Fig. 3.7b). During warm years the opposite process occurs. Although there is not a clear anticyclonic system, there is a divergence area. The
positive anomalous precipitation certifies our explanation.

In conclusion, the seesaw rainfall pattern during cold/warm years can be explained independently using upper wind circulation. The confluence/diffluence wind and the cyclonic/anticyclonic systems show an interesting phase reversal in harmony with the
reversal phase of the warm/cold ENSO. This is in agreement with what was described in lower levels (850 mb).

3.5 The interannual time scale process

Finally, this interannual variation of the rainfall regime over tropical South America is part of a complex multiple time scale interaction process. Although, we have described that the SST plays an important role in these multiple scale processes, there are other processes
embedded in this interannual scale variation. Therefore, in the next chapter those processes are going to be described in detail. However, we want to end this chapter with a plot that portrays the interannual variation of the whole latitudinal section (from 180° to 20°E degrees). This plot is shown in Fig. 3.8 for DJF and in Fig. 3.9 for FMA. Fig 3.8 is included to show the comparison between the mature SAMS monsoon life cycle and the analyzed season (FMA). The main difference between DJF (Fig. 3.8) and FMA (Fig. 3.9) season is that during DJF there is a weak connection between the central Pacific (180° degrees) and western South America. This disconnection is more notable especially during weak warm years (1987, 1992). Thus, during FMA a continuous precipitation line which links the central Pacific and western South America can be observed. It will be discussed further in detail the kinds of perturbations that make this possible.
CHAPTER 4. THE RAINFALL REGIME OVER TROPICAL SOUTH AMERICA
ASSOCIATED WITH ITS MULTIPLE SCALE PROCESS

4.1 Introduction

Eastward and westward rainfall propagations have been identified in the tropical Amazon Basin. Kousky (1980) and Cohen et al. (1995) found a type of westward propagation perturbation which originates in the mouth of the Amazon River. Kousky (1980) claimed that the mechanism responsible for this perturbation is diurnal cycle forcing. Also, Cohen et al. (1995) studied several cases of those westward disturbance propagation and relates their continental incursion to squall line systems. Using the Geostationary Operational Environmental Satellite (GOES) imagery, Cohen et al. (1995) portrayed the inland penetration of rainfall, which traveled 2 days to arrive over the western Amazon. Although, the convective disturbance propagation is very clearly observed, no clear evidence of a synoptic perturbation moving into the continent was presented, because the vertical wind component used to illustrate the penetration was very noisy.

Nuyen and Duvel (2008), using OLR (Outgoing Long Radiation) data sets and EOF (Empirical Orthogonal Functions) analysis, identified eastward propagation perturbation generated at the western Amazon which triggered rainfall over western Africa and western South America. Using a spectral analysis technique, they found that the eastward perturbation seems to be associated with a 5-6 day mode frequency. In addition, Wang and Fu (2007) showed that a convective-coupled Kelvin-like wave could be responsible for this propagation over the Atlantic Ocean. Recently, Liebmann et al. (2009) identified two types
of eastward propagation. One originated over the Pacific Ocean and the other over the western Amazon.

Precipitation exhibits a strong quasi-periodic behavior as was illustrated by Chen et al. (1995a) and Chen et al (1995b) on a global scale. The structure of these quasi-periodic perturbations show eastward and westward propagation properties which are determined to constrain the range of rainfall variation over a regional scale such as the tropical Amazon Basin. Thus, those precipitation patterns described in previous paragraphs must be associated with synoptic disturbances, and those synoptic disturbances must act under the influence of low frequency periodic phenomena; however, few efforts have been made to relate those high frequency perturbations with the low frequency warm and cold ENSO phase over the tropical Amazon Basin.

Therefore, in this chapter we are going to explain the possible mechanism for which the rainfall regime over tropical South America is generated using a downscaling approach, which consists in an analysis of different time scale processes like the intraseasonal, synoptic scale, and the diurnal variation. Also, the results from the interannual variation from the previous chapter are applied.
4.2 Modulation of the precipitation over the Amazon Basin by the global 30-60 and 12-24 day modes

It has been proven that the 30-60 day mode has an impact on global circulation. After the first indication of its existence by Madden and Julian (1971), many publications have confirmed the importance of this global perturbation mode, and a lot of research has been conducted to investigate the effect of the 30-60 day mode on diverse meteorological phenomena (e.g., Chen et al. 1995a). Also, Chen et al. (1995b) showed for first time that the 12-24 intra-seasonal mode plays an important role in global precipitation. Using two significant cases, they showed that in the Asian-Australian monsoon system area the 12-24 day mode plays an important role in the maintenance of the seesaw behavior between the east-west hemispheric AA-monsoon (Asian-Australian Monsoon System) and extra-AA-monsoon. They concluded that the propagating properties of those intraseasonal modes help to explain the maintenance of the seesaw oscillation due to the convergence of water vapor transport. Therefore, in this section we are going to discuss the role of the 12-24 and 30-60 day modes in the modulation of the rainfall regime over the Amazon Basin affected by the influence of the cold/warm ENSO phase.

4.2.1 Interannual scale separation

The primary source of data used in this section is precipitation in different time scales like yearly, monthly, pent-day, daily, and hourly. The information was not available from a single source and format. Thus, monthly and pent-day precipitation data sets were used from CMAP, and daily and hourly data sets were taken from TRMM. Since there is not reliable daily precipitation data set for South America before 1998, OLR was used as a proxy for
precipitation. We assumed the approach by Arkin and Ardanuy (1989), who defined $\Delta OLR=235-\text{OLR}$ as a good estimation for rainfall. OLR was compared with precipitation showing acceptable results. Moreover, in all x-t plots of interannual variation show (Figs. 4.2.3 to 4.2.10), the time series were detrended to avoid spurious representation of the precipitation departure. In addition, when it was needed the data was filtered using the technique proposed by Murakami (1976).

As it was discussed in the previous chapter, the interannual variation of the rainfall regime has a strong influence on the ENSO. Therefore, in this research, we have separated the analysis according to the phases of the ENSO (Fig. 3.2), and these phases are sub-divided by strong and weak events. The strong warm cases are years 1983 and 1998, and the weak warm cases are years 1987 and 1992. Although, years 2003 and 2007 have a positive anomaly when they are compared with the index area Niño 3.4, we did not consider them as a strong or weak case because the rainfall pattern over the continent is

![Figure 4.1](image)

Figure 4.1 The variance of 5-day mean precipitation seasonal (FMA) composite for warm (a), cold (b), and neutral (c) ENSO.
not close to warm ENSO phase cases. On the other hand, strong cold cases are years 1989 and 2000, and weak cold cases are 1985, 1996, and 2006. Although, 1985 shows a strong signal in the ΔSST interannual histogram (Fig. 3.2), it is only focused in the western continent (Area 1), so 1985 is considered a weak cold case. Using a x-t diagram (Fig. 3.8b), it can be observed that 1989 and 2000 have more uniform rainfall over the whole tropical continental area, so they are chosen as strong cases.

Two areas of maximum precipitation over continent are observed in the Hovmöller diagram shown in Fig. 3.10a. It shows the seasonal mean (FMA) average from 5°S to EQ over the oceanic area and the average from 10°S to 5°N over continental area. It can be observed that those centers of maximum precipitation show an interannual variation. The one located at 73°W does not show a strong linear relation with the Niño 3.4 index (correlation < 0.4), and the other located at 50°W shows a good correlation with the ΔSST (correlation = 0.8). The former area has less precipitation, and the latter has more. Those centers of precipitation will be the focus of further attention when we shall discuss the formation of cumulus convection that has the property to propagate eastward (from the western maximum) and westward (from the eastern maximum). Therefore, the analysis of the intraseasonal rainfall regime over tropical South America is divided into warm and cold ENSO years.

4.2.2 Identification of maximum perturbation areas

In order to analyze where maximum perturbations are located, the variance of 5-day mean rainfall is computed. The 5-day mean acts as a low-pass filter in order to eliminate signals lower than 5 days where synoptic disturbances are included. The 5-day mean precipitation variance is shown in Fig. 4.1, and is grouped into warm (83/87/92/98), cold
(85/89/96/99/00/06), and neutral (all – [cold+warm]) cases. The composite was made over the primary rainy season FMA. It can be observed that there are two different types of connections to the Amazon Basin, one is from the west during warm years (Fig. 4.1a), and other is from the east during cold and neutral years (Figs. 4.1b and 4.1c). These results show that the rainfall regime over the Amazon basin responds to two different forcing: one during warm cases coming from the Pacific Ocean, and the other during cold cases coming from the Atlantic Ocean. We are going to prove that this sub scale (5-day mean) behavior has an important connection with the modulation of the interannual variation of the upward/downward branch of the west-east Walker Circulation. The west-east Walker Circulation suppresses tropical convection in its descendant branch and enhances them in its ascendant branch. We are looking to find in what way this release/suppression mechanism is expressed in the intraseasonal time scale.

Figs. 4.1b and 4.1c show similar patterns, which is logical because cold cases are the enhancing state of the neutral years. However, it can be noted in Fig. 4.1b that in eastern 40°W and western 60°W the perturbation is more intense during cold years than neutral years. This is a good indication that the convective activity increases in the areas of maximum rainfall observed previously in the seasonal mean. The consequence of this fact will gain importance later when the formation of eastward and westward disturbance propagation over the continent will be explained.

On the other hand, Fig. 4.1a shows the 5-day mean variance for warm cases. Two maximums are observed. One maximum center is located at 50°W at the mouth of the Amazon River. The rainfall is suppressed during the warm phase of the ENSO according to the global divergence circulation, but the strong diurnal cycle of precipitation over the mouth
of the Amazon River can not be suppressed, so this is the reason why the perturbation signal is strong. It will be shown later that this fact plays an important role in the formation of the eastward disturbance propagation during warm years. The other maximum is observed offshore from the western continent, which is seen as an extension of the maximum zone in the central Pacific. We shall discuss the role of this maximum variance precipitation in the analysis of synoptic systems under the effect of the intraseasonal global mode. Therefore, perturbations are located mainly in the western and eastern continent, but their intensity and extension varies according to the phase of the ENSO.

4.2.3 Identification of the regional 12-24 and 30-60 day modes

Madden and Julian (1971) discovered the existence of a global 30-60 day mode, and later Chen et al. (1995b) showed the existence of the global 12-24 day mode. The question we need to ask is: are these low frequency disturbances presented in our analysis dataset? Although, we have shown that there is an important interannual variability of precipitation over the Amazon Basin, other sources of perturbation may exist in a sub-range scale of the interannual.

To investigate what low frequency disturbances are presented, a spectral analysis is performed. It was demonstrated by Chen et al. (1995a) and Chen et al. (1995b) that the 12-24 and 30-60 day modes are global, and they proved it using a homogeneous field of global precipitation estimated from satellite data at the Goddard Laboratory for Atmosphere. Although they did an analysis only for the years of 1979 and 1980, the general conclusion was that the 12-24 and 30-60 day modes have global implications in the hydrological cycle, especially in its east/west representation. Therefore, we are going to perform a power spectral
analysis over the Amazon Basin focusing the areal average of the precipitation proxy over the maximum variance described in the previous section, this is because the areal mean acts as a filter, so it can reduce the amplitude of the signal in the power spectral. The power spectral plots are shown in Fig. 4.2 for: warm (Fig. 4.2a), cold (Fig. 4.2b), and neutral (Fig. 4.2c) years. The black solid line is the 95% confident level, which means that peaks over the line are statistically robust and significant. As can be observed, all years have both the 12-24 and 30-60 signals. Therefore, the intraseasonal oscillation (12-24 and 30-60) is observed in the rainfall pattern of the tropical Amazon Basin. Also, the amplitude of the 30-60 mode is bigger than the 12-24 as was pointed out by Chen et al. (1995b). This fact can be observed in years 1983 (W), 1985 (C), and 1982 (N). However, in some cases the 12-24 signal can dominate (e.g., 2008, 1988, and 1984). Chen et al. (1995b) concluded that a tendency of exclusion exists between 30-60 and 12-24.

Although, the signal for the 12-24 and 30-60 day modes is present in every case, in some years the peak is weak. For example, 1998 has a weak 30-60 peak, and 2000 has a weak 12-24 peak. This happens because exist processes associated to the 12-24 and 30-60 mode which govern the rainfall behavior. Thus, strong peaks dominate against the weak ones, and clear evidence can be observed in Fig. 4.3, where the 5-day [ΔP] running mean and the filtered 12-24 and 30-60 of [ΔP] are superimposed. In those examples one mode dominates over the other, which is reflected in the power spectral representation, so both signals exist.

Moreover, the contribution of both signals is very important in the rainfall pattern. For example, figure 4.3b shows the combination of the 30-60 band-pass filtered,$\tilde{P}$, and the
12-24 band-pass filtered, $[^\hat{P}]$, of precipitation for years 1998 and 2000. In both cases, if they
are compared with the total time series, Fig. 4.3a, they match very well. Thus, the
contribution of both signals to the total time series is 42.0% and 32.8% for year 1998 and
2000 respectively (Fig. 4.4). Also, the maximum peaks in the total time series for both cases
are in good agreement with the maximum peaks of $[^\hat{P}]$ (Fig. 4.3d). The 30-60 and 12-24 day
modes are present in both warm and cold cases of the interannual variation, and they are not
an exclusive characteristic of the cold or warm ENSO phase. Therefore, the reason for the

![Graph showing time series of precipitation departure and band-pass filtered precipitation]

**Figure 4.3** Left panel: a) the time series of 5-day mean of precipitation departure, \( [\Delta P] \), for 1998, b) the combination of the 30-60 band-pass filtered, \( [\hat{P}] \), for 1998, and the 12-24 band-pass filtered, \( [\tilde{P}] \), of precipitation for 1998, c) the 30-60 band-pass filtered, \( [\hat{P}] \), of precipitation for 1998, and d) the 12-24 band-pass filtered, \( [\tilde{P}] \), of precipitation for 1998. Right panel: same as left panel except for year 2000.

different rainfall pattern could be the way they interact with the diurnal cycle and other high
frequency modes. An interesting feature appears, which is the tendency of exclusion between
\( [\hat{P}] \) and \( [\tilde{P}] \). This was noted before by Chen et al. (1995b). He suggested that the not well
coupled between Kelvin waves and cumulus convection as a possible reason for this
processes. Although this question is beyond the scope of this investigation, we can do some
comments: as we can observe, the contribution of $\hat{P}$ and $\tilde{P}$ together is very important and good enough to portray the seasonal behavior of the total time series. Also, $\hat{P}$ matches very well with the peak of maximum rainfall. Later it will be shown that the eastward propagation is embedded inside the positive phase of $\hat{P}$. Also, a well organized pattern of propagation is observed in the positive phase of $\hat{P}$ (Figs. 4.13f and 4.14f). Therefore in some way the atmospheric circulation is coupling to organize the cumulus convection, which only happens during the positive $\hat{P}$ phase. Inside the positive $\tilde{P}$ active precipitation is also observed, but not well organized as in the $\hat{P}$ phase. Therefore, we speculate that the exclusion the exclusion of $\hat{P}$ and $\tilde{P}$ is related on how much efficient the rainfall production is needed, which can be linked to how much water vapor is available in the system in a specific time period.

4.2.4. Intraseasonal modes and the ENSO modulation

Early in this chapter, it was established that the 12-24 and 30-60 day modes exist and play an important role in the rainfall pattern over the Amazon Basin. However, we did not comment about the characteristics of these intraseasonal modes on the modulation of the rainfall behavior. Although, the power spectral analysis portrays that the intraseasonal mode exists in warm and cold years, we are going to show in this section how the horizontal distribution of rainfall is modulated by those intraseasonal modes under the influence of the interannual variability caused by the ENSO. Thus, the analysis is divided into two parts: warm and cold cases.
4.2.4.1 Warm cases

The effect of the intraseasonal mode on rainfall can be analyzed splitting the entire time series into its 30-60 and 12-24 day modes. To do this, a band-pass filter is applied to the precipitation proxy. This precipitation proxy is formed by averaging daily ΔOLR over the Amazon Basin between 5°S to 0° and 80°W to 40°W. We are going to analyze common features observed in Figs. 4.5 and 4.6 which collect strong and weak warm cases (x-t for warm years 1983, 1987, 1992, and 1998).

The first characteristic appearing in all cases is the activity convection west of 80°W, which is more intense during strong cases (1983, 1998) than during weak cases (1987, 1992). Also, those perturbations propagate eastward with the center of the formation located around 160°W in most of the cases. In 1998, an early activation of the eastward rainfall propagation is observed compared with the other strong warm case (1983). The second characteristic is the modulation pattern. The first box for every case represents an x-t diagram with a 5-day resolution. The envelope of the 30-60 mode is clearly noted modulating the propagating extension of the precipitation. Thus, the minimum coincides with the inactive phase (dry),
and the maximum with the active phase (wet).

The number of eastward propagating perturbation varies with the intensity of the ENSO. Thus, during weak warm cases, only a few cases of eastward perturbation over the ocean can be observed. This can be also noted in the interannual variation (Fig. 3.10), and inferred from the small amount of rainfall over that area during 1987 and 1982 compared with 1983 and 1998. For example, strong cases like 1998 and 1983 have 5 and 4 cases of
propagation respectively, and for weak cases 3 (1987) and 1 (1992) propagating perturbations are observed, so the more (less) intense ENSO means more (less) number of propagating cases. When these propagations reach the west coast of tropical South America (~80°W), they stop in some cases, and continued propagating in others. Cases of stopping propagation can be observed in Fig. 4.5c, for the strong case, during Feb/1 to Feb/15, and a case of continue propagation is observed in Fig. 4.6c for a weak case (Apr/1). Thus, precipitation over eastern Pacific is produced by the westerly propagation system. The mechanism formation of precipitation over the central Pacific during warm ENSO was discussed by Chen et al. (2008). They found that the influence of subtropical anticyclones is the main source of rainfall forcing. Therefore, the seasonal rainfall pattern over the eastern Pacific is wet due to the increased number of these perturbations which enhance the convective activity.

Stopping propagation cases are more common during strong than weak ENSO. This happens because the strong suppression of the rainfall over the continent during extreme warm cases. Also, if the propagation continues moving to the continent with the same intensity, the cumulus convection developed could reverse the downward branch of the east-west Walker circulation and end the warm case configuration. Thus, the seesaw pattern between eastern continental and eastern Pacific is maintained by this propagating property of the low frequency mode. During weak warm years, not only more cases of continuous propagating states are observed, but also continental eastward perturbations generated in the western tropical South America are noted. However, those perturbations are not generated at 160°W as in the strong cases. The more existence of these perturbations over the continent contributes to a reduction of rainfall during the warm years.
When the 30-60 band-pass filtered \(\Delta\text{OLR}\) is superimposed, its maximum values coincide with the maximum rainfall over the continent. This happens not only for strong cases, but also for weak cases. Thus, the 30-60 filtered data becomes the envelope of the modulation pattern. Therefore, we can conclude that the 30-60 day mode modulates the rainfall over Tropical South America. The perfect example can be observed in 1987 (Fig. 4.6a). Four maximum peaks are observed, which match well with the four cycles of maximum rainfall. The period of the cycle is approximately 40 days, which is between the 30-60 day mode. However, during 1998 (strong warm) the 30-60 band-pass filtered time
series does not show a clear relationship, but the 12-24 band-pass filtered time series for the same area corresponds well with the diurnal Hovmöller diagram (Fig. 4.5d).

However, what is the dynamic forcing responsible for driving those perturbations? It can be observed that all these cases have a quasi-cycle behavior with a wet and dry period. Hence, we have chosen a complete cycle and analyzed what happens in both the maximum (wet) and minimum (dry) phases, and then the 30-60 filtered potential function is computed for 200 and 850 mb. The results are shown in Fig. 4.9, which shows an inverse relationship between both the dry and wet phase. An inverse relation between 850 and 200 mb is also shown in each phase. In the dry phase (box (b)), the configuration is a divergence
in low level and a convergence in upper level (suppression), and in the wet phase (box (c)),
the configuration is opposite compared to the dry case: convergence in low level and the
divergence in upper level (enhance). Therefore, the maximum/minimum oscillation property
of this low frequency precipitation mode is dynamically controlled by the 30-60 day mode of
\( \chi \).

In addition, the east-west 30-60 filtered divergent circulation is plot to illustrate the
downward motion during the dry phase and the upward motion during the wet phase, in both
cases the upper and lower levels are well coupled. According to Chen et al. (1995b),

Figure 4.8 Same as figure 4.5 except for the years 1985 and 2006.
precipitation is modulated by the 12-24 day mode too, also an eastward propagation of $\chi$ is observed in agreement with the rainfall perturbation, and the maintenance of these propagations is due to water vapor transport (Chen et al. 1995a, 1995b).

The 5-day mean acts as a low-pass filter for precipitation, so high frequency perturbations are removed. Although this is useful to analyze low frequency modes, we need to observe the interaction between 12-24 and 30-60 day mode. Thus, a 1-day mean x-t diagram is plotted in Figs. 4.5b, 4.5d, 4.6b, and 4.6d, and the 12-24 filtered areal precipitation index is superimposed in every case. In all cases (strong and weak), it can be observed that the intraseasonal 12-24 mode is embedded in the 30-60 day mode. For example, in the strong case 1998, the rainfall during 3/25 to 4/5 is not only controlled by the 30-60 day mode, but also by the 12-24 day mode as it is observed in Fig. 4.5d. There, the maximum perturbation state rainy coincides with the maximum amplitude of the 12-24 filtered time series. Moreover, in all the cases the maximum rainfall over the continent matches well with the maximum of the 12-24 filtered time series, which happens in both cases penetrating and non-penetrating. A good example is shown in the weak case 1992 (Fig. 4.6d). Three maximum rainfall areas are in agreement with the maximum amplitude of the 12-24 filtered time series. In the 5-day mean, it is observed that the maximum rainfall coincide with the maximum 30-60 filtered time series. This finding gives us the supporting material to claim that the rainfall over Tropical South America is modulated by the 30-60 intraseasonal global mode. However, amplifying our scope of observation another interesting feature emerges: the 12-24 intraseasonal mode control the intensity of the rainfall.

Therefore, it can be concluded that the intraseasonal mode strongly modulates the rainfall over Tropical South America during strong and weak warm ENSO. Also, some
propagating properties of precipitation can be attributed to the intraseasonal modes.
4.2.4.2 Cold cases

Rainfall during cold cases shows an opposite relationship compared to warm cases. As it was shown in the previous chapter, the rainfall suppression in the eastern Pacific due to the Walker circulation during cold years can be observed in the x-t diagram of proxy precipitation (Figs. 4.7 and 4.8). West 80ºW, the area is dry as it was presented for warm cases; that is, no rain over the eastern Pacific. On the contrary, inside the continent (80ºW~40ºW) the rainfall is very intense in all the cases.

The rainfall is driven by the 30-60 day mode in every case. This can be observed using the 30-60 band pass filtered time series superimposed with the 5-day mean x-t diagram. In all cases, the maximum rainfall inside continent shows a good relationship with the peaks of the filtered time series. For example, in cases 1985 (Fig. 4.8a) and 2006 (Fig. 4.8c) the two peaks in the time series form the enveloping of the rainfall, where the maximum match well with the peaks of filtered [ΔOLR].

Eastward/westward propagations observed in the 5-day mean x-t diagram (Fig. 4.8a) are driven under the influence of the 30-60 day mode as it is inferred from the 30-60 filtered time series. Inside the 5-day propagations synoptic perturbations are embedded as it is shown in the daily timescale (Fig. 4.8b). For example, during the period Mar/1 to Apr/15, it can be noted that this 5-day mean westward propagation is coupled with the eastward synoptic propagation of a shorter time period (~5 days). The coupling of these signals portrays a typical rainfall pattern during a cold ENSO phase. We want to note that when perturbations of opposite propagations emerge more rainfall is expected than in cases when propagation came from the same side. As can be noted in the case of Apr/15 to May/1, one large storm can be tracked, but the others around are weak. Therefore, the rainfall production might be
linked to the propagation property of the disturbance.

Although the 5-day mean portrays a general feature of the rainfall, the 1-day mean reveals other characteristics. The principle is that 12-24 day mode is linked to the maximum rainfall, which coincides with enhanced eastward propagation. In all the cases presented, it can be noted that the eastward propagation is more intense when it coincides with the positive phase of the 12-24 filtered time series. This does not mean that there are not cases of eastward propagation outside the positive range of 12-24, but those cases are weak.

Also, we want to point out that for every 30-60 cycle there is at least two 12-24 cycles, and in each 12-24 cycle there is at least 1 big storm moving eastward. As in warm cases, this big storm can be comparable to the inland rainfall penetration. Therefore, we can speculate about some kind of teleconnection from the central Pacific through Tropical South America during cold years. This teleconnection is clearly observed on the surface during warm years; however, in cold years it may be noted in the upper levels.

When both the 12-24 and 30-60 day mode are in phase the rainfall over tropical South America experiences a maximum enhancement. For example, in case 1985, it can be seen how the peak of 12-24 and 30-60 approach at the same time. In most of the cases the 12-24 day mode is out of phased from the 30-60 day mode. It is common that the maximum 30-60 mode coincide with the minimum 12-24 and vice-versa. The consequence of this configuration is that more cases of westward disturbances are forced by the 12-24 day mode and others by the 30-60 day mode.

A special case is year 1985, which is wetter than 2000 over the eastern part of the continent (~40ºW) [Fig. 3.2b]. What make cold cases have intense rain using a perspective of the intraseasonal mode? This can be due to the different way in which the 12-24 and 30-60
day modes interact. If we observe the four cold cases presented, in case 1985, the two maximum peaks of 30-60 component are in phase with the maximum peaks of 12-24 component, and coinciding with this constructive interference, three cases of enhanced storms propagating eastward can be observed. Nevertheless, in the other cases the 30-60 and 12-24 components are delayed in temporal phase one to each other. The consequence of this special arrangement is that in the case of constructive interference (1985) strong cases of storm can be observed as the one which hits Brazil early in this year, but this constructive interference also produce prolonged dry periods as is noted in case 1985. According to the interannual variation of the FMA season (Fig. 3.2b) year 1985 is the highest peak of our record, but when it is compared in a longitudinal variation (Fig. 3.10b) it is observed that this high precipitation only occurs over the eastern continent, and the western continent looks as dry as in warm years. In cases when the interference is not coherence there is a suppression of rainfall or destructive interference state. What occurs is that rainfall is less enhanced, but more storms are formed. For example, the case 2006 (Fig. 4.8d), between Feb 15 to Apr 1, the 30-60 component is dephased half 12-24 cycle, in the way that the 30-60 maximum coincide with the 12-24 minimum. This configuration allows the formation of storms around the peak of both the 30-60 and 12-24 components, and more storm cases are observed. Thus, in case 2006 six storms are observed in half 30-60 cycle and only one storm in 1985 is observed.

The eastern side of the tropical continent is the end part of the eastern propagation. Therefore, the difference in the rainfall amounts between 1985 and 2000 should be related to this characteristic. In case 2000, the peaks of the 30-60 time series is not in phase with the peaks of the 12-24 time series. As it can be observed (Fig. 4.7d), the 12-24 maximum is
located ahead or behind the maximum signal of the 30-60 time series. The same pattern can be observed for years 1989 and 2006. However, in some cases, the maximum of 12-24 and 30-60 approach very much as in Apr/25, and when this happens an enhanced and prolonged rainfall propagation is expected. These maximums coincide with a long extension penetration case observed in Fig. 10, which is enhanced in the eastern side of the continent via the strong diurnal cycle (Kousky 1980). It was shown that the forcing is driven by a 12-24 and 30-60 low level convergence and upper level divergence illustrated with the potential function ($\chi$). Therefore, eastern maximum rainfall observed in 1985 is due to a high order of phase coherence between the 12-24 and 30-60 day modes.

In conclusion, the following aspects can be highlighted regarding the influence of the intraseasonal mode over the rainfall region in the Tropical Amazon basin associated with the warm/cold ENSO phase: 1) the 12-24 and 30-60 modes are observed in both cold and warm ENSO phases, 2) the 12-24 and 30-60 day modes were observed to control the maximum amplitude of the rainfall; also, the dry period corresponds well with the minimum amplitude of the 30-60 day mode, 3) at least 2 maximum 12-24 perturbations can be found embedded in a single 30-60 mode, 4) during cold years, the maximum rainfall is observed when the phase difference of 12-24 day mode and 30-60 day mode is minimum, 5) during warm years the maximum inland penetration of the eastward propagation is associated with a maximum phase of the 12-24 day mode, and 6) during extreme cold years a double co-existent eastward perturbation over the continent is observed during the maximum phase of the 12-24 mode.
Same as Figure 4.9 except for the year 2000.
4.3 The Multiple time scale processes

Processes at two time scales have been identified as being responsible for the rain-producing system over tropical South America. Thus, the interannual variation due to the ENSO was the main constraint discussed in chapter 3, and the role of the intraseasonal modes (12-24 and 30-60 day) was also analyzed earlier in this chapter. The main conclusion of the latter was that rainfall is strongly modulated by the propagating property of the intraseasonal modes. Thus, the rain-producing system over tropical South America is a complex system that involves different ranges of time scale processes. However, the role of the synoptic and high frequency modes were not discussed, and neither the function of the diurnal cycle. Therefore, in this section, we are going to analyze how the interaction of synoptic disturbances propagating systems contributes to the enhancement/suppression of rainfall in tropical South America.

4.3.1 High frequency disturbances

The annual and diurnal cycle constitutes the most clear cycle variation in the atmospheric circulation. Therefore, the understanding of how this mode of perturbation can affect superior modes is important in order to portray a complete picture of the rainfall pattern. We need to identify whether exist perturbations of high frequency in both the cold and warm ENSO years. Thus, the variance of precipitation is computed to have a general view of how these diurnal and high frequency perturbations behave in tropical South America. The left panel of Fig. 4.11 shows the daily precipitation variance, where precipitation was estimated from satellite by TRMM. Three different composites are shown for warm, cold, and neutral years. The only year for the warm composite was 1998 because
the TRMM data set began in Jan 1998, but the other cases include a span of 4 years.

Perturbations over the tropical continent are limited for warm cases, but some perturbations clustering in a small area close to the mouth of the Amazon River are noted. This area is distinguished as an important source of dynamic forcing to generate westward propagation. On the other hand, the perturbation activity is more intense over the oceanic area in the eastern Pacific. These areas of intense activity are in agreement with the ENSO signal during warm years. On the contrary, perturbations over the Pacific Ocean are suppressed for cold years, and the continental area looks more active. Those cases are in
agreement with the east-west global (enhance/suppression) circulation. Neutral years have a similar behavior to that of the cold ones, but the intensity of the perturbation is lower than in cold years, as it is expected. In addition, we want to highlight the existence of two maximum centers over 73°W and 40°W, which coincides with the areas where the eastward and westward propagation are triggered. Therefore, perturbations are intense over eastern Pacific during warm years and over the continent during cold years.

4.3.2 The power spectrum

A power spectral analysis should show a pronounced peak around the frequency where the signal is more important. Thus, we can be able to identify what is the frequency of the perturbation previously found in the precipitation variance plot. The power spectrum analysis is computed using an areal average precipitation index, and the area is selected taking as reference the maximum value of variance over the continental area shown in Fig. 4.11. In this computation, 3 hour data sets were used in order to portray the amplitude of the diurnal mode. Fig. 4.12 shows the power spectrum for warm, cold, and neutral years. It can be observed that the diurnal cycle is present in all the cases, as it was mentioned earlier the diurnal cycle is the most important climate signal.

However, during cold years like 2000 and 2008 the amplitude of the diurnal mode is reduced and other modes appear with important impact on the total spectrum. For instance, in 2008 two peaks are clearly observed in the high frequency range. The first is located around 6-day and the second one at 4-day. Therefore, more perturbations can be expected between ranges of 3-6 days for that case. This fact opens the option to argue about the interaction/coupling between the diurnal mode and high frequency modes. As it was
established early, the enhancement of the wet rainy season should be due to some type of interaction of processes, so during cold years the way the system responds is through an activation of high frequency mechanisms. They couple not only with the diurnal variation, but also couple with low frequency modes such as the intraseasonal. On the contrary, during warm years those high frequency modes are suppressed, as can be observed in the power spectral plot for year 1998, so the inhibition of the cumulus convection might be through the suppression of the synoptic mode. The diurnal cycle can not be totally suppressed due to its natural climate domination as it is shown in the power spectral plot. However, a pronounced peak in the diurnal cycle for 1998 compared with the other cases is observed, so there is

Figure 4.12  The power spectral of 3-hour mean rainfall [TRMM] averaged over the Amazon Basin for warm (a), cold (b), and neutral (c) ENSO years. The lower black curve in each plot represents the 95% confident level, while the upper black curve represents the 99% confident level.
some degree of variation in the diurnal cycle.

Other notable differences between cold and neutral years observed in the power spectrum is the existence of a single peak in the high frequency in neutral years (2002, 2004) and consecutive double peaks observed in cold years (2000, 2008). The year 2000 as an extreme year is very wet, and we found that double peaks in the power spectral are a response to a double coexisting eastward propagation (Fig. 4.14f), which has the function to release more latent heat and rain during the extreme cold ENSO. On the contrary, neutral years, like 2004 or 2002, do not show many cases of double eastward perturbation, and less rain is produced as a consequence. Also, comparing the high frequency mode (6-2 day) during 1998 and 2000, the amplitude in 2000 is bigger than in 1998.

4.3.3 The multiple time scale processes during the ENSO

Up to this point several ranges of perturbation were found in the power spectral analysis for low and high frequencies, which should be responsible for generating the rainfall pattern over tropical South America. However, how do they act together? And how do they couple under the influence of the interannual variation of the ENSO? In this section, we are going to discuss those issues.

Warm

Fig. 4.13 shows a graphical representation of the interaction of multiples scales. The seasonal FMA rainfall for 1998 is shown, and the enhanced area over central and eastern Pacific is observed (Fig. 4.13a). The seesaw between central-eastern Pacific and tropical South America is very well depicted in the seasonal FMA departure (Fig. 4.13b). In Fig.
4.13c, it can be observed that the anomalous precipitation over continent is produced by eastward and westward propagations controlled by the intraseasonal modes (30-60 and 12-24). Also, a deficient amount of rainfall over the continent is shown by the histogram in the lower panel. We concentrate in two cases of eastward propagation (Fig. 4.13d). Case 1 propagates along the maximum phase of the 30-60 day mode, and case 4 propagates along the minimum phase. In Fig. 4.13e, we observe that together with cases 1 and 4 emerge a cluster of eastward propagations which correspond to a synoptic scale perturbation. The interesting finding is that those synoptic eastward perturbations are embedded in the 12-24 day mode perturbation. A 3-hour time step x-t diagram is presented in Fig. 4.13d. There, westward disturbances are identified, which were not observed in any of the previous plots. The important point of the rain-production over tropical South America is formed by the interaction of the high frequency westward perturbation and the synoptic eastward perturbation. It can be observed that for this warm year that interaction is weak, which can be observed in the predominant effect of the diurnal mode portrayed by the westward rainfall perturbation. This is the reason why there is deficit of rainfall, and also it might explain the mechanism how the propagation observed in the intraseasonal scale is stopped. Thus, the system responds suppressing the synoptic perturbation, because the diurnal cycle can not be suppressed efficiently. Therefore, this type of weak interaction is fundamental to maintain the rainfall regime during the warm phase of the ENSO.

Cold

The graphical representation of the multiple scale processes for the cold year 2000 is shown in Fig. 4.14. Similar to year 1998, Fig. 4.14a and 4.14b portray the FMA seasonal
rainfall, and a seesaw pattern between central-eastern Pacific and tropical South America is observed. In this case, the seesaw is reverse to the seesaw for the warm year 1998. Two features can be observed in Fig. 4.14c. First, the eastward/westward perturbation over central-eastern Pacific is suppressed. Second, the intraseasonal rainfall over the continent is enhanced, which is shown in the histogram at the bottom. A kind of eastward/eastward propagation similar to the warm case is shown, but maybe not clearly portrayed because the longitude of the propagation is short. Two cases of eastward propagation are selected from Fig. 4.14d. Cases 5 and 7 are embedded in the maximum phase of the 30-60 intraseasonal mode as it is verified by the superimposed 30-60 filtered time series. Two more eastward propagations emerge near case 5 and 7 (Fig. 4.14e), and the amount of precipitation is more intense than in warm cases. The synoptic eastward disturbances are embedded in the 12-24 intraseasonal mode as it can be observed with the superimposed 12-24 filtered time series. The 3-hour x-t diagram (Fig. 4.14f) is used to portray the role of the high frequency westward perturbation identified in the warm case, and the following features emerge. First, the high frequency westward perturbation exists also in cold cases. Second, it is strongly coupled with the synoptic eastward perturbation. Third, precipitation is enhanced because the coupling between those propagating systems. Fourth, the westward disturbance can link two consecutive eastward propagating systems in the way that a maximized double-coupled eastward propagation is observed. Fifth, the distance traveled by the westward propagation is shorter than its reciprocal in the warm scenario. Therefore, during cold years exists a positive interaction between the high frequency westward propagation and the synoptic eastward propagation. The latter is not suppressed, and the ways both systems interact define the rainfall regime during the cold ENSO phase.
4.3.3.1 Synoptic perturbations

Other characteristics emerge when high frequency ranges are included. Thus, Fig. 4.13e and 4.14e show a daily Hovmöller diagram of precipitation in each figure analyzed in the previous section. Sub cases of eastward disturbance propagation embedded in the main intraseasonal eastward propagation. The eastward propagations are located over the continent inside the intraseasonal modulation, the intraseasonal mode define periods of suppression and enhancement of rainfall too. Although, over the continental there is a high instability order, regular patterns can still be noted when the 12-24 mode is presented. In the next section, it will be discussed what synoptic forces are responsible for this type of process; in advance, we can say that different types of low level convergence occur in the eastern side of the Andes Mountains to trigger those perturbations.

The diurnal scale of the Hovmöller diagram for the cold and warm cases shows interesting features of how the synoptic scale responds against the ENSO. First, the number of eastward propagations in cold cases is larger than in warm cases. For example, in a 30-60 cycle, at least 8 eastward propagations are identified for the cold case (Fig. 4.14e), against 3 eastward propagations for the warm case (Fig. 4.13e). Other characteristic is that the synoptic eastward during the cold case is located over continent, and during warm years it forms over ocean and propagates into the continent. Although rainfall propagation exists into the continent during warm years, the rainfall is not as intense as over ocean. This must behave in that way because there is a suppressing interannual mechanism illustrated using the east-west circulation (Fig. 3.4a). The eastern-propagating distance is longer in the cold case than in the warm case. Also, during the warm and cold cases these synoptic perturbations match very
well with the maximum phase of the 12-24 intraseasonal mode. The rainfall during cold years is more enhanced than the rainfall during warm years, which follows the enhance mechanism during cold cases. The separation period for one perturbation and the next is about 5 days in the warm case, and about 2-3 days in the cold case, which match with it was found in the power spectral. Thus, more eastward synoptic propagation is expected during cold years. During warm years the eastward propagation can reach 45°W, but in many cases it stops at 60°W. However, in cold cases, it can pass 40°W and continue propagating to tropical Africa. These characteristics may be the synoptic distinction between cold and warm years.

4.3.3.2 Modulation of the diurnal cycle

High time resolution data was employed to investigate the effect of the diurnal cycle on the rainfall during warm and cold ENSO phases. A Hovmöller diagram of 3-hour resolution for years 1998 and 2000 are shown in Fig. 4.13f and 4.14f respectively. The most notable feature is the westward propagating disturbances due to the role played by the diurnal cycle. The characteristics and effects of these westward propagating disturbances are the following.

During the cold year westward disturbances form everywhere over the continent and they couple with the synoptic eastward disturbance in order to enhance it. It can be observed that cases 5, 6, 7, and 8 (Fig. 4.14f) are enhanced because the positive interaction between the diurnal westward propagation and the synoptic eastward disturbance. Also, the presence of intense diurnal westward propagation coincides with the maximum (wet) phase of the 12-24 filtered time series. During the warm year the diurnal westward propagation is found
mainly over eastern continent and not well coupled with the synoptic eastward propagation. Although the amount of rainfall produced is less than the cold case, it can propagate farther west than in cold year. In several cases, the westward perturbation reaches the Pacific Ocean and generates big storms in the area of big convection which keep moving westward. Thus, small diurnal perturbations generated over eastern tropical South America can produce big storms over western Pacific.

Other characteristic of the diurnal westward perturbation is that during cold years it can link two consecutives eastward perturbations producing a special enhancement stage. For example cases 5 and 6 (in Fig. 4.14f) are seen as a single eastward propagation case when synoptic and diurnal variation is removed (Fig. 4.14d). However, according to Fig. 4.14f storm 6 is linked with 5 through the diurnal westward propagation. In other words, this is another mechanism responsible for the enhanced precipitation during cold years. During warm cases, a connection exists between two eastward propagations, but not in the way observed in cold cases.

Moreover, there is no signal of westward propagation over ocean in cold cases, but in warm years a clear signal of westward propagation exists over ocean. This characteristic not only enhances the eastward propagation over ocean, but also it can produce strong westward disturbances as was commented in the previous paragraph. Therefore, it can be concluded that the rainfall regime over tropical South America is generated by the way the synoptic scale is coupled with the diurnal scale under the influence of the intraseasonal modes governed by the ENSO.
Figure 4.13  Seasonal, (FMA), a) and departure b) of precipitation for year 1996.

c) The x-t diagram of 5-day mean precipitation average over 5°S-0°, and below the histogram of $\Delta[P]$. d) The x-t diagram of the 5-day mean $P$ superimposed with the 30-60 band-pass filtered $P$ time series. e) The x-t daily $P$, and superimposed with the 12-24 and 30-60 band-pass filtered $P$ time series. f) The x-t diagram for 3-hour mean $P$. Superimposed black dots are eastward disturbances, and red dots are westward representing disturbances.
The roles of surface diurnal heating on the diurnal cycle of precipitation, and its implication on the Asian monsoon system were explained by Chen (2006). He found a clockwise 'oscillation' of the diurnal rainfall associated with the eastward rotation of the Earth. The consequence of this fact is a westward retarding of heating which produces a westward propagation of the diurnal rainfall. If the reverse rotational direction property is applied to our case in the Southern Hemisphere, we have a counter-clockwise rotation, but the surface heating still maintains a westward direction, so a westward propagation is expected. Our result (Figs. 4.13f and 4.14f) shows clearly this westward diurnal propagation and those might be due to surface diurnal heating. However, other local factors contribute to modulate this diurnal behavior. First, the continental land-sea breeze located at $50^\circ$W can be noted as a recurrent forcing over this area. It can be observed that for cases during Feb 20 to Mar 1 of 1998 the rainfall over eastern continent is not suppressed, although it is in the central continent. Second, it is clearly observed that the diurnal cycle modulates the synoptic scale, and both are modulated by the intraseasonal modes.

In conclusion, all those processes act in different scales in time and space, and they interact to constitute the rainfall pattern portrayed for tropical South America in this discussion. In addition, the multiple scale processes for the cold and warm cases are similar, as it was discussed; however, some features can be distinguished that make cold cases unite and opposite to warm cases. First, there is no rainfall propagation over the ocean in cold cases, but an oscillation pattern of the rainfall exists due to the intraseasonal mode, which was verified using the potential function ($\chi$) in upper and low levels. Second, inside the maximum phase of the low frequency activity, the number of high frequency activity is larger in cold than in warm cases, and this role is played by the large scale divergent
circulation. During warm years a convergence was observed in low levels and divergence in high levels which drives the circulation in a 30-60 window perspective, but still there is suppression due to the interannual variation of the large scale circulation; that is the reason why the rainfall is weak in warm years. Thus, the synoptic scale is modulated not only by the intraseasonal modes, but also by the large scale interannual circulation influenced by the ENSO. Third, the diurnal variation modulates the synoptic scale and links the coexisting cases.
4.4 Synoptic disturbances responsible for the rainfall pattern

Synoptic perturbation varies with time, and those variations may depend on the influence of large scale processes and low frequency phenomena. It was discussed in previous sections how the interannual and intraseasonal modes affect the rainfall pattern in tropical South America. However, the discussion has not addressed the impact on the synoptic representation of the weather systems during the cases exposed. In this section, typical weather systems are portrayed in order to illustrate the synoptic dynamic involved in these complex processes at multiple time scale, and to provide evidence about how the weather system responds under the influence of large-scale/low-frequency modulations. The discussion is divided in two branches: the analysis of warm cases and the analysis of cold cases, and they are separated into eastward and westward categories according to their disturbance propagation.

4.4.1 Warm cases

The power spectral analysis of 3-hour precipitation has shown the existence of high frequency disturbances (Fig. 4.12), which can be observed in years 1998, 2003, and 2007. Year 1998 is chosen as the representative strong warm case. The amplitude of its diurnal cycle is larger than other high frequency modes, so there is a significant high frequency signal. As it was concluded in the previous chapter, showing the coupling between high and low frequency modes is a way to explain the interaction of processes at different scales. Also, it was pointed out that during 1998 a predominant westward propagation existed, and the coupling between the eastward and westward propagation is weak over the continent, so a dry season is observed in order to ensure the suppression by the east-west Walker circulation.
Therefore, in this section synoptic perturbations associated with these restrictions are analyzed in order to portray how the synoptic system behaves under the influence of the warm ENSO phase.

### 4.4.1.1 Westward synoptic propagating systems

Westward rainfall propagation systems are the most important type of disturbance over the Amazon basin during the warm ENSO. This was noted by several authors (e.g., Cohen et al. 1995, and Yamazaki and Rao 1977). Although the synoptic representation of those disturbances is not clearly noted over the continent, they can be observed over eastern Pacific when rainfall propagation is presented during warm years. For example, Figs. 4.18 and 4.19 (left panels) show a case of westward propagation during warm ENSO years. There the propagation starts over the continent, 75°W, and propagate through eastern Pacific. The period of this propagation is approximately 1 day, which support the concept about the diurnal cycle forcing the westward propagating systems. Also, the first time step (00GMT 2/26/98), and the last time step (00Z 2/27/98) have a similar configuration, which can be noted in the rainfall at the mouth of the Amazon river (eastern continent). It seems that convergence at low level is the principal triggering mechanism, as can be noted by the easterly wind at 850 mb. At 00GMT 26 February 1998 the trade winds blows straight up the east slope of the Andes Mountains, then the disturbance cross the Andes and reaches the ocean, and two anti-cyclonic vortex emerge one in the Southern Hemisphere and other in the Northern Hemisphere, which creates a convergence zone to maintain the transport of water vapor to feed the convective storm. Also it is observed that the double anti-cyclonic vortices move westward with the precipitation since 12GMT 2/26/98 to 00GMT 2/27/98.
In lower levels the perturbation is difficult to detect as can be observed in the weather maps shows in Fig. 4.23 for this case. However, in upper level (Fig. 4.19) the signal is more clearly observed than in lower levels. The Bolivian High in the Southern Hemisphere and an anti-cyclonic vortex located over Panama in the Northern Hemisphere are coupled and form a diffluence zone which propagates along the precipitation disturbance. The dots indicate the location of the diffluence zone at every time step. Therefore a typical synoptic chart representation of the westward propagation exists, and its horizontal representation suggest that convergence in low levels and diffluence in upper level are the main forcing to maintain the system moving. Thus, the westward disturbance propagation is important because it generates most of the rainfall during strong warm cases.

The westward disturbance was noted to be associated to the diurnal cycle and was very recurrent during the warm ENSO phase. This can be noted at times 00GMT 2/26/98 and 00GMT 2/27/98. At 850 mb, the rainfall patterns in those two plots are similar especially in the eastern continent and in the coastal Pacific Ocean. Noting how the storm at the eastern continent is followed by an amplified through, and an anticyclonic vortex by the storm in the Pacific coast. Thus, these two plots might show the important role of the diurnal cycle.

The most peculiar signature during the warm case is the lack of rainfall inside the continent between 70°W to 50°W (Fig. 4.18, red box in the left panel), and the opposite happens during the cold case (Fig. 4.21, red box in the left panel). It can be observed that the rainfall pattern in those plots are opposite. The reason of this configuration can be observed at upper levels using 6-hour resolution data sets. For the cold case (Fig. 4.22, right panel), the rainfall propagating perturbation is associated with a diffluence zone generated by two anticyclonic systems which propagate along the rainfall perturbation, and it is located in the
middle of the continent. For the warm case (Fig. 4.19, right panel) two diffluence zones are observed. One in the western continent which propagate along the rainfall pattern, and other at eastern continent which keep stationary. In this case, the middle continent is dry because it is the area of subsidence. This matches very well with the representation of the Walker Circulation over tropical South America. At lower levels, we found that this dry period over continent occur when the trade wind has a NEN direction as it is shown at time step 12GMT 2/26/98 or 12GMT 2/27/98. This synoptic representation not only occurs during warm years, but also during cold year between the dry periods (in the minimum phase of the 12-24/30-60 intraseasonal modes). In another case not shown here, it was observed that the NEN wind can come from the Northern Hemisphere (~20ºN), and be controlled by the North Atlantic subtropical anticyclone. On the contrary, during the cold ENSO phase (Fig. 4.22), the trade wind is easterly. Therefore, we speculate that during the dry period relative cold air flow into the continent and this $\Delta T$ reduces the buoyancy condition, which can be consider another suppression factor.

4.4.1.2 Eastward synoptic propagating systems

It was identified that the eastward propagating perturbation is the most efficient mechanism for rainfall-production over tropical South America. First, because its propagation velocity is slow compared to the westward case, and second because of its coupling with the westward propagating perturbation. Those characteristics make this type of perturbation the best way to produce considerable rainfall and to release important amount of latent heat during the water vapor transformation. However, during the warm ENSO phase the rainfall over the continent is suppressed, as it is observed in the in the synoptic chart (Fig.
4.24, left panel) for a case of eastward propagation during the warm ENSO 2010, so the weather system has to adjust to a dry environment. Thus, the eastward synoptic disturbance during the warm ENSO phase behaves in different ways in order to fulfill the suppression mechanism.

The main feature of the eastward propagation is that precipitation is reduced in amount and extension when it reaches the continent. Fig. 4.15 shows a 5-day mean streamline chart at 850 and 200 mb, which are superimposed with precipitation. The eastward propagating precipitation is observed and highlighted with dots and joined by dashed lines. This is the same case 1 showed in the x-t precipitation diagram (Fig. 4.13d). At time 00GMT 2/25/98 precipitation reaches the coastal area, and it appears very intense. Simultaneously, another area of precipitation is enhanced around 50°W. These two areas of cumulus convection characterize this case, but can these two areas be linked in some way? Yes, they can. We found that these systems are linked in upper levels. In upper levels two diffluence zones are formed over the enhanced storms, one at 80°W and other at 50°W. Each diffluence zone is maintained by two pair of vortices. Two located in the Southern Hemisphere and one in the Northern Hemisphere. In low levels, an anticyclonic vortex at 80°W near the equator in the Southern Hemisphere might promote the convergence. In the center of precipitation (over the continent), the anti-cyclonic vortex is intensified comparing with the time step before and after. This intensification generates a trough east of the anti-cyclone, in a way that the precipitation resides ahead of the trough. Therefore, the lower and upper levels are synoptically coupled in order to have an area of favorable environment for deep cumulus convection forming with the flow converging at low levels and diverging at upper levels.
Therefore, during warm years the synoptic eastward propagation is very well defined in the streamline. The low-level convergence driven by cyclone/anticyclone vortex seems to be the main mechanical forcing to produce convection (Fig. 4.15), and a diffluence zone in upper levels is observed along the propagation. Thus, convergence in low levels and divergence in upper levels is the main synoptic characteristics which can be verified using $\chi$ (Fig. 4.16).

**The maintenance**

A better illustration of this lower and upper coupling can be portrayed using the potential function ($\chi$). Fig. 4.16 shows $\chi$ for lower and upper levels for the same case 1 in the previous section. It is clearly noted how flow converges in low levels and diverges in upper level along the rainfall propagation. Also the eastward propagating characteristic is observed. To complement the $\chi$ representation, the stream function ($\psi$) is shown in Fig. 4.17 for the same case. These plots show the quadrature relationship between $\chi$ and $\psi$ in both levels 850 and 200 mb. This spatial quadrature relationship means that the dynamics governing this process is ruled by the Sverdrup approach.

**The blocking propagation**

The rainfall over the continent was shown to be suppressed according to the interannual variation of the season. In the previous section we showed that the rainfall is suppressed in order to maintain the seesaw balance between continent and eastern Pacific. As it was shown in the x-t diagram of equatorial precipitation (Fig. 4.13) the rainfall propagation is blocked in some cases. Thus, what can be the synoptic response to this blocking
configuration?

The blocking state is defined when the perturbation does not continue moving and stops in the western continent. The precipitation disturbance formed in the central Pacific moves eastward helped by a cyclonic/anticyclonic vortex system in most of the cases. However, when the perturbation reaches the coast, it is not possible to develop convergence at low levels on the east Andes Mountains, and the eastward propagation is interrupted. It was pointed out that westward disturbances over the continent are very common during extreme warm ENSO years, and also those westward disturbances are clearly observed over the ocean. Those facts may have some relation of why it does not rain so much over continent. Therefore, we argue that blocking propagation might play two roles: 1) it contributes to enhance precipitation over the ocean via coupling between the westward and eastward precipitation, and 2) it maintains the subsidence over the continent avoiding the triggering of cumulus convection via the blocking of eastward weather disturbances. This blocking type of weather disturbance is usually observed during extreme warm years like 1983 and 1998, and it may be the reason why these years are extremely dry. In addition, a semi-permanent anticyclonic system is observed over the continent and it can help to transport moist air from the Amazon Basin to eastern Pacific as was pointed out by Horel and Cornejo-Garrido (1986).

Fig. 4.15 shows the eastward rainfall propagation using a 5-day time scale. It is observed at time step 2, (00GMT 2/25/98), that the rainfall is inhibited over the continent along the Equator; however, at the next time step, (00GMT 3/02/98), an important center of cumulus convection is located south of the equator. Therefore, it seems that the propagation take a detour southward in order to maintain the mass continuity balance. Using the daily
time scale to analyze this disturbance, Figs. 4.18 and 4.19 (left panel), it can be noted that at time 00GMT 2/27/98 cumulus convection is formed along the east slope of Central Andes. This happens in several cases, not only in warm cases but also in cold ones. In this case the forcing is also produced by the low level convergence. Thus, the trade winds curves southward around the equator flowing parallel to the Andes slope, then the flow lifts when it reaches the bottom north face of the Altiplano. When the storm is formed, it moves in the opposite direction (northward or northeastward) of the trade winds. We believe that this system is maintained by a mesofront which is formed between the trade wind and the downdraft zone of the convective (complex) system (Cotton, 1990). Although we have evidence for the time evolution of this process, our spatial resolution data set is not high enough to support the idea of this mechanism. Therefore, a way to inhibit the precipitation formation in tropical South America is moving the convective system southward which can be observed in all the time steps in Fig. 4.18 (left panel).

Another way to block the propagation is through the appearance of reflexive synoptic systems, which split the coming easterly perturbation through wave reflections. If we observe Fig. 4.15, a kind of westward propagation can be noted from 80ºW to 100ºW at times 2/25/98 and 3/02/98. This propagating perturbation is not only noted in precipitation field, but also in the streamline. This observation is confirmed using daily mean data (Fig. 4.18). There the westward propagation is clearly observed, and an anticyclonic vortex, H, is moving along the perturbation. Moreover, it seems that the perturbation starts on the western continent, and the anti-cyclonic system is an extension of the anti-cyclonic vortex over continent. This kind of propagating anti-cyclonic vortex has been observed in other cases. In addition, Fig. 4.13f can confirm that the perturbation genesis is over continent, as it is seen in case A.
The importance of these reflexive synoptic systems is that they enhance the incoming eastward propagation. In other words, the eastward propagation is merged with those reflective waves which have a different frequency as consequence of the reflection. However, these reflective waves are not the only way to enhance precipitation along the eastward propagation over the ocean, the diurnal mode also plays a similar role. This enhancement has the same principle as what occurs over the continent during cold cases, where the diurnal cycle modulates the eastward perturbation enhancing the rainfall, but the advantage during warm years is that the synoptic perturbations are observed more clearly over the ocean than over the continent.

4.4.2 Cold cases

The synoptic perturbation during cold cases is better observed over the continental area, and there is no perturbation coming from the central Pacific as in warm cases. This happens because the subsidence flow over the eastern Pacific inhibits all potential formation of cumulus convection. However, the inhibition state over the eastern Pacific is not an inert state because its main role is to provide energy as latent heat to the tropical convection in the western Pacific to maintain the east-west Walker circulation as it was pointed out by Cornejo-Garrido and Stone (1977). Also, during the cold ENSO phase, two kinds of synoptic perturbation (eastward and westward) can be observed over tropical South America, and both are responsible for the rainfall formation.

According with the interannual scale analysis, rainfall is suppressed over the eastern Pacific and enhanced over the continent. However, an important storm is observed propagating westward since 17 March 2000 to 01 April 2000 (Fig. 4.20), also a cyclonic
system (L) is clearly observed. Can this case contradict the theory about the rainfall suppression over eastern Pacific during cold cases? No, it cannot, because this is the only case observed in the whole season as can be noted in Fig. 4.14c. However, others weak-westward propagating cases have been observed in other cold and neutral ENSO years. This is the reason why the precipitation departure for 2000 (Fig. 4.14b) shows a positive anomaly between 120°W and 80°W, where the case for 2000 of westward propagation is observed. This happens during late March most of the time, and it is also noted in the warm year 1998 (Fig. 4.13). In warm years the origin of these perturbations was detected in the eastern continent by the effect of the diurnal cycle forced by the continental land sea-breeze, but for cold years we can not confirm that the origin is in continent (as it seems in Fig. 4.14c), because there is a gap area (no rain), between 80°W and 90°W due to the cold water of the coastal ocean current of Peru, which makes it impossible to track the signal. The situation is reversed in warm years and the continuity of the westward propagation can be observed.

4.4.2.1 Westward synoptic propagating systems

The diurnal cycle has important amplitude, and signals in the order of [1.5-6] day are distinguished (Fig. 4.12). For instance, in the year 2000, two peaks are very clearly identified at day 2.5 and 1.5, and another with less amplitude around day 4. The former can be considered the high frequency modes, where the diurnal cycle can be included, and the latter can be considered as the synoptic disturbance. It was shown using the Hovmöller diagram that the synoptic and the high frequency modes interact to enhance the rainfall pattern during cold years in the active phase of the interannual modes (12-24 and 30-60). This fact reflects
the importance of the coupling of the westward propagation (represented in these high frequency modes) with the eastward perturbation (synoptic mode). However, what is the horizontal structure of those weather disturbances under the influence of the interannual and intraseasonal modes?

Westward propagation over continent also exists during cold years, and it seems to be ruled by the diurnal cycle, but it can live longer than 24 hours. Figs. 4.21 and 4.22 (left panels) show a storm case that propagates from the central continent to the western continent. The trade winds flow parallel to the equatorial line, and the perturbation follows that direction. Wave-like perturbations are not observed in the stream lines, neither convergent flow is evident in this case. The main forcing should be in upper levels. In upper levels (Fig. 4.22), a diffluence area is observed located over the main precipitation area. When the divergence field is computed, it shows that the diffluence area is a divergent one.

The interaction between the synoptic and high-frequency scale can be better noted in upper levels (Fig. 4.22). It can be observed that the big amount of rainfall on 3/25/00 (daily scale) is consequence of the superposition of the eastward (daily scale) and the westward (6-hour scale) perturbation. In the 6-hour scale, a diffluence perturbation is moving along the rainfall propagation. At the same time a diffluence zone move eastward along the rainfall perturbation in the 1-day scale. Therefore, the westward perturbation is embedded inside the synoptic scale and it might be controlled by the diurnal cycle. As in the opposite situation of the warm cases, the coupling occurs in the middle of the continent.

4.4.2.2 Eastward synoptic propagating systems

Eastward disturbances are the most efficient way to produce rainfall over tropical
South America during the cold ENSO phase. Thus, during warm years, few cases of eastward perturbation over the continent are observed in order to maintain a rainfall suppression state. However, during cold years the situation is inverted, and more cases are observed. Also, when the year is extremely wet the number of eastward disturbances increases compared to neutral or less extreme years. Therefore, the eastward perturbation is an important one to ensure maximum release of water vapor over continent, and maintain the upward branch of the Walker circulation during the cold ENSO phase.

The genesis of eastward synoptic disturbances appears over the western Amazon basin in the Colombian and Peruvian Amazon, and they are formed when low level convergence is coupled with a diffluence system in upper levels as in Fig. 4.21. Then, they propagate until reaching the eastern continent in a time period of 4-6 days. Fig. 4.24 (left panel) shows a synoptic chart in which is observed a low system (B) moving eastward. It was pointed out by Wang and Fu (2007) that in some cases the propagation can reach the African continent, and the activity over the western Amazon can modulate the rainfall pattern over the Atlantic ITCZ through this propagation mechanism (Nguyen and Duvel 2008).

The eastward propagation can be observed in the 5-day mean stream lines (Fig. 4.20), as it is marked by dots and lines. The perturbation is a trigger on the east Andes Mountains and it seems to be forced in low levels by an anticyclonic system, which keeps the trade winds easterly and straight to the Andes Mountains. In upper levels, a diffluence zone is observed at the beginning of the propagation, and other diffluence zone at the end. Although we can think that there is not connection between these two diffluence zones, Fig. 4.21 proves that these are the same, and part of the propagating system.

The interaction between lower and upper level is better portrayed using $\chi$. As in the
warm case, it shows convergence in low levels, and divergent flow in upper levels. This result proofs the existence of favorable conditions for deep cumulus convection formation. The existence of simultaneous propagation makes it difficult to have a isolated signal, but it is solved using different time steps. The interaction between the eastward and westward propagation can be observed at upper levels in Fig. 4.22. In the left panel, the diurnal time step shows the synoptic eastward propagation. There a diffluence weather configuration is observed following the perturbation. In the right panel, 6-hour time step variation portrays the westward perturbation. Also, a diffluence configuration is observed following the rainfall perturbation, which seems to be governed by the diurnal mode. Therefore, the rainfall observed in the red box, which occurs in the middle of the continent. It is consequence of the diurnal (westward) perturbation, which is embedded in the synoptic system.

In conclusion, the synoptic scale and the high-frequency scale interaction were portrayed using horizontal maps in this section. Many of the examples were used to illustrate the findings described early in this chapter. The main characteristic of the synoptic system during the warm ENSO is the blocking propagation, and the synoptic system during the cold ENSO is its constraint to continental area. Thus, differences of synoptic systems were found during warm and cold ENSO years. However, some basics mechanism exists in both ENSO phases. During both warm and cold cases there is a strong coupling between low and upper levels. The frequency of occurrence and the intensity establish the particularity of each ENSO phase. Also, the enhanced rain over the ocean during warm ENSO years seems to have a similar mechanism to the processes over continent where the synoptic signal is not clear observed.
Figure 4.15  Left panel: The 5-day mean streamlines at 850 mb for a case of moving perturbation beginning on 20/Feb/98 superimposed with precipitation for the same period. The red dots represent the moving disturbance. Right panel: same as left panel except for streamlines at 200 mb. Letters H represent anticyclonic systems.
Figure 4.16  Left panel: The 5-day mean potential function and divergent wind at 850 mb for a case of moving perturbation beginning on 20/Feb/98 superimposed with precipitation for the same period. The red dots represent the moving disturbance. Right panel: same as left panel except for streamlines at 200 mb.
Figure 4.17  Left panel: The 5-day mean stream function at 850 mb for a case of moving perturbation beginning on 20/Feb/98 superimposed with precipitation for the same period. The red dots represent the moving disturbance. Right panel: same as left panel except for streamlines at 200 mb.
Figure 4.18  Left panel: The 1-day mean streamlines at 850 mb for a case of moving perturbation beginning on 18/Feb/98 superimposed with precipitation for the same period. Right panel: The 6-hour mean streamlines at 850 mb for a case of moving perturbation beginning at 00GMT 26/Feb/98 superimposed with precipitation for the same period. The red dots represent the moving disturbances, and letters H represent anticyclonic systems.
Figure 4.19  Same as figure 4.18 except for 200 mb.
Figure 4.20  Left panel: The 5-day mean streamlines at 850 mb for a case of moving perturbation beginning on 17/Mar/00 superimposed with precipitation for the same period. The red dots represent the moving disturbance. Right panel: same as left panel except for streamlines at 200 mb. Letters L represent cyclonic systems.
Figure 4.21  Left panel: The 1-day mean streamlines at 850 mb for a case of moving perturbation beginning on 21/Mar/00 superimposed with precipitation for the same period. Right panel: The 6-hour mean streamlines at 850 mb for a case of moving perturbation beginning at 0000 UTC 24/Mar/00 superimposed with precipitation for the same period. The red dots represent the moving disturbances.
Figure 4.22  Same as figure 4.21 except for 200 mb.
Figure 4.23 Left panel: Surface weather charts for the case of moving perturbation beginning on 18/Feb/1998. Right panel: Surface weather charts for the case of moving perturbation beginning on 21/Mar/00. The red dots represent the moving disturbance.
Figure 4.24  Left panel: Surface weather charts for a case of moving perturbation beginning on 02/Jan/2009 (considered a Normal ENSO year). Right panel: Surface weather charts for a case of moving perturbation beginning on 14/Feb/2010 (Considered a Warm ENSO year). The red Bs represent the moving low pressure disturbance.
CHAPTER 5. GENERAL CONCLUSIONS

The rainfall regime over tropical South America was studied using a downscaling approach, so a multiple time scale technique were used to isolate and analyze the contribution and role of several modes of 'oscillation'. In this investigation, the intraseasonal modes (12-24 and 30-60), the high synoptic modes (2-8 days), and the diurnal mode were found responsible for the rainfall pattern observed during the season FMA. Also, these modes interact in a particular way ruled under the influence of the interannual regulation of the ENSO.

It was found that the interaction of synoptic disturbances and diurnal disturbances are responsible for the rainfall regime over tropical South America. The synoptic disturbances propagate eastward and the diurnal disturbances propagate westward. The ways they interact are ruled by the ENSO as a first order of approximation. Thus, during the warm ENSO phase the synoptic eastward propagation is suppressed and a weak interaction is observed, and a dry season dominates this period. During the cold ENSO phase, a strong interaction between those opposite propagation is observed, so a wet season is experimented. However, the ENSO is not the only agent which can play a role in the rainfall regime in tropical South America, so other modes were found to be important in this multiple scale processes rainfall-production. In the following paragraphs a brief description of the principal findings in this research are explained.
The ENSO

The role of the ENSO is to modulate the interannual variation of the rainfall in tropical South America. This modulation is made through the east-west Walker circulation, so the ascendant (descendent) branch of that circulation will rule the rainfall regime during the cold (warm) ENSO. During the cold (warm) phase, more (less) rainfall is produced. The main mechanism forcing is the east-west global circulation (Fig. 3.5). This mechanism inhibit (enhance) the coupling of synoptic disturbances during the warm (cold) ENSO. To maintain the balance a seesaw pattern is observed between the eastern part of the continent and the eastern Pacific Ocean. The former is associated with a continental land-sea breeze (Kousky 1980), and the latter with an extension of the central Pacific precipitation during warm years which may have some local effect including the oceanic upwelling and local land-sea breeze (Horel and Cornejo 1986).

The intraseasonal modes

In this research, we were able to link the interannual scale and the synoptic scale through the intraseasonal scale. Before this investigation, it was not clear the role of the 12-24 day mode on tropical South America. We found that the 12-24 day mode embedded in the 30-60 day mode controls the maximum amplitude of the rainfall during cold and warm years. In addition, during warm years, the inland penetration of eastward disturbances is associated with the 12-24 day mode.

The modulation of rainfall by the 30-60 day mode was very well portrayed over the continent. During warm years, this 30-60 day mode perturbation propagates eastward from central Pacific to South America, this quasi-cyclic perturbation is noted over the continent as
an oscillation state of dry/wet/dry periods, which are observed during warm and cold years. The dry/wet period over continent are in agreement with the potential function \(\chi\) filtered in the 30-60 day mode, so the dynamical forcing illustrated in this research follows Chen's approach (Chen et al. 1995b). The phase reversal of \(\chi\) in 850 and 200 mb support the Sverdrup dynamics approach as can be noted by the low level convergence and upper level divergence coupling.

Although, the 12-24 day mode can play a global role on the hydrological cycle (Chen et al. 1995b), we analyzed it in a regional prospective (over the Amazon basin). Thus, it was found that during warm years the penetration of the eastward (30-60) perturbation occurs when it is coupled with the 12-24 day mode. During cold years, it was found the maximum event of rainfall pattern over continent coincide with the maximum phase of the 12-24 day mode.

**The synoptic mode and the diurnal cycle**

Using a power spectral analysis several peaks were found in the range of high frequency, and the diurnal cycle was identified in the spectral analysis. However, warm and cold years show different characteristics. During the warm year it was found a suppression of the synoptic perturbation, which was verified with the Hovmöller diagram. This condition makes easy to observe a more recurrent synoptic westward perturbation forcing by the diurnal cycle and maintained by the diurnal diabatic heating. Thus, the mechanism to suppress rain over continent during warm years is via the inhibition of the synoptic disturbance. This concept is also supported by the power spectral plot where the big amplitude of the diurnal cycle for 1998 is observed.
Moreover, during the cold phase of the ENSO, all the perturbation mode act together to enhance the rainfall over tropical South America. Thus, the westward perturbation modulated by the diurnal cycle coupled with eastward synoptic perturbation which is embedded in a regional 12-24 day mode. In addition, the eastward synoptic perturbation can co-exist as double line propagation as a way of maximum enhance coherence which occurs specially during strong cold events. This works in a special configuration where maximum water vapor is released and maximum water resource over the continent is obtained.

**The weather system**

All the early finding was observed in synoptic charts. Thus, we found the following characteristic properties. First, during warm years a special synoptic configuration which blocks the eastward propagation is observed. In this blocking case, a semi-permanent anticyclone system is observed in central continent, which pushes the water vapor southern and trigger big storms in the northern face of the Altiplano. Thus, the eastward perturbation moves southern from the Equator. This behavior, during the warm ENSO phase, sustains the rainfall suppression over tropical continent.

Second, during cold years not perturbation are observed over eastern Pacific, and rainfall is enhanced over continent because of the coupling between synoptic eastward and the diurnal westward propagations. The formation of the westward propagation is due to low level convergence between continental land-sea breeze and the trade winds (Kousky 1980), and the formation of the eastward perturbation is due to low level convergence in the east side of the Andes Mountains. Several mechanism were found responsible for this low level convergence, and the most common is an anticyclonic vortex located at western continent.
and a cyclonic system below the previous one. This type of dipolar High/Low system can intensity further on time and produce a easterly flow which converge with the trade in central continent; this is a similar generation pattern compared with the westward perturbation.

Finally, we believe that a complete understanding of the process responsible for the rainfall regime over tropical South America is crucial. In this research some of them were identified, isolated, and analyzed their interaction. However, future work should be focus on the understanding of how this concept can be inserted in a mathematical/computational model in order to use it for forecasting and numerical diagnostic purpose. A brief analysis of precipitation data produced by the model of the Hydro-climatology group of Princeton University shows up not existence of these high frequency mode propagation. Thus, the miss-representation of those processes will induce arbitraries forcing as a consequence of the uncoupled system between upper and lower level through deep cumulus convection, and this spurious forcing will works against the predictability of the weather and climate system.
REFERENCES


Halley, E., 1686: An Historical Account of the Trade Winds, and Monsoon, observable in the Sea between and near the Tropicks, with an attempt to assign the Physical cause of the same Winds, *Phil. Trans.*, **16**, 153-168.


Ropelewski, C., and M. Halpert, 1989: Precipitation Patterns Associated with the High Index


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