Coupling between the Andes and the Amazon River Basin and their feedbacks with the Tropical North Atlantic and Tropical Pacific Ocean

Luis Alejandro Builes Jaramillo

Universidad Nacional de Colombia
Facultad de Minas, Department of Geosciences and Environment
Medellín, Colombia
2017
Coupling between the Andes and the Amazon River Basin and their feedbacks with the Tropical North Atlantic and Tropical Pacific Ocean

Luis Alejandro Builes Jaramillo

A thesis submitted in partial fulfillment of the requirements for the degree of: Doctor of Engineering in Water Resources

Director:
(Ph.D) Germán Poveda Jaramillo

Research Track: Hydroclimatology
Research Group: Posgrado en Aprovechamiento de Recursos Hidráulicos (PARH)

Universidad Nacional de Colombia
Facultad de Minas, Department of Geosciences and Environment
Medellín, Colombia
2017
“It is not the strongest of the species that survives, not the most intelligent that survives. It is the one that is the most adaptable to change.”

Charles Darwin
Acknowledgments

First I would like to thank my advisor Prof. Germán Poveda for saying yes when I first came to his office seeking for advice. He helped me to maintain my curiosity and interest regardless of the circumstances and he let me propose and try new things that I really enjoyed during this learning process.

I would like to express my gratitude to Prof. Jürgen Kurths and Dr. Norbert Marwan from the Potsdam Institute for Climate Impact Research at Potsdam, Dr. Antônio Ramos from the National Institute for Space Research at Sao Jose dos Campos and Prof. Paola Arias from the Universidad de Antioquia at Medellin for their valuable and insightful reviews and comments during my Ph.D studies.

I would like to thank also all the Institutions that supported my research: Institución Universitaria Colegio Mayor de Antioquia, Universidad Nacional de Colombia, German Academic Exchange Service, Potsdam Institute for Climate Impact Research and Humboldt University of Berlin. Also I am thankful with my fellow graduate students and friends Alejandra, Edna and Johanna for their support and comprenhension.

I have a heartfelt thanks to my parents and my brother for being there, for their care and loving support, to Nati, Juan Camilo, Yully, Rubén, all my friends and Kiwi for putting my thoughts out of the academy.

Last but not least, I would like to acknowledge those that for whatever the reason did not endure till the end of this journey, but that had a special place in it, Laura and Lola.
Abstract

In this thesis we study the hydroclimatic interrelations between the Andes and the Amazon River basin with the Tropical North Atlantic and the Tropical Pacific. Particularly, we estimate the separate and conjoint surface and atmospheric water budgets in the entire Amazon River basin, its major subcatchments, as well as in the Andean and low-lying portions of the Amazon River basin. The water balances approach provided an update regarding the spatial patterns of the water budgets closure in the basin, and a quantification of the two-way interactions and coupling existing between the Andean and low-lying regions of Amazonia. We studied the complex two-way feedback dynamics between the Amazon River basin and the Tropical North Atlantic with linear and non-linear methodologies, and the analysis showed that at seasonal and interannual timescales the Amazon River basin plays an active role modulating the Tropical North Atlantic Sea Surface Temperatures. Finally, with the use of a new non-linear methodology we provided new insights of the function that the Amazon River basin plays in the connection between the Pacific Ocean and the Tropical North Atlantic during an El Niño event as "land-atmosphere" bridge.

Keywords: Amazon River basin, Water balances, Hydroclimatology, Tropical North Atlantic, Tropical Pacific, Recurrence

Resumen

En esta tesis se han estudiado las interrelaciones hidroclimáticas entre la cordillera de los Andes y la cuenca del Río Amazonas con el Atlántico Tropical Norte y el Pacífico Tropical. En particular, hemos estimado de forma independiente y conjunta los balances de agua superficial y atmosférico en toda la cuenca del Río Amazonas, sus principales subcuentas, y en las regiones Andina y de tierras bajas de la cuenca del río Amazonas. El análisis de balances de agua proporcionó una actualización sobre los patrones espaciales del cierre de los balances hídricos en la cuenca y una cuantificación de las interacciones de doble vía y el acoplamiento existente entre la región Andina y de tierras bajas de la Amazonía. La dinámica compleja de retroalimentación bidireccional entre la cuenca del Río Amazonas y el Atlántico Tropical Norte fue estudiada con metodologías lineales y no lineales, y este análisis mostró que en las escalas de tiempo estacional e interanual la cuenca del Río Amazonas desempeña un papel activo en la modulación de las temperaturas superficiales del océano Atlántico Tropical Norte. Finalmente, con el uso de una nueva metodología no lineal se encontraron nuevas evidencias de la función que la cuenca del Río Amazonas cumple como puente "suelo-atmósfera" en la conexión entre el Océano Pacífico Tropical y el Atlántico Tropical Norte durante un evento El Niño.

Palabras clave: Amazonia, Balance Hidrológico, Hidroclimatología, Atlántico Tropical Norte, Pacífico Tropical, Recurrencia
# Table of Contents

**Acknowledgments** ............................................................................................................. VII

**Abstract** ............................................................................................................................. IX

**Introduction** .......................................................................................................................... 1

1. **Conjoint Analysis of the Surface and Atmospheric Water Balances in the Andes-Amazon System** ................................................................. 5
   1.1 Introduction .................................................................................................................. 5
   1.2 Methods and Data Sets ............................................................................................. 8
      1.2.1 Surface and Atmosphere Water Balances ......................................................... 8
      1.2.2 Data ................................................................................................................... 10
   1.3 Results and discussion ............................................................................................... 11
      1.3.1 Closure of water balances ................................................................................. 11
      1.3.1.1 Closure of the surface water balance ......................................................... 11
      1.3.1.2 Closure of the atmospheric water balance .................................................. 15
      1.3.1.3 Closure of the conjoint surface-atmospheric water balances .. 18
      1.3.2 Annual Cycle of the long-term water balances .............................................. 21
      1.3.3 Long term imbalances between surface-atmospheric water balances ........ 25
      1.3.4 The Andes-Amazonia System......................................................................... 28
   1.4 Concluding remarks .................................................................................................... 29

2. **Seasonal and Interannual Two-Way Feedbacks between Tropical North Atlantic SST and Amazonian Hydroclimatology** .......................... 34
   2.1 Introduction ............................................................................................................... 34
   2.2 Data and Methods .................................................................................................... 36
   2.3 Results and Discussion ............................................................................................. 38
      2.3.1 Feedbacks at seasonal time scales ................................................................. 38
      2.3.2 Feedbacks at the interannual time scale ....................................................... 41
      2.3.3 Spatial extent of the Amazon influence on TNA ........................................... 42
      2.3.4 Physical evidences of feedbacks .................................................................... 44
      2.3.4.1 Physical mechanisms at play at seasonal timescales ................................. 45
      2.3.4.2 Physical mechanisms at play at interannual timescales ............................. 48
   2.4 Concluding remarks .................................................................................................. 52

3. **Nonlinear interactions between the Amazon River basin and the Tropical North Atlantic at interannual timescales** ................................ 53
   3.1 Introduction ............................................................................................................... 54
   3.2 The proposed mechanism and data sets ................................................................. 56
   3.3 Methods .................................................................................................................... 58
      3.3.1 Recurrence Analysis ...................................................................................... 58
      3.3.2 Significance test ............................................................................................. 61
   3.4 Results and discussion ............................................................................................. 62
   3.5 Concluding remarks ................................................................................................ 74
4. Atmosphere-land Bridge between the Pacific and Tropical North Atlantic SST’s through the Amazon River basin during the 2005 and 2010 droughts

4.1 Introduction

4.2 Methods and Datasets

4.2.1 Recurrence Measure of Conditional Dependence

4.2.2 Significance testing

4.2.3 Data

4.3 Results and discussion

4.3.1 Information Transfer from the Pacific Ocean to the Tropical North Atlantic through the Amazon River basin

4.3.2 Information Transfer from the Tropical North Atlantic to the Amazon River basin

4.3.3 Information Transfer from the Tropical Pacific to the Tropical North Atlantic without mediation of the Amazon River basin

4.3.4 Information Transfer from the Tropical Pacific to Major Amazon River Sub-Basins

4.3.5 Information Transfer from the Tropical Pacific to the Andes and Amazonia regions

4.4 Summary and Conclusions

5. General conclusions and future work

A. Appendix A: Chapter 1

B. Appendix B: Chapter 2

C. Appendix C: Chapter 3

References
Introduction

The hydro-climatological importance of the Amazon River basin at global and continental scales cannot be overstated, given its size of more than six million km$^2$, and its equatorial setting with a large portion covered by rainforests. All those characteristics contribute to explain the more than 200,000 m$^3$ s$^{-1}$ of mean annual discharge at the outlet [Dickinson, 1989; Nobre and Borma, 2009; Medvigy et al., 2011; Hartmann, 2015; Swann et al., 2015]. The Amazon River basin plays a fundamental hydro-climatological role on Earth’s climate, and exhibits different stability characteristics associated with land cover types and extent [Lenton et al., 2008; Keller et al., 2009; Nobre and Borma, 2009; Hirota et al., 2011].

Changes in Tropical North Atlantic (TNA) Sea Surface Temperatures (SST) are crucial for the availability of moisture in the Amazon River basin, since the dynamics of the Intertropical Convergence Zone (ITZC) follows the annual cycle of SST over the TNA, which is associated with well-known precipitation patterns over land [Fu et al., 2001; Poveda et al., 2006; Nobre et al., 2009; Yoon and Zeng, 2010; Gimeno et al., 2012; Yin et al., 2014; Arias et al., 2015]. The activity of the trade winds over the TNA region is related with the South American Low-level Jet (SALLJ), the South American Monsoon System (SAMS), and with the dynamics of aerial rivers that combined configure the transport of moisture from Amazonia all the way to the eastern flank of the Andes, and reaching as far as northern Argentina [Marengo et al., 2004; Vera et al., 2006; Moraes-Arraut et al., 2011; Poveda et al., 2014]. Also, TNA SSTs can modulate the interannual variability of fires in southern Amazonia [Chen et al., 2011; Fernandes et al., 2011].

The impacts of El Niño-Southern Oscillation (ENSO) in the Amazon River basin are well documented, and are mainly related with interannual anomalies in precipitation triggered by a displacement of the Walker circulation over South America, that may lead to droughts (floods) all over the basin [Ropelewski and Halpert, 1987; Marengo and Hastenrath, 1993; Coe et al., 2002; Poveda and Salazar, 2004; Drumond and Ambrizzi, 2006; Poveda et al., 2006; Araújo Gonzalez et al., 2007; Davidson et al., 2012; Espinoza et al., 2012]. The increase of anomalous extreme events that may be induced by climate change [Marengo and Espinoza, 2016] could be related with a possible change in the Amazon ecosystems and in the equilibrium state of the Amazon from forest to savannah [Holmgren et al., 2001; Nepstad et al., 2004, 2008; Salazar et al., 2007, 2016; Nobre and Borma, 2009; Li et al., 2011; Swann et al., 2015].

The connections between the Pacific Ocean and the TNA SSTs during ENSO have been a research topic for several decades, and there is evidence of Atlantic SST anomalies change during El Niño events. Such alterations have been typically explained through a vertical stabilization of the tropical troposphere that may induce such feedbacks with a delay related with weaker trade winds and fluxes over the Atlantic due to an anomalous Walker
circulation [Saravanan and Chang, 2000; Giannini et al., 2001], and also through an “atmospheric bridge” [Ngar-Cheung and Mary, 1996; Latif, 2001; Roy and Reason, 2001]. On the other hand, Poveda and Mesa [1997] showed evidence of a connection between the Tropical Pacific and the Tropical North Atlantic occurring through a “land-atmosphere” bridge, whereby hydrological processes acting over northern South American, mainly over the Amazon River basin, play a key role in teleconnecting both oceanic regions during El Niño events.

In spite of all the studies that have been carried out, conjoint analysis of surface and atmospheric water budgets in distinctive regions within the Amazon River basin (sub-catchments, Andes and low-lying Amazon) or the interdependences between those regions have not been addressed completely by the literature. Also, the complex interrelation between the Amazon and the Tropical North Atlantic have been underestimated, taking the Amazon as a simple spectator of the climate dynamics, but not as a potential forcing of the tropical hydroclimate.

The Andean Amazonia exhibits strong physiographic, orographic, ecological, hydrological and climatic gradients, including tropical glaciers, páramos, yungas, montane forests, tropical rainforests, and savannas, whereas the low-lying Amazonia exhibits small slopes, less strong hydrological gradients and less land cover variability. In fact, the upper Andean and the low-lying regions of Amazonia constitute two coupled sub-systems, given that the Andean region imports atmospheric water by the trade winds from the low-lying Amazon forests, and in turns the former region exports water (including sediments, nutrients, and biogeochemical tracers) to the latter [Poveda et al., 2006; Poveda and Zapata, 2016]. The complex two-way dynamics between Andean and low-lying Amazonia is a relevant topic for research since it has been neglected in research initiatives like the Large Biosphere-Atmosphere experiment in Amazonia (https://daac.ornl.gov/LBA/lba.shtml).

The main objective of this dissertation is to contribute to the understanding of the Amazon River basin hydrological balance and the role that the basin may play in the tropical hydroclimatology. With such goal in mind we study the behavior of the conjoint and separate surface and atmospheric water balances in the basin, and next we explore the complex relation between the Amazon River basin hydroclimatology with the surrounding Tropical North Atlantic and Pacific Ocean. This dissertation is organized as follows: each chapter is a self-contained manuscript that has the potential to be submitted to a journal or that has been already submitted.

The first chapter Conjoint Analysis of the Surface and Atmospheric Water Balances in the Andes-Amazon System studies the surface and atmospheric water balances for the complete Amazon River basin, its major sub-catchments, and for the Andes and low-lying Amazon regions, with a special emphasis on the water (im)balances between both budgets. This chapter represents and update of the state of knowledge of water budgets over the
Amazon River basin, and provides important insights about the regional closures of the water (im)balances in the region. This chapter is already submitted to *Water Resources Research*, and some of the results were presented in the *European Geosciences Union Assembly 2017* held in Vienna, Austria, during 23-28 of April.

The second chapter is entitled *Seasonal and Interannual Two-Way Feedbacks between Tropical North Atlantic SST and Amazonian Hydroclimatology*. It examines the relationships between the Amazon River basin and the Tropical North Atlantic Sea Surface Temperatures at seasonal and interannual timescales, and proposes the dynamics of physical mechanisms connecting both regions. The main results from this chapter are the statistical and physical evidences of the influence that the Amazon River basin convection exerts over the Tropical North Atlantic SST’s from 0 to 2 months. The results from this chapter were presented in the *2015 AGU Fall Meeting*, held in San Francisco, CA, USA during 14-18 of December, and in the *2016 EGU "Leonardo Conference: From Evaporation to Precipitation"*, held at Orense, Spain during 25-27 of October.

In the third chapter, *Nonlinear interactions between the Amazon River basin and the Tropical North Atlantic at interannual timescales*, we present the research developed as a visiting researcher in the Potsdam Institute for Climate Impact Research, Germany, during the second semester of 2015. This chapter deepens in the study of the physical mechanisms connecting the most relevant hydroclimatic processes of the Amazon River basin and the Tropical North Atlantic SSTs at the interannual timescale by using a recurrence joint probability based analysis that accounts for the lagged nonlinear dependency between time series. In this chapter we present statistically significant evidence of how the proposed mechanism is triggered by several extreme precipitation events in the Amazon River basin. This chapter is already under review in the *Climate Dynamics* Journal.

The fourth chapter is entitled *Atmosphere-Land Bridge between the Pacific and Tropical North Atlantic SST’s through the Amazon River basin during the 2005 and 2010 droughts*. It presents an in-depth analysis of the causal information transfer between the Pacific Ocean and the Tropical North Atlantic with and without the presence of the Amazon River basin. This chapter provides further evidence on the way the Amazon River basin act as an Land-Atmosphere bridge between the two tropical oceans for the particular drought of 2010. The analysis of causal information transfer is developed using a new recurrence measure of causality, which was developed in the framework of this doctoral research in collaboration with researchers from the National Institute for Space Research in Brazil and the Potsdam Institute for Climate Impact Research, published in *Physical Review E* as *Recurrence Measure of Conditional Dependence and Applications*. The measure and applications has been presented in the *2016 AGU Fall Meeting* held in San Francisco, CA, USA, during 12-16 of December, and the *8th SOLSTICE Conference*, held in Aveiro, Portugal during 20-22 of June, 2016.
Finally, in Chapter 5 we draw the general conclusions and propose new perspectives and future work.
1. Conjoint Analysis of the Surface and Atmospheric Water Balances in the Andes-Amazon System

Alejandro Builes-Jaramillo and Germán Poveda

Abstract: A comprehensive analysis of surface and atmospheric water balances in the Amazon River basin is performed with an emphasis on the linkages between the Andes and low-lying Amazonia. Estimates are made regarding the closure of these regions’ long-term water budgets and their respective imbalance. Analyses are performed using observational and reanalysis data sets covering five different periods of analysis (depending on the length of evapotranspiration data sets) for the entire basin, major subcatchments and for the Andes and low-lying Amazonia. Regarding the surface water balance, although the results show that the entire basin might be considered as being in balance, spatial disaggregation of data produced a less clear picture, presenting major discrepancies between observations and reanalysis data sets. The atmospheric budget exhibits no closure and positive residuals regardless of data set, with the magnitude of residuals for the entire Amazon basin highly dependent on evaporation data source. The imbalance between the two branches of the hydrological cycle (14%-16%) is driven by higher values of runoff and by an abrupt change in runoff when changing from dry to wet seasons in the Amazon. By disentangling the entire river basin into the Andes and low-lying Amazonia sub-systems, we unveil two shortcomings of the available data, namely poor quality in the representation of surface processes by reanalysis, and flaws and scarcity of information in the high Andes that induce uncertainties and errors in both water budgets. The results of the present study highlight the importance of the Andean region for the hydrological integrity of the entire Amazon River basin. Our results confirm the paramount importance of the joint analysis of atmospheric and surface water budgets at the basin level with respect to achieving a complete understanding of the whole hydrological cycle at the continental scale.

Key words: Amazon River basin, Amazonia, Andes, Surface Water Balance, Atmospheric Water Balance, Tropical Hydroclimatology

1.1 Introduction

The hydrological cycle connects diverse water reservoirs (oceans, soils and groundwater, cryosphere, atmosphere and biosphere) through diverse flows (precipitation, evapotranspiration, infiltration, river flows, and advection of water vapor by wind). The conservation of the amount of water on Earth is a
well-known fact, with the physical description of the water balance in the hydrological cycle an important research topic for hydrologists and climatologists alike [Budyko and Miller, 1974; Trenberth et al., 2007, 2011]. In general, most water balance studies have focused on either the surface or atmospheric branches of the hydrological cycle; only a few have dealt with their coupling at different temporal and spatial scales [Peixoto and Oort, 1993; Healy et al., 2007].

It is also widely known that the atmospheric water balance over oceanic regions is not well represented in either reanalysis or satellite-based data sets, especially for high precipitation regions such as the Intertropical Convergence Zone (ITCZ) [Trenberth and Smith, 2005; Park et al., 2013]. Errors and uncertainties may be attributed to shortcomings associated with interpolation algorithms, the parameterization of diverse hydro-meteorological processes in various reanalysis products, or to the fact that they ignore that the atmosphere’s total ‘dry-air mass’ is invariant [Takacs et al., 2016]. Whereas estimates of the atmospheric water budget are largely constrained by the scarcity of actual measurements, in particular regarding the transport of water vapor by the winds through the troposphere, it is much easier to estimate the water balance over land, given the availability and small uncertainty of river flow data.

The hydro-climatological importance of the Amazon River basin at global and continental scales cannot be overstated, given its size of around 7.0x106 km2 and its tropical setting with a large portion covered by rainforests, which together contribute to explaining its mean annual discharge at the outlet of more than 200,000 m3s⁻¹ [Dickinson, 1989; Nobre and Borma, 2009; Medvigy et al., 2011; Swann et al., 2015; Trumbore et al., 2015]. The basin plays a fundamental hydro-climatological role in Earth’s climate and exhibits various stability characteristics associated with land cover types and extent [Lenton et al., 2008; Keller et al., 2009; Nobre and Borma, 2009; Hirota et al., 2011]. Understanding the dynamics of the hydrological cycle of the Amazon River basin has been the focus of many previous studies, some interested in individual branches of the hydrological cycle [Costa and Foley, 1997; Zeng, 1999; Marengo, 2004; Getirana et al., 2014], others in one particular variable of the hydrological budget over the Amazon River basin, such as precipitation in particular seasons [Costa and Foley, 1998; Liebmann et al., 1998; Liebmann and Marengo, 2001; Espinoza et al., 2009], runoff [Coe et al., 2002; Syed et al., 2005; Azarderakhsh et al., 2011], evaporation [Hasler and Avisar, 2007; Carmona et al., 2016; Mallick et al., 2016], and wind divergence [Liebmann et al., 1998; Zeng, 1999].

The northwestern regions of the Amazon River basin are subject to the largest mean annual rainfall and are mostly covered by tropical rainforests. Whereas Andean Amazonia exhibits strong physiographic, orographic, ecological, hydrological and climatic gradients, including tropical glaciers, páramos, yungas, montane forests, tropical rainforests and savannas, low-lying Amazonia is characterized by small slopes, less strong hydrological gradients,
and less land cover variability. In fact, the upper Andean and low-lying regions of Amazonia constitute two coupled sub-systems, with the former importing atmospheric water via trade winds from the latter’s forests, and in turn exporting water (including sediments, nutrients, and biogeochemical tracers) [Poveda et al., 2006; Poveda and Zapata, 2016].

The Amazon River basin exhibits a strong north-south differential in terms of its hydrological regime, with distinctive annual cycles observed in the two regions [Marengo, 2009]. The present study focuses on the hydrological interactions between the Andes and the low-lying portion of the Amazon River basin (Amazonia, hereafter), given that the portion of the basin located in the Andes is well known for its regulation of surface water supply to the entire basin [McClain and Naiman, 2008]. In addition, the Andes drive wind circulation from the Amazon to southeastern South America and also exert an important influence on the forcing of winds and convection along the hillsides of the mountain range [Marengo, 2004; Poveda et al., 2006, 2014; Garreaud et al., 2009], thus playing a fundamental role in the water balance of the entire Amazon.

Several studies investigating the long-term water balance of the Amazon River basin have focused on understanding the surface branch [Villa Nova et al., 1976; Marques et al., 1980; Jordan and Heuveldop, 1981; Leopoldo et al., 1982; Franken and Leopoldo, 1984; Shuttleworth, 1988; Vörösmarty et al., 1989; Russell and Miller, 1990; Costa and Foley, 1999; Zeng, 1999; Marengo, 2004]. In general, these studies have addressed water balance at the Obidos gauging station, with some reporting an exact (see Appendix A, Table A1) closure accounting for a perfect surface balance in the basin. However, such results of perfect closure are highly influenced by the methodologies employed when estimating evapotranspiration, with estimations that are mostly based on the precipitation product used in the closure calculations providing results that are not completely accurate due to dependence between data sets. The imbalance of the surface water budget found in the Amazon River basin [Marengo, 2004, 2006] has previously been explained as a misrepresentation of the biological fluxes over the basin that may in turn affect the estimation of evapotranspiration [Makarieva et al., 2012, 2013; Jasechko et al., 2013; Drumond et al., 2014].

Despite the aforementioned studies examining the surface water balance of the Amazon River basin, no consistent effort has been made to understand/quantify the conjoint dynamics of the water balance over the land and atmospheric branches of the hydrological cycle [Peixoto and Oort, 1993]. Therefore, the present study was aimed at quantifying the conjoint long-term water balance over the surface and atmospheric branches of the Amazon River basin and its major sub-basins, with a particular emphasis on the functioning of the Andes-Amazônia system.
The paper is organized as follows: In section 2 we present the methods and data used in the analysis. Section 3 presents the results obtained and a subsequent discussion regarding the closure of the surface and atmospheric water balances for the entire Amazon basin and its main sub-catchments, as well as the closure of the conjoint water balances (3.1), long-term imbalances between the surface and atmospheric balances (3.2), and the analysis of water (im)balances for the Andes-Amazônia system (3.3). Finally, section 4 ends the paper with some concluding remarks.

1.2 Methods and Data Sets

1.2.1 Surface and Atmosphere Water Balances

Following the two-box model of the hydrological cycle [Zeng, 1999] presented in Fig.1-1, the surface water balance (SWB) and atmospheric water balance (AWB) can be expressed in the form of the following differential equations representing the conservation of mass over both control volumes:

\[
\frac{\partial S(t)}{\partial t} = P(t) - E(t) - R(t) \quad (1-1)
\]

\[
\frac{\partial W(t)}{\partial t} = -P(t) + E(t) + C(t) \quad (1-2)
\]

where \( S \) represents soil water storage [L], \( P \) is precipitation rate [LT\(^{-1}\)], \( E \) is evapotranspiration rate [LT\(^{-1}\)], \( R \) is runoff rate [LT\(^{-1}\)], \( W \) is the amount of precipitable water [L], and \( C \) is the net moisture convergence (\( C = -\nabla Q \) where \( \nabla Q \) is the net vertically integrated moisture divergence) [LT\(^{-1}\)]. The change in both storage terms over time as represented by \( \partial W/\partial t \) and \( \partial S/\partial t \) is negligible in the long term [Rasmusson, 1968; Peixoto and Oort, 1993; Marengo, 2004; Poveda et al., 2007; Zhang et al., 2008], such that for \( t \to \infty \), equations (1-1) and (1-2) become, respectively

\[
\langle R \rangle = \langle P \rangle - \langle E \rangle \quad (1-3)
\]

\[
\langle C \rangle = \langle P \rangle - \langle E \rangle \quad (1-4)
\]

where the symbol \( \langle \ \rangle \) denotes the long-term mean (\( t > 20 \) yr) of the respective variable. For simplicity, we can disregard the \( \langle \ \rangle \) symbol and thus \( P, E, R, C \) will hereafter denote long-term mean values.
Figure 1-1. Illustration of the hydrological cycle and its component control volumes as defined for surface and atmospheric branches. Variables inside a circle represent terms associated with precipitable water storage and soil water storage (\(W\) and \(S\)). Precipitation and Evapotranspiration rates (\(P\) and \(E\)) are present in both branches, while Convergence and Runoff (\(C\) and \(R\)) are considered as inflows and outflows for the atmospheric and surface branches, respectively.

Equations (1-3) and (1-4) imply that, in the long term, atmospheric water convergence is equal to long-term surface runoff, or \(C = R\). One of the main objectives of the present study was to verify the validity of such equality in the long-term water balances of the surface and atmospheric branches of the entire Amazon River basin, its major sub-basins, and the Andes-Amazonia system. Large uncertainties in the estimation of \(E\), possible losses of underground water and shortcomings in the models used to represent the atmospheric budget may result in an imbalance between the two branches of the hydrological cycle.

The imbalance between the SWB and AWB can be quantified through the following expression [Marengo, 2004]:

\[ I = \frac{C}{R} - 1 \] (1-5)

Regarding the AWB over the Amazon River basin, Makarieva et al. [2012, 2013] have argued that the results of previous studies that report such an imbalance [Marengo, 2006; Espinoza Villar et al., 2009] can be explained by the misrepresentation of water vapor released to the atmosphere from forest evapotranspiration and the reduction in precipitation that may be attributable to deforestation in the Amazon, which upon condensation can be converted into potential energy available to drive winds and moisture into the continent.
1.2.2 Data

Evapotranspiration information and data sets were obtained from five different sources: (1) The ORCHIDEE Land Surface Model (LSM) [Krinner et al., 2005] data set that provides estimates of total evaporation based on runs of the SECHIBA hydrological model coupled to the STOMATE carbon model (STOMATE). The model is forced with NCEP/NCAR reanalysis and ORE-HYBAM project precipitation (http://www.ore-hybam.org/), while the data set itself has a spatial resolution of 1° x 1° and a monthly time resolution spanning the period from 1970 to 2008; (2) Max Plank Institute (MPI) ensembles of monitoring networks, meteorological observations and remote sensing were used to produce estimates of evaporation at 0.5° x 0.5° spatial resolution and a monthly time resolution for the period from 1982 to 2008 [Jung et al., 2010]; (3) The Global Land Evaporation Amsterdam Model (GLEAM), which uses satellite information from different vegetation indices and estimates values of total evaporation [Miralles et al., 2011], at 0.25° x 0.25° spatial resolution and using monthly data for the period from 1984 to 2007; (4) The MODIS MOD16A product includes estimates of evapotranspiration based on MODIS surface temperature and vegetation indices, at a spatial resolution of 0.01° x 0.01° every 8 days, here aggregated at 1.0° x 1.0° and monthly timescales for the period from 2000 to 2010 [Mu et al., 2011; Nadzri and Hashim, 2014]; (5) ERA-Interim reanalysis variable total evaporation data with a spatial resolution of 0.75° x 0.75° and a monthly time resolution for the period from 1979 to 2014 [Dee et al., 2011].

Monthly precipitation data sets were obtained from the Global Precipitation Climatology Center (GPCC), produced by the Deutscher Wetterdienst (gpcc.dwd.de) at 1° x 1° spatial resolution [Schneider et al., 2014] for the period from 1901 to 2010, and from ERA-Interim reanalysis with a spatial resolution of 0.75° x 0.75° [Dee et al., 2011] spanning the period from 1979 to 2014. Monthly river flows for the period from 1982 to 2008 were obtained from the SO-HYBAM Observation Service (formerly the ORE-HYBAM Environmental Research Observatory), available at http://www.ore-hybam.org/. River discharges (D) were transformed into runoff (R) considering that $R = D/A$, with A being the drainage area of each sub-catchment within the Amazon River basin. R was also obtained from Era-Interim reanalysis at a spatial resolution of 0.75° x 0.75° and monthly resolution for the period from 1979 to 2014, together with monthly vertically integrated moisture flux divergence ($\nabla Q$) data for the same resolution and time period. Values of $P$, $E$ and $\nabla Q$ were averaged for the entire basin and for each sub-catchment within the basin for which information regarding $R$ was available.

Two types of data set were defined in order to estimate the water balances, one comprising mostly observations or land surface models, and the other obtained from reanalysis products. All data sets corresponding to $P$, $E$, $R$ and $C$ were employed to estimate the long-term water balances and their
independent (surface and atmosphere) and conjoint closures. The observational data sets included those of \( P \) (GPCC), \( E \) (ORCHIDEE, GLEAM, MPI and MODIS) and \( R \) (HYBAM), while the reanalysis data sets of \( P \), \( E \), \( R \) and \( R' \) were retrieved from Era-Interim reanalysis. For estimation purposes, the long-term surface water balance over the Andes-Amazonia system was estimated using data pertaining to 115 sub-catchments out of 146 (due to the availability and quality of the HYBAM data set), whereas the reanalysis data sets covered all 146 sub-catchments.

It is worth noting that all data sets contained information recorded at different spatial resolutions, the most extreme example being the \( E \) data sets ranging from 10,000 km\(^2\) (ORCHIDEE) to 1 km\(^2\) (MODIS), while the sub-catchment areas also varied, from 465 km\(^2\) (Base do Cachimbo) to 4.6x10\(^6\) km\(^2\) (Obidos). Therefore, we expect that such differences in data set resolution and catchment areas represent a major source of uncertainty in our analysis. In addition, some uncertainty in the results may also be attributed to the way in which information was produced; while \( P \) and \( E \) are gridded data, \( R \) values were derived from continuous observation of actual outflows measured along the catchments.

### 1.3 Results and discussion

#### 1.3.1 Closure of water balances

##### 1.3.1.1 Closure of the surface water balance

From eqn. (1-3) it is possible to state that, in the long term, \( P-E-R = 0 \), and thus one can define the closure of the SWB in terms of such a residual. If the result is significantly different from 0, one can conclude that there is no closure of the water balance equation (imbalance) in a river basin, with \( P-E-R > 0 \) representing an excess or surplus of water, and \( P-E-R < 0 \) representing a deficit of water in the basin. Errors in closure can be explained by: (i) data quality, (ii) record length, (iii) availability of observations within the basin, (iv) indirect or modeled measures of evapotranspiration and precipitation, (v) changes in soil water storage, and (vi) trends in the hydrological variables associated with climate change and/or deforestation, which violate the mean value theorem implicit in going from eqn. (1) to eqn. (3). Such trends have been reported for the Amazon River basin in many studies [Marengo, 2009; Gloor et al., 2013; Espinoza et al., 2016; Posada and Poveda, 2017]

Table 1-1 shows the results of SWB closure reported in previous studies, as well as those obtained in the present study using different combinations of data sets for various periods of analysis. Using different combinations of \( P \), \( E \) and \( R \), it is possible to identify positive and negative residuals of diverse
Conjoint Analysis

magnitudes regardless of observation or reanalysis data set. According to Table 1-1, the SWB with the best closure (11 mm year\(^{-1}\), 0.5\% of \(P\)) is that calculated with GPCC rainfall data, GLEAM for evapotranspiration, and HYBAM for runoff. This percentage of \(P\) required to define the closure is around the same order of magnitude as values reported in previous studies (see Table 1-1), but lower than those found in balances computed using variables from remote sensing and observations [Sahoo et al., 2011].

Here the use of observations produced positive and negative residuals ranging from 197 mm year\(^{-1}\) to -30 mm year\(^{-1}\), while the reanalysis data set provided a negative residual of -51 mm year\(^{-1}\) (Table 1-1). Computing mean \(P\), \(E\) and \(R\) based on observations and reanalysis, we found a positive budget residual of 41 mm year\(^{-1}\) for the entire basin. Our results, similar to those of most of the earlier studies presented in Table 1-1 and Table A1, confirm that the Amazon River basin can be considered in balance in terms of the SWB, assuming a closure criterion of ± 10\% \(P\) is acceptable to consider a river basin as being in balance. Nevertheless, an important question remains regarding the validity of water balance closure along the different major sub-basins of the entire Amazon River basin; this issue is discussed in the following section.

Table 1-1: Comparison between previous studies of surface water balance (SWB), atmospheric water balance (AWB) and imbalance, and the results obtained for the data sets used in the present study.

### Surface water balances (SWB)

<table>
<thead>
<tr>
<th>Previous Studies</th>
<th>(P) (mm y(^{-1}))</th>
<th>(E) (mm y(^{-1}))</th>
<th>(R) (mm y(^{-1}))</th>
<th>(P-E-R) (mm y(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>[Salazar, 2004]</td>
<td>2189</td>
<td>1248</td>
<td>940</td>
<td>1</td>
</tr>
<tr>
<td>[Marengo, 2004]</td>
<td>2117</td>
<td>1570</td>
<td>1059</td>
<td>-511</td>
</tr>
<tr>
<td>[Carmona, 2015]</td>
<td>2266</td>
<td>1189</td>
<td>1163</td>
<td>-86</td>
</tr>
<tr>
<td><strong>Present study sources P-E-R</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(period)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GPCC-ORCHIDEE-HYBAM (1979-2008)</td>
<td>2225</td>
<td>1062</td>
<td>965</td>
<td>197</td>
</tr>
<tr>
<td>GPCC-MPI-HYBAM (1982-2008)</td>
<td>2225</td>
<td>1181</td>
<td>965</td>
<td>78</td>
</tr>
<tr>
<td>GPCC-GLEAM-HYBAM (1984-2007)</td>
<td>2225</td>
<td>1248</td>
<td>965</td>
<td>11</td>
</tr>
<tr>
<td>Era-Interim (1979-2012)</td>
<td>2190</td>
<td>1301</td>
<td>939</td>
<td>-51</td>
</tr>
<tr>
<td>Mean value</td>
<td>2218</td>
<td>1216</td>
<td>960</td>
<td>41</td>
</tr>
</tbody>
</table>

### Atmospheric water balances (AWB)

<table>
<thead>
<tr>
<th>Previous studies</th>
<th>(P) (mm y(^{-1}))</th>
<th>(E) (mm y(^{-1}))</th>
<th>(C) (mm y(^{-1}))</th>
<th>(P-E-C) (mm y(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>[Zeng, 1999]</td>
<td>2044</td>
<td>1679</td>
<td>292</td>
<td>73</td>
</tr>
<tr>
<td>[Costa and Foley, 1999]</td>
<td>2322</td>
<td>1384</td>
<td>905</td>
<td>33</td>
</tr>
<tr>
<td>[Marengo, 2004]</td>
<td>2117</td>
<td>1570</td>
<td>511</td>
<td>36</td>
</tr>
<tr>
<td><strong>Present study sources P-E-C</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(period)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GPCC-ORCHIDEE-ERA (1979-2008)</td>
<td>2225</td>
<td>1062</td>
<td>810</td>
<td>353</td>
</tr>
</tbody>
</table>
Imbalances between SWB and AWB

<table>
<thead>
<tr>
<th>Previous studies</th>
<th>C (mm y(^{-1}))</th>
<th>R (mm y(^{-1}))</th>
<th>R-C (mm y(^{-1}))</th>
<th>I (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>[Zeng, 1999]</td>
<td>292</td>
<td>1095</td>
<td>803</td>
<td>-73</td>
</tr>
<tr>
<td>[Costa and Foley, 1999]</td>
<td>905</td>
<td>937</td>
<td>32</td>
<td>-3</td>
</tr>
<tr>
<td>[Marengo, 2004]</td>
<td>511</td>
<td>1058</td>
<td>547</td>
<td>-51</td>
</tr>
<tr>
<td>Present study sources (C-R)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(ERA-HYBAM)</td>
<td>810</td>
<td>965</td>
<td>155</td>
<td>-16</td>
</tr>
<tr>
<td>(ERA-ERA)</td>
<td>810</td>
<td>939</td>
<td>129</td>
<td>-14</td>
</tr>
</tbody>
</table>

Long-term SWB for the major sub-catchments of the Amazon River basin were thus estimated using observational and reanalysis data sets. Long-term average values of \( P \), \( E \) and \( R \) were estimated for each sub-catchment, with observed values of \( R \) obtained using the HYBAM data set. Fig. 1-2 shows the estimated SWB closures for the Amazonian sub-catchments using each data set.

Figs. 1-2a to 1-2d present the results for the four combinations of observational data sets as detailed in Table 1-1; differences between the data sets include the \( E \) source (a. ORCHIDEE, b. MPI, c. GLEAM, d. MODIS) and the length of the study period (a.1979-2008, b. 1982-2008, c. 1984-2007, d. 2000-2008). Analysis of Figs. 1-2a to 1-2d reveals a common geographical pattern of positive residuals (green colors) in the southeast portion of the basin, with negative residuals appearing in the central and northwest regions, as well as in sub-catchments with extensive areas such as Obidos and Itaipue. There is a prevailing negative residual in two Andean sub-catchments, Borja and Rurrenabaque, both of which lie inside the region of positive residuals; this might reflect the high evapotranspiration and low precipitation rates recorded in these high altitude regions (Fig. 1-2). In addition, a negative closure pattern is observed in the northwest region that is characterized by the outflow of high amounts of runoff towards low-lying Amazonia [McClain and Naiman, 2008].

Fig. 1-2a displays the results of SWB closure for those observational data sets with longer record lengths. For this particular combination of \( P \), \( E \) and \( R \) data sets (GPCC-ORCHIDEE-HYBAM), 63 sub-catchments are in balance (±10% of \( P \)), indicating that there is balance in only 55% of catchments, while out of the 115 sub-catchments 75% have a positive residual. This suggests that (using observations) \( P \) has the largest influence on water balance closure as the dominant term, as was also found for the long-term water balance (see section 3.2), with residuals varying from -2000 mm year\(^{-1}\) to 1121 mm year\(^{-1}\).
(90% of \( P \)). Similar analyses were performed for the remaining combinations of observational data sets (Figs. 1-2b to 1-2d).

The SWB estimated using reanalysis data (Fig. 1-2e) is characterized by higher negative residuals in all sub-catchments, in particular in the Andean region. Compared with those estimated from observational data sets, the reanalysis residuals are lower in magnitude, ranging from -340 mm year\(^{-1}\) to 40 mm year\(^{-1}\). Our results indicate that out of the 146 sub-catchments, 135 (92%) are in balance (±10% of \( P \)) and 110 (85%) have negative residuals. Unlike those based on observation data, the sub-catchment reanalysis results are in agreement with the results obtained for the entire basin, taking into account balance closure and the sign of residuals. The negative residuals found in the reanalysis SWB may be attributed to the higher amplitudes in the magnitudes of \( R \) and \( E \), not only for the long-term mean balance, but also for the annual cycle (Table 1-1). The prevalence of negative residuals in the reanalysis data set is indicative of more \( R \) than \( P-E \) in the basin at the monthly scale, and therefore one may assume that other sources of water, such as soil moisture or groundwater storage, may be overestimated in the models.

The above results demonstrate that SWB closure is independent of drainage area regardless of the evapotranspiration data set, and the same conclusion also holds for the closure sign (Appendix, Fig. A3). Such random closure behavior in catchments may be related to a possible misrepresentation of \( E \) throughout the entire river basin, with this particular variable the most uncertain due to difficulties in obtaining precise in-situ measurements and regional estimates. Shortcomings in \( E \) data, regardless of their origin in land-surface regional models, remote sensing or reanalysis, are the reason why this variable is not completely understood or well represented at the sub-catchment level. Finally, one can conclude that the picture of near balance found for the entire basin (11 mm month\(^{-1}\), 0.5% of \( P \)) tends to change upon spatial downscaling at the sub-catchment level, such that a conclusion of balance depends on scale.
1.3.1.2 Closure of the atmospheric water balance

We also studied the long-term atmospheric water balance (AWB; eqn. 1-4) for both the whole Amazon River basin and its main sub-catchments. Table 1-1 shows the results for AWB closure reported in previous studies, as well as those obtained with the different data sets used in the present study. Overall, the $P-E-C$ residuals are positive, varying from 353 to 78 mm year$^{-1}$ (15% – 4% with respect to mean annual $P$), indicating that closure depends on the observational data sets used to calculate the AWB (according to the ± 10% of $P$ criterion). Balance was achieved here using the GLEAM (7.5%) and MODIS...
(5.6%) observational data sets of E, but not with ORCHIDEE (15.8%) and MPI (10.5%) observational data.

The closure results also show that residuals produced here via the observational data sets are larger than those reported by previous authors [Costa and Foley, 1999; Zeng, 1999; Marengo, 2004]. This discrepancy is partly due to our use of recently available E data sets, but also due to differences in periods of analysis, advances in the representation of C by reanalysis and the averaging of R in the basin, and the occurrence of extreme drought events in the Amazon such as those of 2005 and 2010. In the present study, we used data sources that may be considered independent, thus assuring that variables used in the balance computations were not derived from one another, with the exception of the reanalysis data set. In the study conducted by Marengo (2004), E and C were obtained from reanalysis and thus these two variables cannot be considered independent. Furthermore, the study period in the latter paper ran from 1970 to 1999. In the study carried out by Zeng (1999), the balance was computed with E derived from P (and thus these two variables were not independent), while the study period spanned from 1985 to 1993. Finally, the results obtained by Costa and Foley (1999) were derived from NCEP/NCAR reanalysis, spanning the period from 1976 to 1996.

The smallest AWB residual was obtained using the reanalysis data set (78 mm year\(^{-1}\); 4 % of P), although this value is still higher than those reported by the other authors (Table 1-1). A positive value of the residual P-E-C indicates a higher amount of atmospheric water available for precipitation, while a negative residual might be related to advection of moisture in the atmospheric column that reduces the availability of water over that particular region. Although the P-E-C residual values obtained here indicate that P > E+C for all data sets (Table 1-1), moisture convergence is always positive in most of the basin (Appendix A, Fig. A1) and the sources of P may vary throughout the year, with C higher in magnitude than E during the wet season (Jan-Apr), and E the main source for P in the rest of the year (Appendix A, Fig. A2). These details confirm the importance of land surface-atmosphere feedbacks and the role of evapotranspiration in precipitation recycling throughout the Amazon River basin. In fact, the role of E in precipitation in the Amazon cannot be overstated, given that almost 25%-50% of E in the basin contributes to the formation of precipitation due to recycling during the dry season [Eltahir and Bras, 1994; Malhi et al., 2008; Zemp et al., 2014, 2017; Rocha et al., 2015].

AWB closure was also computed for 146 Amazonian sub-catchments (Fig. 1-3), with values ranging between -1720 mm year\(^{-1}\) and 2270 mm year\(^{-1}\) obtained using the roughest spatial resolution data sets (GPCC(P)-ORCHIDEE(E)-ERA(C)). For these particular data sets, none of the sub-catchments is in balance, and the one in near balance (Base Alalau) has a closure of 27 mm year\(^{-1}\) (14.8% of P in that catchment). Imbalances of the same order of magnitude were also obtained with the remaining observational data sets (not shown). Residuals calculated using the reanalysis data set vary
from -295 mm year^{-1} to 1,211 mm year^{-1}; only four basins are balanced with a residual of ±10 mm year^{-1} (-1.5% to 3.5% of $P$), while most sub-catchments exhibit positive residuals (more than 75% in all data sets). This pattern of positive residuals is observed regardless of observational or reanalysis data set, or the employed P source (GPCC and ERA-Interim reanalysis; see Appendix A; Fig. A4).

The spatial distribution of residuals is characterized by positive values for most of the Amazon River basin. Negative AWB residuals are observed over the Altiplano region, specifically in the Rurrenabaque catchment and in the eastern-most part of the basin in the Boa Esperança catchment (Figs. 1-3a to 1-3d). This relationship between AWB closures and drainage areas illustrates the latter’s independence (Appendix A, Fig. A6).

The high AWB closure values reported in the present study for all data sets may be attributed to several factors, such as uncertainties in estimates of $E$ due to a lack of available accurate observational data sets for the region, limitations in the modeling of $C$ due to the misrepresentation of the influence of vegetation on biogeochemical fluxes [Makarieva and Gorshkov, 2007; Brando et al., 2010], as well as the contribution of evapotranspiration to atmospheric water content and the misrepresentation of the influence of the Andes on the dynamics of diverse hydrometeorological processes in upper Amazonia [McClain and Naiman, 2008; Insel et al., 2009].
Figure 1-3: Closure of the AWB in the Amazon basin sub-catchments based on the E data sets used: a) ORCHIDEE, b) MPI, c) GLEAM, d) MODIS, and e) ERA. Green values denote positive closures; yellow to red values denote negative closures.

### 1.3.1.3 Closure of the conjoint surface-atmospheric water balances

The conjoint closure of long-term surface (SWB) and atmospheric (AWB) water balances (eqn. 1-5) was studied for the entire Amazon River basin, using runoff data derived from both observations (HYBAM) and reanalysis (Era-Interim). For the period of analysis (1982 – 2008), estimated values of $R$ are larger than those of $C$, such that $R-C= 356 \text{ mm year}^{-1}$ (observations) and $R-C= 129 \text{ mm year}^{-1}$ (reanalysis), thus producing negative residuals (Table 1-1). According to these results, one can infer that there is water that is not included in the atmospheric convergence values, and therefore $C$ cannot
match long-term runoff in the basin. The obtained results are also influenced by all the aforementioned sources of uncertainty and errors affecting both the SWB and AWB.

Previous studies have found similarly negative imbalances \((R>C)\) for the entire Amazon River basin \cite{Costa, Foley, 1999; Zeng, 1999; Marengo, 2004}. Zeng \cite{1999} reported the highest negative imbalance, with \(R\) 50\% higher than \(C\), which might be related to an underestimation of moisture convergence using GEOS-1 data. Costa and Foley \cite{1999} produced an imbalance of 3\%; such a small imbalance may be explained by the use of the same reanalysis for both variables (NCEP-NCAR), with these data sets thus interrelated and dependent. Although Marengo \cite{2004} used reanalysis and observations as in the present study, the former reported an imbalance of around 50\%, much higher than the value of around 15\% calculated here, suggesting that reanalysis is providing a more accurate representation of water balance variables over time.

We then quantified conjoint SWB and AWB closure residuals for all sub-catchments (115 for observations and 146 for reanalysis) in the Amazon River basin (Figs. 1-4a and 1-4b). A stringent criterion was also defined when assessing the imbalance in order to find significant differences between data sets, with a sub-catchment defined as being in balance when the value of I (eqn. 1-5) is in the range of \(\pm 5\%\). According to this criterion, 4 sub-catchments (3.5\%) are in conjoint balance based on observation data \cite{Estirao Da Santa Cruz, Serrinha Fazenda Sao Jose, Boa Sorte}, while 26 sub-catchments are in balance (17.8\%) based on reanalysis, suggesting that the balances estimated via the latter are more accurate. However, it is worth noting that in such cases we are comparing \(R\) and \(C\) from the same source, which may induce false conclusions as information may be redundant and dependent. For both data sets, most sub-catchments exhibit negative imbalances \((R>C)\), with 87\% recorded for observational \(R\) (HYBAM) and 90\% for reanalysis \(R\) (ERA). The spatial distribution of imbalances exhibits a clear geographic pattern of negative values throughout Amazonia and positive values in the Altiplano region \cite{Rurrenabaque} sub-catchment. Furthermore, the imbalances do not seem to depend on drainage areas (Figs. 1-4c and 1-4d).
Figure 1-4: Imbalance between SWB and AWB in the Amazon basin sub-catchments: (a) R from HYBAM and (b) R from Era-Interim. Log-log relationship between imbalance and catchment areas in the Amazon river basin, estimating R using (c) HYBAM and (d) Era-Interim.

Fig. 1-5 displays a summary of all the variables involved in SWB and AWB for both observational and reanalysis data sets. The results show that precipitation is the main driver of dynamics in both SWB and AWB, and that observation data produced higher values of P and R than those calculated via reanalysis.
1.3.2 Annual Cycle of the long-term water balances

Estimation of the long-term mean annual cycles of the water balance variables was carried out by calculating monthly means based on averaged time series for the Amazon River basin, with the budget annual cycle calculated using the storage terms $\partial S(t)/\partial t$ and $\partial W(t)/\partial t$ as the residuals of the remaining variables in the balance equations (eqn. 1-1 and eqn. 1-2). The annual cycle of $P$ in the Amazon River basin is unimodal, with a wet season from November to April and a dry season from May to October, as shown in Fig 1-6a. Both data sets (GPCC and Era-Interim) exhibit similar behavior in the amplitude and phase of the annual cycle of $P$ when averaged for the entire basin. The main differences are observed in the maximum values of precipitation from January to April and in the peak of the dry season, which occurs in August for GPCC and July for Era-Interim data. Similarities in the spatial distribution of $P$ (Figs. 1-6c and 1-6d) can be observed between the data sets, with maximum precipitation values in the northwest of the river basin and in the southwest along the Andes. Precipitation patterns are similar to those reported in previous studies [Espinoza et al., 2009; Garreaud et al., 2009; Nobre et al., 2009].
Figure 1-6b shows the annual cycle of $E$ for both data sets, as well as the annual cycle computed from Era-Interim reanalysis in the Amazon River basin (dotted line) and a composite of the remaining data sets (ORCHIDEE, MPI, GLEAM and MODIS) for the 1979-2008 period, computed as the mean of the four annual cycles (dashed line). The long-term mean annual cycle of $E$ exhibits unimodal behavior, with higher values from October to March and lower values from April to September. The pattern of evapotranspiration in the Amazon is nearly in phase with precipitation, meaning that the main drivers of evapotranspiration in the basin are the availability of water given by precipitation and the energy given by insolation. The dynamics of water/energy availability in the basin also configures the recycling of precipitation from evapotranspiration [Malhi et al., 2008; Zemp et al., 2014; Rocha et al., 2015].

The results shown in Fig 1-6b evidence that although both data sets exhibit a similar annual cycle of $E$, the magnitude of these cycles differs, with that derived from reanalysis always smaller than that obtained from observations, ranging from 10 mm month$^{-1}$ in June to less than 5 mm month$^{-1}$ in February. Appendix A (Fig. A7) presents a comparison of the annual cycles for all sources of $E$. The differences in cycles may be explained by data assimilation in each source. During the dry season, the Amazon forest tends to increase its activity due to the availability of radiation, enhancing vegetation transpiration and thus providing more moisture to the atmosphere [Huete et al., 2006; Samanta et al., 2010; Harper et al., 2014; Salazar et al., 2016]. As climate models and parameterizations do a poor job in representing these processes, any water not accounted for in the atmosphere might be related to shortcomings in the representation of evapotranspiration. There are several indices that account for vegetation activity (VI) as processed and estimated using remote sensing information, such as the normalized difference vegetation index (NDVI), the enhanced vegetation index (EVI), as well as composite indices constructed with several variables, such as the leaf area index (LAI). Era-Interim reanalysis uses LAI in its evaporation from vegetation scheme [Boussetta et al., 2013]. However, the estimation of evaporation from LAI is highly sensitive to LAI estimation itself [Rogers, 2013]; according to several studies, Amazon vegetation activity and the formation of new leaves in the dry season are better represented by EVI or MODIS data [Brando et al., 2010; Hilker et al., 2015].

One can also observe that in four months of the year (August through November), estimates of $E$ based on observations are lower those derived from reanalysis, while for the remaining months $E$ is consistently higher in observations. The Era-Interim data set (Fig. 1-6e) exhibits a homogeneous spatial pattern of evapotranspiration over central Amazonia, with values around 100 mm month$^{-1}$, while ORCHIDEE (Fig. 1-6f) exhibits more spatial variability of between 50 and 100 mm month$^{-1}$. The long-term mean annual cycles of $E$ for both data sets exhibit similar phases, with maximum values in March and minimum values in June. Fig. 1-7 shows the long-term monthly mean values of $R$ and $C$. The long-term mean values of such variables should
be in balance, allowing the connection of the two water budgets. Indeed, the runoff data set presents similar phases and amplitudes, with a dry season from January to May and a wet season from July to November. Maximum values of $R$ are around 110 mm month$^{-1}$ in the peak wet season and minimum values are around 30 mm month$^{-1}$ in the peak dry season.

Although the amplitudes of the cycles presented in Fig. 1-7 are similar to those estimated in previous studies [Zeng, 1999; Marengo, 2009], the peaks are out-of-phase, with the peak of $R$ in the present study (Mar-Apr) occurring before those reported elsewhere (Apr-May). Such differences in phase are likely caused by the previous use of information regarding basin outflow in Obidos, whereas we used an average of basin runoff. The annual cycle of $C$ exhibits two peaks, one in February of 110 mm month$^{-1}$ and a minimum of around 10 mm month$^{-1}$ in August; these peaks do not coincide with those of $R$ in phase or magnitude, but are in phase with those of $P$.

Fig. 1-8 shows the long-term monthly means of the surface water budget for both data sets. The two exhibit common features, such as a higher magnitude of $P$, which is the dominant variable throughout the year and is always higher than $E$, thus confirming that the basin acts a moisture sink ($P>E$) over the year. One can observe that for the Era-Interim data set, the monthly values of $R$ are higher than those of $E$ for the period Feb-May, while for the observational data set $R$ is higher than $E$ in the period Jan-July. It is worth noting that for both data sets the maximum and minimum values of the annual cycle of $R$ lag between one and three months behind those of $P$ (dry and wet), as also found in previous studies [Marengo, 2004, 2009].
Figure 1-6: Long-term mean annual cycles and spatial distribution of $P$ and $E$ in the Amazon River basin for the period 1982-2008: (a) Long-term annual cycle of $P$ based on GPCC observations and Era-Interim reanalysis; (b) Long-term mean annual cycle of $E$ derived from Era-Interim data, together with a composite based on observations from MPI, MODIS, ORCHIDEE, and GLEAM. Panels (c) and (d) show the spatial distribution of long-term mean annual values of $P$ for Era-Interim and GPCC data, respectively. Panels (e) and (f) show the spatial distribution of long-term mean values of $E$ for Era-Interim and ORCHIDEE data, respectively.

Figure 1-7. Annual cycle of Runoff ($R$) based on HYBAM and Era-Interim data, and Convergence ($C$) based on Era-Interim data.
1.3.3 Long term imbalances between surface-atmospheric water balances

The results reported by Marengo [2004] indicate lower imbalances in the Amazon River basin during La Niña years, which may imply that convergence is over-estimated in episodes of higher than normal precipitation during the dry season, with a surplus of available atmospheric moisture. Thus, one can assume that shortcomings exist regarding the representation of the water balance that may be influenced by the modeling of terms such as precipitable water or the poor representation of evapotranspiration [Guiana et al., 2009; Jasechko et al., 2013], potentially leading to the underestimation of evapotranspiration in the basin.

In order to further study imbalance in the Amazon River basin, we here assume that monthly mean values of the variables involved in the atmospheric balance, and thus all terms in eqn. (1-2), are of equal magnitude [Rasmusson, 1968]. Therefore, the rate of change in the storage of precipitable water \( \frac{\partial W(t)}{\partial t} \) must be included in water balance estimation. By the same token, we also introduce a term to include the rate of change in
soil water storage \( \partial S(t)/\partial t \), such that the corrected imbalance is estimated as

\[
CI(t) = \frac{C(t) - \frac{\partial W(t)}{\partial t}}{R(t) + \frac{\partial S(t)}{\partial t}} - 1 .
\] (1-6)

Monthly means of the corrected imbalance, \( CI(t) \), \( t = 1, 2, \ldots, 12 \), were thus calculated for the entire Amazon River basin using two sets of data: (1) \( C \) (ERA-Interim), \( R \) (HYBAM), \( W \) (ERA-Interim) and \( S \) as the residual of \( P-E-R \) (GPCC-ORCHIDEE-HYBAM), and (2) \( R \) (Era-Interim), as shown in Fig. 6a. This figure reveals a clear pattern of negative residuals (more runoff than moisture convergence) throughout the entire year for both data sets, regardless of runoff source. Monthly mean values of \( CI \) are higher during the peak dry season (August) using reanalysis runoff data, while the observational data set produced higher \( CI \) values throughout most of the year with the exception of August-September. For the entire basin, \( R \) is mainly driven by precipitation and \( C \) is a function of wind speed, surface pressure and specific humidity in the atmospheric column \([\text{Rasmusson, 1968; Peixoto and Oort, 1993; Trenberth and Smith, 2005}]\). The role of water storage (over land and atmosphere) in the basin is of the utmost importance and is related to precipitation recycling and runoff regulation.

According to previous studies, the main sources of moisture for the Amazon are the tropical North Atlantic (TNA) and the migration of the Intertropical Convergence Zone (ITCZ) \([\text{Aceituno, 1988; Marengo, 1992; Fu et al., 2001; Poveda et al., 2006; Nobre et al., 2009; Yoon and Zeng, 2010; Gimeno et al., 2012; Yin et al., 2014; Arias et al., 2015}]\). According to our results, \( C \) is under-represented/calculated by reanalysis (Table 1-1, AWB), leading us to argue that there might be a general under-calculation of the specific humidity and wind velocities driving advection, as modeled in reanalysis, that can lead to lower convergence values in the region. Another plausible explanation for the imbalance may be the misrepresentation of evapotranspiration and subsequent erroneous modeling of precipitation. When comparing the four annual cycles of evapotranspiration in the Amazon (Appendix A, Fig. A7), one can see that all data sets but the Era-Interim set tend to estimate higher levels of evapotranspiration for that particular period. These results indicate a likely underestimation of evapotranspiration in reanalysis that may be related to reduced moisture in the atmosphere and therefore higher imbalance residuals during the dry season.

In the present study we estimated higher values of \( CI \) using \( R \) data obtained from observations, which strongly decrease during August-September. In this particular period we also found that the minimum values of \( C \) and water storage in the basin occur earlier than minimum \( R \) by two months (Fig. 1-7 and Fig. 1-8). Thus far, our results demonstrate the importance of soil and atmospheric water storage at the monthly time scale for the conjoint water balance in the Amazon River basin, as well as supporting the cautionary
caveat when computing monthly water balances as proposed by many other authors [Rao et al., 1996; Curtis and Hastenrath, 1999; Marengo, 2004; Morales-Salinas et al., 2012].

Figure 1-9: Long-term monthly means of the corrected imbalance, $CI$ (eqn. (1-6)), estimated using two sources of $R$ (HYBAM and Era-Interim) for three regions: a) the entire Amazon River basin, b) low-lying portions of the Amazon River basin (Amazonia), and c) the Andes.
1.3.4 The Andes-Amazonia System

In order to further understand the imbalance between the SWB and AWB for the entire Amazon River basin, we then computed the respective water balances over the Andean and low-lying portions of the basin. The Andes comprises those regions located above 500 m a.s.l., while low-lying Amazonia is defined as the areas below the same threshold. Although the former represents only approximately 10% of the total Amazon River basin area [McClain and Naiman, 2008], it contains the regions with the highest values of both rainfall and runoff.

As before, both observational and reanalysis data sets were employed to estimate the SWB and AWB for both regions. Figure 1-9 and Table 1-2 show the calculated long-term mean SWB and AWB variables for Andes and Amazonia. Both regions exhibit a significant imbalance between long-term mean SWB and AWB, ranging from -10% to +11% for Amazonia (greater than the ±5% previously used to define balance) and between -19% and -27% for the Andes. The results for I (row 7 in Table 1-2) reveal a negative imbalance \( R > C \) in both sub-systems, as was also found for the entire basin and in the sub-catchment analysis. For both Andes and Amazonia, agreements in the residuals depending on the used data sets were also recorded, with observational data producing positive residuals \((P-E-R)\) for the SWB and reanalysis data producing negative residuals. This discrepancy may be explained by the fact that observation data reported higher \( P \) and lower \( E \) values, which in turn inverts the sign of the residuals. In both data sets, the values of \( P \) and \( E \) are higher in Amazonia than in the Andes, while observational \( R \) values are higher in the Andes than in Amazonia, both of which patterns are in agreement with the negative residuals in SWB estimated earlier for the sub-catchments (Fig. 1-2).

According to the \( P-E-R \) SWB residuals calculated for both regions, the observational data sets (Table 1-2, row 6) produce a positive residual, likely related to the increase in soil storage in both systems for the period of analysis, while the reanalysis data produce negative residuals for both regions, related to additional water sources besides precipitation. In terms of the AWB, both data sets produce positive residuals \((P-E-C)\), indicating that the two subsystems act as moisture sinks, as found for the entire basin. Fig. 1-10 summarizes the results obtained regarding the variables involved in the SWB and AWB in the Andes and Amazonia regions for the two data sets used. Analysis of the observational results suggests that the Andes portion of the basin acts as a source of surface water that is used in Amazonia to maintain not only runoff but also the evapotranspiration processes. Indeed, evapotranspiration is more intense in the low-lying portion of the basin, as can be seen from the values of \( E \) in each of the regions (Figure 1-9), which in turn favors precipitation and atmospheric moisture convergence.

We also studied the behavior of the annual cycle of residuals for both regions in terms of \( CI \) (Fig. 1-9b and Fig. 1-9c). In both regions the monthly residuals
tend to follow the same pattern of negative values that peak during the dry season, as found for the entire basin (Figure 1-6). Regarding the imbalance residuals obtained using $R$ values derived from reanalysis, both regions reach a maximum negative value in August, of approximately -160% in the Andes and approximately -30% in Amazonia. Similarly, the use of $R$ values derived from HYBAM revealed distinctive behavior in the dry season in both regions, with the Andes reaching a maximum negative imbalance (-100%) and Amazonia a negative maximum (-250%) followed by a positive maximum (+50%) in September.

1.4 Concluding remarks

According to our joint analysis of the SWB and AWB it is evident that there is no such thing as a perfect zero closure of water budgets in the Amazon River basin. Nevertheless, both observation and reanalysis data indicated that the SWB is in fair balance, as the values of closure fell within the range of ±10% of $P$.

Spatial disaggregation of the SWB carried out to study the closure for the main sub-catchments revealed that the results obtained from reanalysis were the opposite to those derived from observations, with negative residuals recorded in the southern portion of the basin. This may be explained by the severe temporal variation in the $P-E$ term, as already reported by [Berrisford et al., 2011] in their analysis of the properties of atmospheric mass conservation in the reanalysis data set. As the performance of reanalysis is sensitive to the spatial disaggregation of data assimilation or the data sources used as input for modeling, we should expect higher uncertainties in sub-catchment closures due to the differences between reanalysis spatial scales and catchment areas.

SWB sub-catchment analysis revealed that the results are independent of the source of $E$, with coherence identified between the closures obtained using observational data sets and spatial patterns. The SWB in sub-catchments showed spatial variation in terms of the sign of the closure, including positive closures in the southern basin and negative closures in the north, indicating that the sub-catchments in the basin do not necessarily exhibit the same closure behavior as the basin as a whole. This is an important result that must be taken into account if the information contained in any one of the data set sources is used for decision-making purposes.
Table 1-2: Comparison of the surface and atmospheric water balances in the two subsystems: Andes (right) and low-lying Amazonia (left). Observational data used reported is the one with the longer period of analysis (1982-2008) with P from GPCC, E from ORCHIDEE and R from HYBAM.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Source</th>
<th>Reanalysis</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Amazonia</strong></td>
<td></td>
<td>mm y(^{-1})</td>
<td></td>
</tr>
<tr>
<td><strong>P</strong></td>
<td>ERA</td>
<td>2271</td>
<td>GPCC 2303</td>
</tr>
<tr>
<td><strong>E</strong></td>
<td>ORCHIDEE</td>
<td>1350</td>
<td>1157</td>
</tr>
<tr>
<td><strong>R</strong></td>
<td>HYBAM</td>
<td>944</td>
<td>931</td>
</tr>
<tr>
<td><strong>C</strong></td>
<td>ERA</td>
<td>841</td>
<td>841</td>
</tr>
<tr>
<td><strong>P-E-R</strong></td>
<td></td>
<td>-23</td>
<td>215</td>
</tr>
<tr>
<td><strong>P-E-C</strong></td>
<td></td>
<td>80</td>
<td>305</td>
</tr>
<tr>
<td>Imbalance, I, eqn. (5)</td>
<td></td>
<td>-11%</td>
<td>-10%</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Variable</th>
<th>Source</th>
<th>Reanalysis</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Andes</strong></td>
<td></td>
<td>mm y(^{-1})</td>
<td></td>
</tr>
<tr>
<td><strong>P</strong></td>
<td>ERA</td>
<td>2043</td>
<td>GPCC 2085</td>
</tr>
<tr>
<td><strong>E</strong></td>
<td>ORCHIDEE</td>
<td>1213</td>
<td>893</td>
</tr>
<tr>
<td><strong>R</strong></td>
<td>HYBAM</td>
<td>929</td>
<td>1034</td>
</tr>
<tr>
<td><strong>C</strong></td>
<td>ERA</td>
<td>753</td>
<td>753</td>
</tr>
<tr>
<td><strong>P-E-R</strong></td>
<td></td>
<td>-100</td>
<td>158</td>
</tr>
<tr>
<td><strong>P-E-C</strong></td>
<td></td>
<td>76</td>
<td>438</td>
</tr>
<tr>
<td>Imbalance, I, eqn. (5)</td>
<td></td>
<td>-19%</td>
<td>-27%</td>
</tr>
</tbody>
</table>

Figure 1-10: Illustration of the components of SWB and AWB in the Andes and Amazonia sub-regions of the Amazon River basin for the observation and reanalysis data sets. Values are in units of mm y\(^{-1}\). The arrow convention in the bottom left represents 1000 mm y\(^{-1}\). Data sources for the variables depicted are as presented in Table 2.
AWB closure exhibited high variability, ranging between 4% - 15% of $P$. No atmospheric balance was found in the Amazon basin for any of the used data sets, with higher closure values recorded for those data sets containing lower values of $E$. This result demonstrates the shortcomings of reanalysis when computing $C$, as well as hinting at a potential compensation of the budget when increasing the value of $E$. Spatial agreement among all data sets was observed in the distribution and sign (positive) of closure in the sub-catchments, although this may have been influenced by the fact that we used the same data source for $C$.

According to the presented analysis, we conclude the existence of imbalance in both of the branches of the water cycle in the Amazon River basin, driven by $R$. Although the imbalances found here of between 14% and 16% are lower than those reported in previous studies [Zeng, 1999; Marengo, 2004], our results also indicate large differences between the amounts of water on the surface and in the atmosphere in the Amazon River basin. These imbalances may be related to both an underestimation of $C$ and the fact that fluxes are not accounted for or are poorly represented in reanalysis. The observed spatial distribution of imbalances among the sub-catchments in the basin is similar to the spatial distribution of AWB closures, with positive AWB closure and negative imbalance. These sub-catchments are those in which reanalysis tends to produce higher values of $C$ that cannot be balanced with the other variables in the budget, and are also characterized by disagreements between observations and reanalysis with respect to SWB closure.

One of the main contributions of the present work is the correction of the imbalance equation, which enabled us to perform an analysis of the annual cycle of imbalance between SWB and AWB. This analysis revealed a common feature of negative imbalances during the year, although the Amazon dry season was associated with higher negative values. The results obtained using the observational data sets also exhibited an abrupt change in the imbalance during August and September that was not present in the reanalysis results. Low values of $R$ measured during the months of August and September were almost in balance with $C$ and the water storage terms, indicating that, for the Amazon’s driest months, uncertainties are lower regarding the modeling of $C$ but higher regarding the modeling of $R$.

Disaggregation of the Andes and Amazonia sub-regions’ imbalance results produced the same negative pattern as found for the entire basin during the year. Amazonia presented a similar abrupt change in imbalance as observed for the whole basin, with positive values of imbalance (+50%) recorded in September. For the Andes, imbalance during the year was always negative, with minimum values recorded during the dry season for both reanalysis and observational data sets. The minimum imbalance in the Andes was also higher for reanalysis data, although the peaks of negative imbalance were in phase for both data sets in August. Our results suggest that reanalysis produces an incomplete representation of surface processes in the low-lying Amazonia sub-
region, such as the influence of the forest and surface ecosystems in the regulation of R.

Imbalance values were higher over the Andes for both data sets. Such results were expected, as this particular region poses the greatest challenge not only for modeling due to the high altitudes, but also due to the scarcity of observations made in the highlands of the Amazon River basin. In general, the imbalances recorded for the Andes sub-region were consistent between data sets, characterized by higher negative values of imbalance in the dry season. During this season of lower P the Andes sub-catchments are associated with higher values of R than of C, confirming that the Andean slopes are the main suppliers of surface water for the basin.
2. Seasonal and Interannual Two-Way Feedbacks between Tropical North Atlantic SST and Amazonian Hydroclimatology

Alejandro Builes-Jaramillo, Germán Poveda and Paola Arias

Abstract: The Amazon River basin (AM) is considered among the most important tipping points of Earth’s climate system. Large-scale macro-climatic phenomena and teleconnections from the tropical Pacific and Atlantic Oceans exert a strong influence on diverse hydroclimatic processes in AM at seasonal and interannual timescales, and the latter feedback on atmospheric and oceanic processes over the Tropical North Atlantic (TNA). Here, we provide statistical and physical evidences of the existence of two-way interactions and a dynamical coupling between sea surface temperatures (SST) on the TNA and the hydro-climatology of AM at seasonal and interannual timescales. At seasonal and interannual timescales the feedback processes act as follows: (a) from the TNA to AM, there is AM uptake of moisture from TNA mainly driven by the TNA SSTs and the zonal winds directed towards the continent with a 4 to 5 months lag in the seasonal timescale and (b) from the AM to the TNA convection anomalies over the AM disrupt the pressure gradient between the two regions thus influencing the zonal winds in TNA and therefore changing the patterns of evaporative cooling in the ocean. At seasonal timescales the feedback process from AM to TNA develops in a time-lapse from 0 to 3 months lag, while for interannual timescales the feedback develops from 0 to 1 months lag. Our results provide clear-cut evidences that the hydrology of the AM exerts a strong influence on the regional climate at both timescales.

Keywords: Amazon River basin, Ocean-Atmosphere Interactions, Hydroclimatological Feedbacks, Tropical North Atlantic, Sea Surface Temperatures

2.1 Introduction

The Amazon River basin (AM) is considered among the most critical tipping points of Earth’s climate system [Lenton et al., 2008]. At seasonal timescales, the hydroclimatic variability of Amazonia is influenced by sea surface temperatures (SST) over the tropical Atlantic and tropical Pacific [Aceituno, 1988; Marengo, 1992; Liebmann and Marengo, 2001; Marengo et al., 2002; Souza et al., 2004; Poveda, 2004; Garreaud et al., 2009; Nobre et al., 2009; Yoon and Zeng, 2010; Gimeno et al., 2012; Satyamurty et al., 2012; Yin et al., 2014; Martins et al., 2015], and in turn the Tropical Atlantic variability may be intensified by processes such the Atlantic Multidecadal Oscillation
The AM hydroclimatic variability is also controlled by the dynamics of the Intertropical and South Atlantic Convergence Zones ([Lenters and Cook, 1999; Paegle and Mo, 2002; Espinoza et al., 2012; Sena et al., 2012; Boers et al., 2013; de Oliveira Vieira et al., 2013], the South American Monsoon System and the South American Low Level Jet [Marengo et al., 2004; Vera et al., 2006; Silva et al., 2009; Boers et al., 2015], by the dynamics of aerial rivers and lakes [Moraes-Arraut et al., 2011; Poveda et al., 2014], by the intraseasonal oscillation [Drumond and Ambrizzi, 2006; Rodrigues et al., 2011; Shimizu et al., 2016] and by land surface-atmosphere feedbacks that involve the interactions between soil moisture and vegetation, and the recycling of precipitation via evapotranspiration [Lettau et al., 1979; Eltahir and Bras, 1994; Poveda and Mesa, 1997; Salazar et al., 2007, 2016; Nepstad et al., 2008; Drumond et al., 2014; Zemp et al., 2017, 2014; Rocha et al., 2015; Mallick et al., 2016].

But hydroclimatic processes in AM are not just passive spectators of climate variability at continental scales. Observational, statistical and modeling studies have shown evidences that AM hydro-climatology affect diverse processes such as precipitation in the lower U.S. Midwest during the spring and summer seasons [Werth and Avissar, 2002]. Modeling studies have shown that the hydro-climatological dynamics of Amazonia affect climate and weather patterns at continental and global scales [Drumond et al., 2014]. Also, anomalous convection processes in AM have shown to be influential in the Inter-Americas Seas (IAS) SSTs [Misra and DiNapoli, 2012], and also in changing the displacement patterns of the Intertropical Convergence Zone [Wang and Fu, 2007a].

At interannual timescales, El Niño/Southern Oscillation (ENSO) is the main modulator of AM hydroclimatology owing to teleconnections arising from the tropical Pacific and Atlantic [Aceituno, 1988, 1989; Marengo et al., 2002; Souza et al., 2004; Yoon and Zeng, 2010; Valverde and Marengo, 2011; van der Ent and Savenije, 2013; Yin et al., 2014; Arias et al., 2015; Marengo and Espinoza, 2016], as well as land surface-atmosphere feedbacks [Costa and Foley, 1997; Poveda and Mesa, 1997; Laurance and Bruce Williamson, 2001; Betts et al., 2004; Poveda et al., 2006; Nobre et al., 2009]. But few studies have focused on the possible effects of Amazonian hydro-climatology on oceanic-atmospheric processes over the Tropical North Atlantic (TNA) at interannual timescales [Poveda and Mesa, 1997; Wang and Fu, 2007a].

The space-time dynamics of hydro-climatological processes in the Amazon River basin depend on large-scale global and continental atmospheric circulation patterns [Nobre et al., 2009], but also on land surface-atmosphere bio-geophysical feedbacks occurring at local, regional and continental spatial scales [Poveda et al., 2006]. The latter processes involve soil moisture storage, vegetation activity, evapotranspiration and recycled precipitation [Eltahir and Bras, 1994; Zemp et al., 2014, 2017; Rocha et al., 2015]. Limitations about the physical space-time dynamics of tropical convection and
rainfall [Gimeno et al., 2012], as well as actual and potential evapotranspiration and soil moisture hamper both modeling and forecasting the hydro-climatic dynamics of Amazonia using general and regional circulation models [Harper et al., 2010]. Therefore, the possible feedbacks between the Amazon hydrology and the regional climate have been underestimated or overlooked. There are two major oceanic forcing mechanisms of the AM hydro-climatology at seasonal timescales, namely SST over both the tropical Pacific and Atlantic oceans [Satyamurty et al., 2012]. These oceanic regions influence hydrological processes over AM (moisture transport by the winds and precipitation, which in turn affect soil moisture, evapotranspiration, and river discharges) at seasonal timescales [Andreoli et al., 2012].

At interannual timescales, the extreme phases of ENSO are strongly associated with precipitation anomalies [Marengo and Hastenrath, 1993; Poveda and Mesa, 1997; Nobre et al., 2009; Marengo and Espinoza, 2016], and many other AM hydro-ecological processes [Poveda and Salazar, 2004; Ramos da Silva et al., 2008]. The anomalously severe droughts of 2005 and 2010 in Amazonia have mainly been associated with SST anomalies, the former over the TNA region and the latter explained by the occurrence of warm phase of ENSO (El Niño) in the tropical Pacific [Marengo et al., 2011; Coelho et al., 2012; Marengo and Espinoza, 2016]. More recently, evidences have been shown about the role of deforestation for the advection of moisture from TNA to AM in the wet season, and in the development of the South American Monsoon System due to reduced moisture and energy fluxes in the forest [Boers et al., 2017]. On the other hand, the study by Poveda and Mesa, [1997] put forward the theory that AM plays the role of a “land-atmosphere bridge” connecting the tropical Pacific and TNA SSTs during El Niño events, supported in high and extensive lagged cross-correlations between the first Principal Component of monthly standardized anomalies of streamflows at 88 rivers in Colombia and TNA SSTs monthly standardized anomalies, peaking when the hydrological processes over the AM lead the TNA SSTs by 4-5 months later. Building on that previous study, this work aims to further investigate the existence of two-way feedbacks between TNA SSTs and AM hydrological processes at seasonal and interannual timescales.

The present work is organized as follows. Section 2 outlines the data and methodologies used. Section 3 provides the results regarding feedbacks at seasonal time scales (3.1), interannual time scales (3.2), the spatial extents of the feedbacks (3.3) and the physical evidence of feedbacks (3.4). Section 4 presents the concluding remarks and the implications of the results.

2.2 Data and Methods

We defined two regions, the Tropical North Atlantic (TNA) located between 75°W to 10°W and 5°S to 29°N, and the Amazon River basin (AM) located between 79.5°W to 50.5°W and 19.5°S to 4.5°N (Figure 2-1). We represent
the AM hydro-climatology through the monthly evaporation index over the entire river basin, $P-E$, where $P$ denotes precipitation and $E$ denotes actual evapotranspiration. Monthly precipitation data was obtained from the Global Precipitation Climatology Centre (GPCC) produced by the Deutscher Wetterdienst (gpcc.dwd.de) at 1° x 1° spatial resolution [Schneider et al., 2014]. Evapotranspiration information was obtained from the Land Surface Model (LSM) ORCHIDEE [Krinner et al., 2005] data set that involves a hydrological model (SECHIBA) and a carbon model (STOMATE) forced with the NCEP/NCAR reanalysis, and with the ORE-HYBAM project precipitation (http://www.ore-hybam.org/); the dataset has a spatial resolution of 1.0° x 1.0° and a monthly time resolution. The monthly $P-E$ index was calculated for the period 1979-2009 that was common to both datasets. SST data was obtained from the third version of the Extended Reconstructed Sea Surface Temperature data set (ERSST v3b, http://www.ncdc.noaa.gov/data-access), produced by the U.S. National Oceanic and Atmospheric Administration (NOAA). The SST data used has a spatial resolution of 2.0° x 2.0° and a monthly time resolution spanning from 1854 to 2014 [Smith et al., 2008].

![Figure 2-1](image)

Figure 2-1: Location of the regions under study, with the Amazon River basin (AM) in green, and the Tropical North Atlantic (TNA) in blue

For the analysis of the seasonal and interannual timescales, the variables $P-E$ and SST were averaged over the defined continental and oceanic regions with the purpose of calculating monthly series and monthly anomalies series. Lagged cross-correlations (CC) were calculated between them for both negative lags (TNA leading AM) and positive lags (AM leading TNA), and a null hypothesis of zero correlation between time series was used in order to construct 95% confidence intervals. Iso-correlation maps were calculated via cross-correlation analysis between the time series of averaged $P-E$ values over the AM region, and each time series of SST over the TNA region. In order to test for the presence of the two-way feedback mechanisms at interannual timescales, we applied a 13-month running mean low-pass filter to all datasets and then calculated the mean monthly values over the studied regions, and estimated the aforementioned CC and the iso-correlation analysis.
To the end of understanding suitable physical mechanisms driving the identified feedbacks we performed a complementary CC analysis, using monthly mean and monthly anomaly time series of surface pressure over the AM and TNA regions, with information obtained from the ERA-Interim Reanalysis [Dee et al., 2011], and the pressure gradient between both regions was calculated as the difference between the averaged surface pressures in the TNA minus the averaged surface pressures in the AM region (TNA-AM). On the other hand, interpolated estimates of Outgoing Longwave Radiation (OLR) were used with the purpose of testing further evidences of the relation between precipitation in the AM and the surface pressure in TNA. We used OLR data provided by the NOAA/OAR/ESRL PSD, in Boulder, Colorado, USA, from their Web site at http://www.esrl.noaa.gov/psd/. The dataset has a monthly time resolution from 1974 to 2013 and a spatial resolution of 2.5° x 2.5°[Liebman and Smith, 1996].

Besides the statistical analyses, we also searched for physical evidences supporting the existence of the two-way feedbacks by plotting composites and maps showing the dynamics of seasonal and monthly anomalies of the variables that may play a key role in the proposed feedback mechanisms. Zonal wind velocities were obtained from the ERA-Interim and total wind vector from the NCEP/NCAR reanalysis, sea surface height from NCEP Global Ocean Data Assimilation System (GODAS) provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, (http://www.esrl.noaa.gov/psd/), and average upper ocean heat content from 0m-400m computed form the Met Office Hadley Centre [Gouretski and Reseghetti, 2010]. We also computed vertically integrated horizontal water vapor transport anomalies (IVT) as proposed by [Lavers et al., 2012; Gimeno et al., 2016] with data form the ERA-Interim reanalysis.

2.3 Results and Discussion

2.3.1 Feedbacks at seasonal time scales

The estimated cross-correlation (CC) shown in Figure 2-2a show significant positive cross-correlations for increasing lags, peaking at -5-month lag (r = 0.85, P > 0.95) when TNA SSTs lead AM hydrology. This observation implies that higher TNA SST will lead to higher AM water availability (P-E and runoff) 4-5 months later. A potential explanation is that higher TNA SSTs lead to higher evaporation and water vapor export from the TNA, which, via the trade winds, contributes to increase water availability (P-E) over AM, peaking 4-5 months later. This result and interpretation is consistent with the long-established fact that the TNA is the most important source of moisture for the Amazon River basin [Hastenrath and Heller, 1977; Marengo, 1992; Curtis and Hastenrath, 1999; Nobre et al., 2009; Moraes-Arraut et al., 2011; Gimeno et al., 2012; Satyamurty et al., 2012; Drumond et al., 2014; Poveda et al., 2014; Arias et al., 2015].
On the other hand, Figure 2-2a also shows a strong peak of negative cross-correlations when AM hydrology leads TNA SSTs peaking at 1 month-lag ($r = -0.87$, $P > 0.95$), suggesting that a decrease in precipitation over land might be related to higher TNA SSTs one month later, as the P-E index is primarily influenced by precipitation over AM (see Figure B1 in the Appendix B). Figure 2-2b shows the long-term mean annual cycles of the TNA SST and AM P-E index. It is clear that the P-E index reaches a minimum value in August; one month earlier than the TNA SSTs reaches its maximum in September. Figure 1c contains the long-term mean annual cycles of the P-E index and TNA-AM surface pressure gradient. It evidences that the decrease of the P-E index, starting in March, could be related also with a reduction in the surface pressure gradient between the TNA and AM, and reaching a minimum value in August. Also, the maximum value of the TNA-AM pressure gradient during January leads the maximum P-E index in February and March. Besides, a reduced pressure gradient between both regions (owing to an increase in AM surface pressure) may explain a lower P-E index, in agreement with lower TNA SSTs.

We also estimated the CC between monthly series of the surface atmospheric pressure gradient between TNA and AM (TNA-AM) and the monthly mean values of the P-E index (Figure 2-2d). Our results indicate that the surface pressure gradient lead P-E with positive correlation reaching a maximum peak at one month-lag ($r = 0.77$, $P > 0.95$). Figure 2-2d also shows that the P-E index leads the pressure gradient with a one month-lag ($r = 0.43$, $P > 0.95$), suggesting that the reduction (increase) in precipitation over AM is therefore related with the reduction (increase) of the pressure gradient. The one month delay in the AM to TNA feedback was found in both CC analyses and can be more easily identified in the August to September period of the annual cycle plots (see Figure 2-2c). The above result evidences that low condensational processes at the AM during the period from August to September, may be related with a decrease in the TNA-AM surface atmospheric pressure gradient during the same period. These conditions are sufficient to trigger the feedback mechanism by which the zonal trade wind velocities towards the Amazon decrease due to a reduction of the atmospheric pressure gradient (see Figure B2 in Appendix). A reduced pressure gradient is related with less steep isobars, causing the reduction of the zonal wind speed and therefore a less active evaporative cooling on the TNA region, and the concomitant increase in SSTs (see Figure B3 in Appendix). These results provide strong indications as to how the hydrological phenomena occurring over the continent (AM) may feedback on the TNA oceanic phenomena at seasonal timescales.
Figure 2-2: a) Lagged cross-correlations between time series of monthly mean values of the P-E index over Amazonia and TNA SST, blue dashed lines represent the 95% confidence bounds. b) Mean annual cycle of the AM P-E index (mm/month) and the TNA SST’s (°C). c) Mean annual cycle of the TNA-AM surface pressure gradient (hPa) and the SSTs (°C) at the TNA region. d) Lagged cross-correlations between the series of the monthly mean values of the surface atmospheric pressure gradient between the TNA and AM regions.
(TNA-AM) and the monthly mean values of the AM $P$-$E$ index, blue dashed lines represent the 95% confidence bounds.

### 2.3.2 Feedbacks at the interannual time scale

We also investigated the existence of two-way feedback mechanisms at interannual timescales through CC analysis; by removing the annual cycle in the relevant monthly time series (see Methods). Fig. 2-3a shows the lagged CC between monthly anomalies of TNA SSTs and $P$-$E$ over Amazonia. There is a significant negative correlation ($r = -0.47$, $P > 0.95$) when the $P$-$E$ index leads the TNA SST by 1 month, suggesting that an interannual negative (positive) anomaly in $P$-$E$ is associated with an increase (decrease) in TNA SSTs with a 1-to-2 months delay. These results reflect the influence of the combined effects of the interannual anomalies in land-surface hydrological processes over the Amazon River basin TNA SSTs [Poveda and Mesa, 1997; Wang and Fu, 2007b; Angelini et al., 2011; Zemp et al., 2014]. The mechanisms by which AM hydrology may affect TNA SSTs at interannual timescales are configured by the interaction between surface pressure, precipitation, air temperature, evapotranspiration, soil moisture and river runoff [Poveda et al., 2006].

The negative cross-correlations found in the CC analysis (Figure 2-3a) suggest that higher TNA SSTs anomalies are associated with a decrease in the AM $P$-$E$ index at interannual timescales, thus meaning a drier land resulting from the differential land-ocean heating. Once water availability over land starts decreasing (May-October), land-driven processes over the continent (precipitation, water recycling, vegetation activity, and runoff) combine to reinforce the positive feedback. The intensification (weakening) of hydrological processes ($P$-$E$ index) over the AM reaching a strong peak at one month lag leads to a gradual increase (decrease) of TNA SSTs that may be related with the release of latent heat from condensation in AM that propagates by the atmosphere to the ocean through equatorial Kelvin waves [Wang and Fu, 2007b].

With the aim of explaining the hydro-climate dynamics involved in our previous statistical results at interannual timescales, we evaluated the relationship between Outgoing Longwave Radiation [Liebman and Smith, 1996] ($OLR$) over the Amazon River basin, and Sea Level Pressures (SLP) over the TNA, after removing the annual cycle. Figure 2-3b shows that AM $OLR$ anomalies lead those of TNA SLP by five months with a negative correlation ($r = -0.37$, $P > 0.95$). This result suggests that an AM $OLR$ positive (negative) anomaly (lower OLR values associated with higher rainfall rates) leads to a negative (positive) SLP TNA anomaly. This result lends support to the idea that a drier (moister) Amazon leads to warmer (colder) TNA SST at interannual timescales.
Figure 2-3: a) Lagged cross-correlation analysis between the low-pass filtered anomalies of the P-E index in Amazonia and those of the TNA SST. b) Lagged-cross correlation analysis between low-pass filtered anomalies of OLR in the Amazon region and those of surface atmospheric pressure in the TNA region. Blue dashed lines represent the 95% confidence bounds.

2.3.3 Spatial extent of the Amazon influence on TNA

Now we focus on the spatial dynamics of the identified two-way feedback mechanisms. To that end, we computed two sets of correlation maps from 6 to -6 month lags deployed over the TNA region. The first set (Fig. 2-4a) is a composite of lagged CC between time series of monthly P-E values averaged over the AM region and TNA SSTs time series. The second set (Fig. 2-4b) is a composite between monthly anomaly time series of the AM P-E index and the TNA SSTs time series, following removal of the annual cycle. In Figure 2-4a (seasonal) there are two maps showing that approximately 65% of the TNA region returns high correlation values (0.75 ≤ r ≤ 1.0, P > 0.95; Fig. 2-5), one at lag +1 (negative correlation) when the AM hydrology leads TNA SSTs, and the other at lag -5 (positive correlation) when TNA leads AM hydrology. At interannual timescales (Fig. 2-4b) we found a map at lag +1 month in which approximately 45% of the TNA region reports high correlation values (0.50 ≤ r ≤ -0.75; Fig. 2-5) when the AM hydrology leads the TNA SST with a positive correlation. These iso-correlation maps lend strong statistical support to the identified two-way feedback mechanisms between the two regions at both seasonal and interannual timescales.
Figure 2-4: a) Lagged cross-correlation maps between the mean monthly values of AM P-E and TNA SST, for ±6 month lags. b) Lagged cross-correlation maps between monthly anomalies (low-pass filtered) of AM P-E and TNA SST time series, for ±6 month lags. Solid black lines represent zones with the same values of correlation coefficient (iso-correlation zones).
44 Seasonal and Interannual two-way feedbacks

2.3.4 Physical evidences of feedbacks

So far, the statistical analysis has given us hints on the possible feedbacks between AM hydrometeorology and TNA SSTs. We have related P-E in the AM with changes in the pressure gradient between AM and TNA and thus, in the present section we will explore the dynamical behavior of diverse processes over the TNA that may respond to changes in the pressure gradient. Figure 2-6 shows a schematic representation of the response variables in atmosphere and surface in TNA.
denoted as \( P \), the surface pressure in both regions as \( SP \), and the TNA SSTs as \( S \). Arrows directed towards the Amazon River basin shows the atmospheric response in zonal winds anomalies (\( ZW \)), and vertically integrated horizontal water vapor transport anomalies (\( IVT \)). Upwards arrows over the TNA illustrate the response variables in oceanic sea surface height (\( SSH \)) and upper heat content (\( UOH \)).

### 2.3.4.1 Physical mechanisms at play at seasonal timescales

With the purpose of evaluating the dynamics of diverse processes over the TNA during contrasting seasons over the AM, we defined the wet season from November to January (NDJF), and the dry season from July to October (JASO). Figure 2-7 shows composites of SSTs and zonal wind seasonal anomalies in the TNA. During the AM dry season (Figure 2-7a) there is a reduction of zonal wind velocities in July over the central TNA region between 2°N and 10°N, even reaching 20°N near the African coast, and SSTs are colder near South America. By August the slower zonal winds flowing towards the AM cover a broader band north of 2°N up to 15°N, and SSTs above the equator are 1°C higher than those found in July. While the reduction of the winds appears to be in phase with the beginning of the dry season, increasing temperatures in the ocean are found to peak after three months. During the AM wet season (Figure 2-7b), starting in November, conditions are similar to those at the end of the dry season (Figure 2-7a), with a warming of almost 2°C above the mean conditions in the northern region of the TNA, and slower zonal wind velocities near the coasts of South America from 5°S to 5°N. In the peak of the South American Monsoon, from December to February, zonal winds start to increase their velocities towards the continent to the north of 5°N up to 15°N as for the dry season, and SSTs become colder reaching values even 3°C below normal conditions.

Figure 2-8 shows composites seasonal anomalies in zonal wind and vertically integrated horizontal water vapour transport (\( IVT \)) in the TNA. During the AM dry season (Figure 2-8a) the TNA exhibits two distinctive patterns, one near northern South America and northeast Brazil from 0°N to 10°N, which is characterized by a reduction of \( IVT \) from July to September, reaching 150 kg m\(^{-1}\) s\(^{-1}\). The second region is above 10°N and presents positive anomalies of \( IVT \) over the Caribbean Sea. During the AM wet season (Figure 2-8b), starting in November there are almost normal conditions with positive anomalies of no more than 3 kg m\(^{-1}\) s\(^{-1}\) located in northeast Brazil. In December, \( IVT \) positive anomalies are found in a region from 5°S to 10°N reaching up to 100 kg m\(^{-1}\) s\(^{-1}\). And during January-February, positive anomalies are located between 0°N and 5°N, with values of approximately 50 kg m\(^{-1}\) s\(^{-1}\). During the wet season, the TNA region above 10°N experiences negative anomalies of \( IVT \). According to these composite analyses, the lower atmosphere response over TNA is characterized by positive (negative) zonal wind anomalies and by negative (positive) anomalies of \( IVT \) in AM dry (wet) season.
Figure 2-7: Composites of SST and Zonal Wind seasonal anomalies in the TNA region for the AM dry season (a) and AM wet season (b). In color are denoted the SST anomalies, in dashed lines the negative zonal wind anomalies, and in solid lines positive zonal wind anomalies.

Figure 2-8: Composites of Zonal Wind and Vertically integrated horizontal water vapour transport seasonal anomalies in the TNA region for the AM dry season (a) and AM wet season (b). In color are denoted the IVT anomalies, in dashed lines the negative zonal wind anomalies, and in solid lines positive zonal wind anomalies.
We have constructed the Hövmoller (longitude-month) diagrams shown in Figure 2-9. We evaluated the response of upper ocean content (UOH) and sea surface height (SSH) in five different regions of the TNA i) -5°S to 0°, ii) 0° to 5°N, iii) 5°N to 10°N, iv) 10°N to 15°N and v) 15°N to 30°N (Figure 9a and Figure 9b). Both UOH and SSH exhibits negative anomalies from January to June in the region from 5°S to 5°N, in association with the wet season in the AM, and positive anomalies from July to December in the same region, in association with the dry season in the AM. Values of UOH in the region from 5°S to 0° also exhibit a similar behavior to the one described above, but with negative anomalies moving farther east from January to September. The dynamical analysis has shown that changes in the AM precipitation patterns are related with a concomitant change in the SSTs over the TNA. At seasonal timescales, wet (dry) seasons in the Amazon are related with the increase (reduction) of zonal winds directed towards the continent, and therefore with a reduction (increase) of SSTs in a time lapse of around three months.

![Longitude-month diagrams](image)

Figure 2-9: Longitude-month diagrams of (a) 0-400m upper ocean heat content in Jm-2, and (b) sea surface height in m. The five diagrams for each variable represent one region in the TNA: i) -5°S to 0°, ii) 0° to 5°N, iii) 5°N to 10°N, iv) 10°N to 15°N and v) 15°N to 30°N.
2.3.4.2 **Physical mechanisms at play at interannual timescales**

At seasonal timescales, the variability of TNA SSTs is strongly forced by the annual cycle, and thus we cannot conclude that the response found in the TNA is solely due to the proposed mechanism. Thus, aiming for a more comprehensive understanding of the feedback mechanisms at the interannual timescale we calculated monthly anomalies in order to perform a dynamical analysis of total winds and SSTs over the TNA during extreme events of convection in the AM. The rationale behind the extreme events analysis is to search for indications of the onset and development of the mechanism after or during a convection anomaly in the AM. For this purpose, we have selected the extreme events of January 1999 with mean precipitation 8% higher than the monthly average, January of 2005 with mean precipitation 15% lower than the monthly average (235 mm month\(^{-1}\)).

Figure 2-10 shows that during the two months previous to January 1999 the total winds are directed westerly and are mainly concentrated in the Caribbean with velocities reaching up to 4.5 m s\(^{-1}\) in December 1998. The SSTs during November-December 1998 exhibit positive anomalies ranging from 0.3°C to 0.8°C. During January 1999, when the mean anomalies of precipitation over AM reached +20 mm month\(^{-1}\) (8% higher than mean monthly values of 235 mm month\(^{-1}\)) the northwest AM was experiencing higher precipitation anomalies and the westerly winds are found from 5°N to 30°N. During January 1999 total winds were increased from 2 m s\(^{-1}\) to 4 m s\(^{-1}\) in the region from 10°N to 30°N, and around 1 m s\(^{-1}\) to 1.5 m s\(^{-1}\) in the region from 0° to 10°N, flowing towards the continent. Figure 2-10 also shows a decrease in SSTs reaching negative anomalies of -0.4°C in the region between 5°N to 25°N and -50°E to -15°E. During the period February-March 1999, just after the precipitation maximum, total winds are not directed towards the AM, and SSTs maintain negative anomalies, in particular near the coast of Africa.

Figure 2-11 shows that during the two months previous to January 2005 total winds are directed towards the east, and that SSTs exhibit positive anomalies with maximum values of 1.1°C covering most of the TNA. These high temperatures in the TNA have been reported as part of an ongoing warming of the TNA that was particularly strong between 2004 and 2005, which was related with the once-in-a-century drought of the AM in 2005 [Marengo and Espinoza, 2016]. During January of 2005, when the mean anomalies of precipitation in AM reached -35 mm month\(^{-1}\) (15% lower than mean monthly values of 235 mm month\(^{-1}\)) the northwest and central regions of AM were experiencing the lower values of precipitation, and the total winds were reduced in the westerly direction from 10°N to 20°N, Fig. 2-11 also shows that total winds do not exhibit a clear-cut defined direction from 0°N to 20°N and that they reached their lower speeds between 5°N and 20°N. There is an
increase in SSTs that reaches a maximum value of 1.3°C on the region between 5°N to 20°N. During the two months after the extreme decrease in precipitation in February-March 2005, total winds do not exhibit clear direction towards the AM and TNA SSTs experience positive anomalies that reach values of 1.6°C near the coasts of northern South America.

Our previous results show that during the AM extreme precipitation events of January 1999 (wet) and 2005 (dry), there is a pattern of reduced (increased) SSTs anomalies over the TNA that can be observed after the increase (reduction) of convection in the AM. The progression of cooling (warming) in the TNA SSTs is therefore in the range of 0-2 months as found with the interannual lagged cross-correlations. Similar conclusions can be drawn from other extreme events of precipitation in the AM in September 1999, June 2005, January 2009 and March 2010 (see Figures B4 to B8 in Appendix).
Seasonal and Interannual two-way feedbacks

Figure 2-10: Time evolution of total wind anomalies and SSTs two months before and two months after extreme precipitation in the Amazon in January 1999. The first row shows extreme precipitation anomalies in January of 1999, the middle row shows total wind anomalies from November 1998 to March 1999, and the bottom row shows SSTs anomalies from November 1998 to March 1999.
Figure 2-11: Time evolution of total wind anomalies and SSTs two months before and two months after the extreme precipitation in the Amazon in January 2005. The first row shows extreme precipitation anomalies in January of 2005, the middle row shows total wind anomalies from November 2004 to March 2005, and the bottom row shows SSTs anomalies from November 2004 to March 2005.
2.4 Concluding remarks

Our results show that the anomalies in atmospheric surface pressure gradient between TNA and AM modulate the strength and direction of the zonal trade winds flowing from the TNA towards the AM, thus influencing its $P-E$ index at seasonal timescales. They also indicate that there is an influence of AM hydrology on the atmospheric surface gradient between both regions that in turn can modulate zonal winds in the TNA and therefore TNA SSTs at seasonal and interannual timescales. The impact of AM hydrological anomalies at interannual timescales on TNA SST anomalies (0 to 2 months ahead) could be explained in terms of equatorial waves generated in the AM, which are transported in the form of precipitation that originates in AM and propagate towards the ocean [Wang and Fu, 2007b].

Strong evaporative and convective processes are intensified in the Amazon by the action of the tropical forests where increased water availability may be related to the positive feedback of precipitation recycling. Therefore, ENSO-driven perturbations can be propagated towards the TNA by the direct influence of the AM hydrology, at one month lag described in the present study, and in agreement with a pretty clear teleconnection between Equatorial Amazonia and Caribbean SST anomalies [Misra and DiNapoli, 2012].

Our results evidence that hydro-ecological processes within the AM play an important role in subsequent TNA ocean-atmospheric dynamics, and highlights the possible consequences of deforestation and land use changes on the integrity of the TNA-AM coupled system. Our results provide evidence that the AM hydrology consistently influences the TNA SSTs from 0 to 2 months ahead at seasonal and interannual time scales, potentially affecting more than 60% of the oceanic region, thus lending further support to the theory that Amazonia indeed acts as a “land-atmosphere bridge” linking Pacific and Atlantic SSTs anomalies, put forward two decades ago by [Poveda and Mesa, 1997].

In deforested Amazon [Nepstad et al., 2008] there could be a significant reduction in precipitation and evapotranspiration and an increase in albedo that could reinforce the identified feedbacks towards a warmer TNA. A continuous warming of TNA SSTs would affect the dynamics of tropical easterly waves and tropical storms, and very likely would induce more and more intense hurricanes and positive feedbacks into the land, thus accelerating desiccation and vegetation dieback, which in turn could lead to a tipping point in the Amazon hydroclimate, of potential consequences at continental and global scales.
3. Nonlinear interactions between the Amazon River basin and the Tropical North Atlantic at interannual timescales

Alejandro Builes-Jaramillo, Norbert Marwan, Germán Poveda and Jürgen Kurths

Abstract: We study the physical processes involved in the potential influence of Amazon (AM) hydroclimatology over the Tropical North Atlantic (TNA) Sea Surface Temperatures (SST) at interannual timescales, by analyzing time series of the precipitation index (P-E) over AM, as well as the surface atmospheric pressure gradient between both regions, and TNA SSTs. We use a recurrence joint probability based analysis that accounts for the lagged nonlinear dependency between time series, which also allows quantifying the statistical significance, based on a twin surrogates technique of the recurrence analysis. By means of such nonlinear dependence analysis we find that at interannual timescales AM hydrology influences future states of the TNA SSTs from 0 to 2 months later with a 90% to 95% statistical confidence. It also unveils the existence of two-way feedback mechanisms between the variables involved in the processes: (i) precipitation over AM leads the atmospheric pressure gradient between TNA and AM from 0 and 2 month lags, (ii) the pressure gradient leads the trade zonal winds over the TNA from 0 to 3 months and from 7 to 12 months, (iii) the zonal winds lead the SSTs from 0 to 3 months, and (iv) the SSTs lead precipitation over AM by 1 month lag. The analyses were made for time series spanning from 1979 to 2008, and for extreme precipitation events in the AM during the years 1999, 2005, 2009 and 2010. We also evaluated the monthly mean conditions of the relevant variables during the extreme AM droughts of 1963, 1980, 1983, 1997, 1998, 2005, and 2010, and also during the floods of 1989, 1999, and 2009. Our results confirm that the Amazon River basin acts as a land surface-atmosphere bridge that links the Tropical Pacific and TNA SSTs at interannual timescales. The identified mutual interactions between TNA and AM are of paramount importance for a deeper understanding of AM hydroclimatology but also of a suite of oceanic and atmospheric phenomena over the TNA, including recently observed trends in SSTs, as well as future occurrences and impacts on tropical storms and hurricanes throughout the TNA region, but also on fires, droughts, deforestation and dieback of the tropical rain forest of the Amazon River basin.

Key words: Nonlinear processes, Amazonia, Tropical North Atlantic, Hydroclimatology, SST
3.1 Introduction

Understanding the space-time dynamics of sea surface temperatures (SST) over the Tropical North Atlantic (TNA) is utterly important, given that this particular region has the potential to strongly modulate tropical and extra tropical climates [Avissar and Werth, 2004]. There is evidence that the TNA SSTs play a key role in the modulation of extreme hydrometeorological events in the boreal summer, such as droughts in the United States and heat waves in Europe [Sutton and Hodson, 2005]. Tropical Atlantic SSTs above 26°C in the TNA are one of the necessary conditions for the development of tropical storms and hurricanes reaching the Caribbean Sea, Central America, the US and the mid-latitude North Atlantic [Goldenberg et al., 2001; Trenberth, 2005; Trenberth and Shea, 2006; Chen et al., 2015]. Even more, warm SSTs anomalies in the TNA region have been suggested as a trigger of El Niño events [Ham et al., 2013; Wang et al., 2017].

Changes in TNA SSTs are also determinant for the availability of moisture in the Amazon River basin, since the dynamics of the Intertropical Convergence Zone (ITZC) follows the annual cycle of SST over the TNA, which is associated with well-known precipitation patterns over land [Aceituno, 1988; Marengo, 1992; Fu et al., 2001; Poveda et al., 2006; Nobre et al., 2009; Yoon and Zeng, 2010; Gimeno et al., 2012; Yin et al., 2014; Arias et al., 2015]. The activity of the trade winds over the TNA region is related with the South American Low-level Jet (SALLJ), the South American Monsoon System (SAMS), and the dynamics of aerial rivers that combined configure the transport of moisture from Amazonia all the way to the eastern flank of the Andes, and upon veering to the southeast reach as far as northern Argentina [Marengo et al., 2004; Vera et al., 2006; Moraes-Arraut et al., 2011; Poveda et al., 2014]. Also, TNA SSTs can modulate the interannual variability of fires in southern Amazonia [Chen et al., 2011; Fernandes et al., 2011].

The dynamics of SSTs in the TNA are controlled by diverse mechanisms acting at multidecadal, interannual, seasonal and intra-seasonal timescales, such as the Atlantic Multidecadal Oscillation (AMO) that operates on time scales of more than 50 years [Steinman et al., 2015], and the North-South Atlantic SST dipole that operates on time scales of 8-12 years [Enfield and Mayer, 1997; Enfield et al., 1999]. At interannual timescales there is a strong influence of El Niño on TNA SSTs positive anomalies, resulting, among others, from an anomalous Walker circulation that drives the teleconnection between the tropical Pacific and Atlantic basins [Saravanan and Chang, 2000; Chiang and Sobel, 2002], while the study by Poveda and Mesa (1997) showed statistical evidence and concluded that such interaction is mediated by land surface-atmosphere feedbacks and convection anomalies over the Amazon River basin, leading to the warming of SSTs over the TNA region following an El Niño event. Those authors went on to propose that tropical South America acts as a land surface-atmosphere bridge connecting the tropical Pacific and TNA.
On the other hand, the hydrometeorology of the Amazon River basin is not a passive spectator receiving the influence from large-scale ocean-atmosphere phenomena. There is evidence of a systematic change in the ITZC location during the boreal spring related with the propagation of coupled convective Kelvin waves emerging from AM and travelling towards the oceanic region [Wang and Fu, 2007a]. Also, wet anomalies in the equatorial Amazon have been related with SST negative anomalies over the Caribbean and the Gulf of Mexico [Misra and DiNapoli, 2012] at seasonal time scales.

Through lagged correlations, the work of Yoon and Zeng (2010) evidenced a possible feedback mechanism between AM precipitation and TNA SSTs, such that the decrease of convective activity in the Amazon leads the increase of TNA SSTs with a 3 to 5 month-lag, although those authors discarded any feedbacks and interpreted it in terms of the influence of the Pacific Ocean over both the AM and TNA regions. We are interested in further investigating the possible influence of the AM over the TNA SSTs and detailing the existence of two-way feedback mechanisms acting between the AM and TNA regions at interannual timescales. Our analysis is based on previous studies such Poveda and Mesa (1997), as well as Misra and DiNapoli (2012), on the influence of the AM precipitation over the Caribbean SSTs.

We aim at investigating the mechanisms proposed by Poveda and Mesa (1997) to provide further support to the existence of feedback processes between the hydroclimatology of the Amazon River basin and ocean-atmosphere processes over the TNA. To that end, we use a dynamical systems approach that overcomes diverse well-known shortcomings of cross-correlation analysis, in particular the highly non-linear character of the processes involved, which may introduce spurious cross-correlations with other hydro-climatic variables, thus clouding the interpretation of the physical mechanisms dynamics [Olden and Neff, 2001; Runge et al., 2014].

We use a recurrence measure to quantify the non-linear lagged dependence based on the evolution of trajectories of the dynamical system in phase space, which does not require an a priori knowledge of the equations governing the system’s behavior, useful in the analysis of hydrological and climatological time series [Marwan and Kurths, 2002; Panagoula and Vlahogianni, 2014], and to unveil the directionality and time delay in the relation between two or more variables [Goswami et al., 2013; Marwan et al., 2013]. Recurrence analysis does not require fitting of any kind of model to data, thus providing an alternative for studying non-linear, non-stationary, and high-dimensional processes [Proulx et al., 2009].

The present work is organized as follows. Section 2 outlines the main physical hypothesis of the feedback mechanism between the two regions and the data used in the present analysis. Section 3 provides a detailed methodological description of the recurrence framework. In section 4 we present the results of
3.2 The proposed mechanism and data sets

Here we propose the following physical processes and variables involved in the proposed two-way feedback mechanisms between the AM and TNA regions at interannual timescales (Fig. 3-1a). Driest (wettest) events in Amazonia (represented by the Precipitation Index, $P-E$), hereafter denoted as $P$, are characterized by a decrease (increase) in rainfall, which in turn acts to increase (decrease) atmospheric surface pressure in the AM region, thus reducing (increasing) the surface atmospheric pressure gradient between the TNA and AM regions, denoted as $G$, which in turn causes a slowing down (speeding up) of the zonal trade wind velocities over the TNA region, denoted as $W$ (Fig. 3-1b). The weakening (strengthening) of the trade winds over the TNA is associated with a reduction (increase) in evaporative cooling and a subsequent increase (reduction) in TNA SSTs, denoted as $S$. Fig 3-2 shows results from lagged cross-correlation analysis between the time series of the $P-E$ Index and TNA SSTs at interannual timescales (filtering out the annual cycles), which indicates that the AM hydrology leads the future TNA SSTs (Fig. 3-1b) from 0 to 2 month-lags (positive lags).
Figure 3-1: **a.** Location of the regions under study, with the Amazon River basin (AM) in green, and the Tropical North Atlantic (TNA) in grey. **b.** Illustration of the physical mechanisms and variables involved in the study. $P$ is the P-E index, $G$ is the surface pressure gradient between AM and TNA, $W$ represents the zonal winds over the TNA region, and $S$ represents TNA SSTs. $SP_{AM}$ and $SP_{TNA}$ are the surface atmospheric pressures on both regions, and the relative position indicates that surface pressure is higher over the TNA than AM. The grey arrow represents the wind flowing between both regions from higher to lower surface atmospheric pressure.

Figure 3-2: **a.** Physical mechanisms involved in the AM-to-TNA feedback. Each one of the nodes represents a variable involved in the process: $P$ is the P-E index, $G$ is the surface pressure gradient between AM and TNA, $W$ represents the zonal winds over the TNA region, and $S$ represents TNA SSTs. The dashed arrow represents the identified lagged cross-correlation found between the P-E index and TNA SSTs, the filled arrows represent the sequence of physical processes connecting the Amazon hydrology with TNA SSTs. **b.** Lagged cross-correlations between low-pass filtered SSTs over the TNA and the AM P-E index. The Amazon hydrology leads the TNA SSTs, with negative correlations peaking at 1 to 2 month-lags, suggesting that lower (higher) $P$ over the AM basin is related with higher (lower) TNA SSTs. The blue dashed lines denote the 95% confidence intervals.

The P-E index is defined as the difference between monthly values of Precipitation and Evapotranspiration averaged over the AM River basin, defined from 79.5°W to 50.5°W and from 19.5°S to 4.5°N [Mayorga et al.,...
We also used time series of zonal wind velocities, $W$, and sea surface temperatures, $S$, over the TNA, and the gradient of surface atmospheric pressures between the TNA and AM regions, $G$.

The precipitation data set was obtained from the Global Precipitation Climatology Centre (GPCC) that produces gridded (1° x 1°) data, based on monthly rainfall collected from more than 80,000 stations worldwide with a rigorous quality control [Schneider et al., 2014]. The evapotranspiration data set was obtained from the land surface model ORCHIDEE (1° x 1°), developed and maintained by the Institute Pierre Simon Laplace [Krinner et al., 2005]. The resulting P-E index was calculated from 1979 to 2008, the shared period among both data sets.

The TNA SSTs are defined for the region from 75°W to 10°W and from 5°S to 29°N. The SST data set was obtained from the third version of NOAA’s Extended Reconstructed Sea Surface Temperature ERSST (v3b), with a spatial resolution of 1° x 1°, and spanning the period from 1979 to 2008. The SST data has a cold bias correction due to satellite data in previous ERSST reconstructions [Smith et al., 2008]. Data sets pertaining to surface atmospheric pressure for both regions (AM and TNA), and zonal wind velocities over the TNA region were obtained from the ERA-Interim Reanalysis [Dee et al., 2011], with a spatial resolution of 0.75° x 0.75°. The analyzed time series were estimated as regional averages, and the surface atmospheric pressure gradient was calculated as the difference in surface pressures between the TNA and AM. We computed the monthly anomalies for each of the series in order to remove the annual cycle. The final time series used in the present study are presented in Fig. 3-3.

### 3.3 Methods

#### 3.3.1 Recurrence Analysis

Hydro-climatic and many other dynamical systems tend to have a recurrent behavior in phase space, so that recurrence can be considered as a fundamental property of these kinds of systems [Marwan et al., 2007a]. This property can be visualized by the trajectories of the system’s dynamics in phase space, and can be quantified by the so-called Recurrence Plot (RP) [Eckmann et al., 1987]. In order to construct an RP from a time series, $x_i$, it is necessary to represent the m-dimensional phase space of a system $X$. In case of single time series $x_i$, the dynamics has to be artificially reconstructed using the time delay embedding technique [Packard et al., 1980; Takens, 1981], whereby the phase trajectories $\tilde{x}_i$ are defined as:

$$\tilde{x}_i = (x_i + x_{i+\omega}, \ldots, x_{i+\omega(m-1)}), \tilde{x}_i \in \mathbb{R}^m$$
Figure 3-3: Anomaly time series used in the present study, from top to bottom: AM P-E index (mm/month), TNA SSTs (°C), TNA zonal wind velocities (m/s), and TNA-AM surface pressure gradient (Pa).

where \( m \) is the embedding dimension and \( \omega \) is the time delay. The most common methodologies to quantify both embedding parameters are the false nearest neighbors for the embedding dimension, and the mutual information function for the time-delay [Fraser and Swinney, 1986; Marwan, 2011]. Once the parameters are set, the RP can be estimated using the pair-wise proximity test, such that,

\[
R_{i,j}^X = \Theta(\varepsilon - \|\vec{x}_i - \vec{x}_j\|) \quad i, j = 1, \ldots N', \quad (3-2)
\]

where \( N' = N-(m-1)k \) is the number of phase space vectors, \( \varepsilon \) is the threshold defined for the proximity between the phase space vectors, \( \|\vec{x}_i - \vec{x}_j\| \) is the
spatial distance between vectors in phase space, and $\Theta(\ )$ is the Heaviside function: $(\Theta < 0 ) = 0$, $(\Theta \geq 0 ) = 1$. The plot of the $R^X$ recurrence binary matrix provides the $RP$ (Fig. 3-4). The probability that one system recurs to a certain state $\vec{x}_i$ is equal to the column-average of the recurrence matrix [Marwan et al., 2013]:

$$p(\vec{x}_i) = \frac{1}{N_f} \sum_{j=1}^{N_j} R^X_{i,j} \quad (3-3)$$

Joint recurrence plots ($JRP$) are used to study the possible influence between two physically different systems [Romano et al., 2004; Marwan et al., 2007b], as they provide a measure of the simultaneous recurrence in both. The $JRP$ matrix is defined as the Hadamard product of the $RPs$ of the single $RPs$ of systems $X$ and $Y$:

$$JR^{XY}_{i,j} = \Theta(\varepsilon - \|\vec{x}_i - \hat{x}_j\|) \times \Theta(\varepsilon - \|\vec{y}_i - \hat{y}_j\|) i, j = 1, \ldots N'. \quad (3-4)$$

The probability of finding a simultaneous recurrence at time $i$ in both systems $X$ and $Y$ is equal to the column-average of the $JR^{XY}$ matrix:

$$p(\vec{x}_i, \vec{y}_i) = \frac{1}{N_f} \sum_{j=1}^{N_j} JR^{XY}_{i,j} \quad (3-5)$$

As we are interested in quantifying the dependency between the time series with a specific time lag $\tau$, we will use the log mean recurrence measure of dependence, $RMD(\tau)$, proposed by Goswami et al. (2013), as a measure of the probability that the trajectory of $X$ recurs to the $\varepsilon$ neighborhood of $\vec{x}_i$, when the trajectory of $Y$ recurs simultaneously to the $\varepsilon$ neighborhood of $\vec{y}_i$, after some lag $\tau$, and is given by:

$$RMD(\tau) = \log_2 \left( \frac{1}{N''} \sum_{i=1}^{N''} \frac{p(\vec{x}_i, \vec{y}_i(\tau))}{p(\vec{x}_i)p(\vec{y}_i(\tau))} \right) \quad (3-6)$$

were $N''=N' - \tau$. If both systems $X$ and the time lagged $Y(\tau)$ are independent, then $p(\vec{x}_i, \vec{y}_i(\tau)) = p(\vec{x}_i)p(\vec{y}_i(\tau))$, which implies that $RMD(\tau)=0$. For $\tau > 0$, a non-zero $RMD(\tau)$ implies that $Y$ is dependent on $X$, and the contrary is true for $\tau < 0$. 

Figure 3-4: Recurrence plots of the time series as presented in Fig. 3, using embedding dimension 4, time delay 1 and a recurrence threshold ensuring a fixed recurrence rate of 5%.

3.3.2 Significance test

Our approach requires implementing a statistical test for the hypothesis of a joint recurrence between the systems $X$ and $Y$ at defined time lags. It is based on the joint recurrence between the trajectory in phase space of the original system $X$ and a representative number of phase space trajectories of the system $Y$, which represent independent copies of the underlying system, known as twin surrogates ($TS$). For the construction of the $TS$ based on the RP of a system, we followed the procedure proposed in Thiel et al. (2006, 2008).

Twins can be defined as points in the phase space trajectories that share the same neighborhood up to the threshold $\epsilon$ such that $R_{i,k} = R_{i,k}^{Y}$, $k = 1, \ldots, N$. Thus, the construction of a twin surrogate $\tilde{x}^{s}(t)$ of $\tilde{x}(t)$ starts with the identification of all the twin points in a trajectory $\tilde{x}$, then one must choose one arbitrary starting point $\tilde{x}^{s}(1) = \tilde{x}(k)$. If this $\tilde{x}(k)$ has no twin then the next point in the
surrogate trajectory is $\tilde{x}^s(2) = x(k + 1)$, but if $\tilde{x}(k)$ does have a twin $\tilde{x}(t)$ then the next point can be either $\tilde{x}(k + 1)$ or $\tilde{x}(t + 1)$ with equal probability, and thus the process is iterated until the constructed surrogate has the same length of the original time series.

The null hypothesis is that each TS trajectory is an independent realization of the system, corresponding to a different initial condition. To test the statistical significance of the $RMD(\tau)$ measure among $X$ and $Y(\tau)$, we generate 500 TS of the $Y$ system, obtain a test distribution of $RMD(\tau)$ calculated with the observations of the system $X$ and each one of the TS of $Y$. Finally, we construct the upper 95% and 90% confidence intervals (CI) based on the respective 95th and 90th percentiles of the test distribution. The lags at which values fall outside the confidence band represent the ones where the null hypothesis is rejected, or that the system $X$ drives system $Y$, i.e. they are dependent at that particular lag.

### 3.4 Results and discussion

We used the false nearest neighbor and mutual information methods to estimate the embedding parameters (Fig. C1, Appendix C) to construct the joint recurrence plots for the time series studied here. We found $\omega = 1$ and $m = 4$ for $P$ and $S$, $\omega = 1$ and $m = 6$ for $W$, and $\omega = 1$ and $m = 5$ for $G$. We decided to set the parameters for all combinations as $\omega = 1$ and $m = 4$ to avoid sparse recurrence plots, as recommended in Marwan (2011). The recurrence threshold, $\varepsilon$, was based on a fixed 5% of the recurrence rate, $RR = 1/N^2 \sum_{i,j} R_{ij}$, for all the time series.

Before presenting the analysis of the dynamical mechanisms involved in the studied feedbacks we will evaluate the recurrence analysis for $P$ leading $S$, whose results are shown in Fig. 3-5. Using the RMD measure, we find that the P-E index ($P$) in the AM region exerts a significant influence over the TNA STTs from 1 to 2 month-lags, and the $RMD$ values are in the 90% confidence band. Comparing this result with the lagged cross-correlations among the two variables presented in Fig. 3-1b, and the results of (Yoon and Zeng 2010, Fig. 10), one can see that although the relation between the variables is well represented by the two measures, the lagged cross-correlations show a (low) peak at 0 month-lag (simultaneous), while the non-linear measure reflects a higher persistence in time that may last for more than two months.

Fig. 3-6 shows the recurrence lagged dependence measures of the four time series studied, according to the leading patterns involved in the AM-to-TNA feedback physical processes. Fig. 3-6a allows us to conclude that $P$ is leading $G$ by one month lag above the 95% CI, and by zero and two month-lags above the 90% CI. The identified influence of the AM hydrology (mainly precipitation-convection) over the pressure gradient between the two regions at the 0 month lag may be related with the Amazon modulation of the
intensity and location of convection in the TNA reported by Wang and Fu (2007) in a time scale from 5 to 7 days.

Figure 3-5: Recurrence lagged dependence between the $P$ and $S$ time series. The arrow denotes the direction of influence between variables; the gray area represents the 90% confidence area, the blue dashed lines represent the 95% confidence intervals and the red line represents the calculated RMD between the variables $P$ and $S$ (lagged).

Furthermore, $G$ (pressure gradient) has a strong influence on $W$ (zonal winds) leading from 0 to 3 month-lags, and from 8 to 12 month-lag above the 95% CI, as shown in Fig. 3-6b. The influence of $G$ on $W$ may be related with the 30 to 70-days oscillation of tropical winds reported by [Foltz and McPhaden, 2008] and the intraseasonal Madden-Julian Oscillation that has important influence in AM [Kayano and Kousky, 1999; Garreaud et al., 2009] and over the Atlantic basin, which has been related to the formation of Atlantic tropical cyclones [Klotzbach and Oliver, 2015]. Finally, $W$ leads $S$ at 0 to 2 month-lags above the 95% CI as presented in Fig. 3-6c. The overall interannual results obtained using our non-linear approach confirm the relation between AM and the tropical Atlantic; although for a much larger region over the TNA than the one found by Misra and DiNapoli (2012) regarding a teleconnection between the Amazon and the Tropical Atlantic region at seasonal timescale, mainly forced by wet anomalies in the Amazon, that in turn are associated with cooler SSTs over the Intra-Americas Seas.

According to the results of [Yoon and Zeng, 2010], after removal of the ENSO influence on both variables, the relation between AM hydrology and TNA SSTs occurs in phase (lag 0). Using the recurrence analysis we found that the feedback in the AM-to-TNA direction occurs from 0 to 2 month-lags. The whole mechanism that involves the influence of AM land surface-atmospheric processes on the TNA SSTs spans around two months with a 90 to 95% of confidence. This result provides further evidence about the influence of AM
hydrological processes over TNA SSTs, and unveils the time span required for convection anomalies over land to affect the oceanic region. These results are also consistent with the empirical analysis of [Poveda and Mesa, 1997] regarding the existence of a strong significative statistical association between the hydrology of Andean streamflows leading the TNA SSTs from 3 to 5 months in advance. This result suggests that the hydroclimatological processes acting at continental scale over northern South America and the AM may influence the TNA from one to almost five months.
Figure 3-6: Recurrence lagged dependence between the four variables according to the direction identified in the AM to TNA feedback mechanism. The arrow denotes the direction of influence between the variables, the gray area represents the 90% confidence area, the blue dashed lines gives the 95% confidence intervals, and the red line represents the calculated RMD between the variables.

These results further support the existence of two-way feedback mechanisms operating between AM and TNA at interannual timescales. Fig. 3-6d shows the recurrence analysis of S (TNA SSTs) leading P (AM hydrology), which point out a simultaneous effect (0 month-lag) with a 95% confidence. We obtained similar results for recurrence rates near 1% (Fig. C2). Fig. 3-7 summarizes the feedback mechanisms and time lags involved in the AM-to-TNA feedback, with the nodes representing the four variables involved.

Figure 3-7: Graph representing the studied feedbacks. The numbers in bold represent the time lag of dependency between the variables, and the percentage inside the parenthesis is the level of confidence for each lag.

Furthermore, our recurrence analyses confirm that two-way feedbacks are set between the two regions at interannual timescales, including a strong influence of AM convection on TNA SSTs, mediated by the surface pressure gradient between the two regions, G, and the zonal trade winds, W. A summary of the proposed two-way feedbacks mechanism is illustrated in Fig. 3-8. Furthermore, our results also unveil dependencies between G and W from 7 to 12-month lags, whose interpretation constitutes a topic of further investigation.

Our work is focused on studying the proposed two-way mechanism from the AM-to-TNA direction. But in order to understand the whole process including the TNA-to-AM direction, we performed a complete recurrence analysis
between the variables involved in the studied mechanism. Our results confirm the two-way character of the AM-TNA coupling, as summarized in Appendix C.

Figure 3-8: Graphical summary of the dynamics of the two-way feedback mechanism originated in the AM region, represented step by step from top to bottom. Left panels show the recurrence analysis results, whereas right panels illustrate the dynamics of the physical processes involved. The arrows beside all variables represent the dynamical changes undergone by the relevant variables. In summary, the mechanism is triggered by a decrease in \( P-E \) over the AM and the ensuing dynamics develops, as shown in pairs of panels a-b, c-d and e-f. Panel (a) shows the recurrence analysis regarding \( P-E \) driving \( G \) from 0-2 months, and panel (b) illustrates that such an influence is mediated by a reduction of \( P-E \), an increase of surface pressure in the AM that leads a decrease in \( G \). Panel (c) shows the influence of \( G \) driving \( W \) from 0-3 months, as discussed in panel (d) through a reduction in the surface pressure gradient between the AM and TNA, and the reduction of the arrow representing the magnitude of wind speed, \( W \). The final step in panel (e) shows the influence of \( W \) over \( S \) from 0-2 months, through the mechanisms discussed in panel (f), with the appearance of an increase in sea surface temperatures, \( S \), due to the ongoing reduction of \( W \).

At this point it is necessary to detail the proposed two-way mechanism by disentangling the physical dynamics between the variables \( P, G, W \) and \( S \). We propose that the mechanism is triggered by anomalous convection processes in AM. Therefore, we conduct the physical analysis by evaluating the behavior of monthly anomalies of \( P, W \) and \( S \) in four particular contrasting extreme convective conditions in the AM, with anomalies greater than \( \pm 20 \) mm month\(^{-1}\). The selected events are those of January 2005 with values of -30 mm month\(^{-1}\), and January 2009 with values of +20 mm month\(^{-1}\) (Fig. 3-9), September 1999 with values +25 mm month\(^{-1}\) and September 2010 with values -25 mm month\(^{-1}\) (Fig. 3-10). The selection of such extreme dry and
wet events in the months of January and September will help us to evaluate whether an anomalous interannual increase or reduction in precipitation may trigger the proposed mechanism.

In January of 2005 the AM is experiencing negative anomalies of $P$ of around 30 mm month$^{-1}$ that covers most of the basin (Fig. 3-9a), with maximum negative anomalies of more than 120 mm month$^{-1}$ in regions over the Andes and central AM. While AM experiences such reductions of $P$ it is possible to observe a reduction in zonal wind velocities from 5°N to 25°N with negative anomalies reaching 1.5 m s$^{-1}$ (Fig. 3-9b). At the same time, the SSTs between 5°N and 25°N exhibit positive anomalies of around 1.5°C. In January of 2009 the AM is experiencing positive anomalies of $P$ that cover the north-western AM (mean positive anomaly of +20 mm month$^{-1}$), with positive anomalies reaching values of more than 110 mm month$^{-1}$ (Fig. 3-9d). In the same month it is possible to observe increased zonal wind velocities from 5°N to 30°N, with positive anomalies ranging from 0.5 m s$^{-1}$ to 3 m s$^{-1}$ (Fig. 3-9e). In phase with the increased zonal wind velocities, SSTs between 5°N and 25°N exhibit negative anomalies of around 0.4°C between 45°E and 20°E (Fig. 3-9f), and the region of negative SST anomalies moves to 0°N in the subsequent February (not shown). In summary, the response of $S$ due to the influence of $P$ anomalies in the two events analyzed in January occurs from 0 to 2 months-lag. These dynamical evidences during the extreme January events are consistent with the results from the recurrence analysis, which indicate that the complete mechanism from $P$ to $S$ may be triggered and active during 0 to 3 months-lag (Fig. 3-7).

During September 1999, there are positive anomalies of $P$ over central AM with values around 100 mm month$^{-1}$ (Fig 3-10a). While AM experiences an increase in $P$, there are two zones of increased zonal wind velocities in the TNA, from 0°N to 10°N, and from 20°N to 30°N there are values of 0.5 m s$^{-1}$ (Fig. 3-10b) and between 10° and 20° there is a zone of reduced zonal winds reaching up to 1.5 m s$^{-1}$. In September 1999, SSTs show an increase of around 0.4°C (Fig. 3-10c) while during the following month (October 1999), the values reached up to 0.6°C (shown in Fig. 3-12b) that goes on until November 1999. In September 2010, the entire AM is experiencing negative precipitation anomalies of around 80 mm month$^{-1}$ (Fig. 3-10d). The zonal wind velocities over the TNA exhibit positive anomalies from 0.5 m s$^{-1}$ to 4 m s$^{-1}$ and from 10°N to 20°N (Fig. 3-10e). For the same month, values of $S$ exhibit positive anomalies greater than 1°C over the TNA from 5°N to 15°N (Fig. 3-10f). In summary, for the extreme convection events in September only the one of 2010 lead to a SST warming over the TNA, consistently with the recurrence analysis (Fig. 3-7). These previous results provide further evidence that convection anomalies over the AM region influence TNA SSTs at interannual timescales.

The two selected extreme events in January and the dry event in September (2010) are consistent with results from our recurrence analysis. For the
January events it is possible to observe the configuration of the mechanism following wet and dry convection anomalies in the AM during the Amazon wet season. The unclear development of the mechanism in September may be attributed to the influence of the Atlantic Meridional Mode (AMM) that exerts a stronger influence on the TNA than the one induced by the studied mechanism, and thus anomalous convection in AM during the dry season (such as the ones discussed for September) has less impact on the AM-to-TNA feedback. The AMM, which develops in the boreal spring (MAM) at interannual time scales, is related with the SSTs anomalies over the TNA due to anomalous displacements of the ITCZ, and may be acting in phase with ENSO events magnifying the SST’s warming during boreal spring to boreal summer [Mitchell and Wallace, 1992; Nobre and Sukla, 1996; Good et al., 2008b; Foltz and McPhaden, 2010; Foltz et al., 2012; Amaya et al., 2016].

Aiming at a more comprehensive identification of the spatial extent of the regions involved in the AM-TNA feedbacks, we define a larger region of the TNA including a broader equatorial portion of the Atlantic and the Intra-American Seas (IAS). From our previous dynamical analysis of the variables $P$, $W$ and $S$ we may infer that convection over AM might be most related with the TNA region north of $0^\circ$N, as is the region experiencing the main changes resulting from the studied mechanism. On the contrary, the more equatorial region of the TNA seems to be more stable and more influenced by the strong seasonality of winds, as proposed by Li and Philander (1997).

We also evaluated the monthly mean conditions of the relevant variables during the extreme AM droughts of 1963, 1980, 1983, 1997, 1998, 2005, and 2010, and also during the floods of 1989, 1999, and 2009 (Marengo and Espinoza, 2016). Our aim is to further understand the dynamics of the two-way interactions between AM and TNA in the context of extreme dry and wet events in the AM region, as a key mechanism involved in the warming/cooling of the TNA (Fig. 3-11). In order to capture the extremes of 1963, 2009 and 2010 that cannot be captured with the $P$-$E$ index (spanning from 1979 to 2008), we used the GPCC average precipitation (AP) over the AM to represent hydrological process over land.
Figure 3-9: Monthly anomalies of (a) precipitation, (b) zonal winds, and (c) SSTs for January 2005. Monthly anomalies of (d) precipitation, (e) zonal winds, and (f) SST for January 2009. Dashed (solid) lines denote negative (positive) anomalies.
Figure 3-10: Monthly anomalies of (a) precipitation, (b) zonal winds, and (c) SSTs for September of 1999. Monthly anomalies of (d) precipitation, (e) zonal winds, and (f) SST for September of 2010. Dashed (solid) lines denote negative (positive) anomalies.
During the extreme AM droughts (red lines in all panels of Fig. 3-11), average precipitation (AP) in January reaches negative anomalies of around 20 mm month$^{-1}$, thus implying a reduction of convective processes during the peak of the wet season. One month later, in February, the surface atmospheric pressure gradient between AM and TNA (G) reaches a minimum value, in phase with minimum zonal wind velocities (W) from the TNA towards the AM. During the first three months of the year, there is a sustained SST warming that continues until May. The further warming of the SSTs from April to May, after the influence of the reduced convection in AM, may be related to the effects of the AMM during boreal spring that induces warmer SSTs at interannual timescales. On the contrary, during extreme AM floods (blue lines in all panels of Fig. 3-11) there is an increased precipitation in January followed by higher pressure gradients, faster winds towards the AM, and cooler TNA SSTs. As suggested in our previous results convection anomalies may trigger the mechanism, but during the anomalous peak of convection found in June in flood years (see AP in Fig. 3-11) there is no evidence of a concomitant increase in G, which may be explained by the control that the ongoing SST warming from April to August exerts over the oceanic surface pressure, and thus over G. Although the 0 to 3 month-lag found in the AM-to-TNA feedback mechanism is well represented in the behavior of the time series of mean conditions during extreme events in the AM (light blue shadings in Fig. 3-11), wet and dry years in the AM have different dynamics and the action of the mechanism is most evident during periods of negative precipitation anomalies in January.

So far we have shown the physical mechanisms that connect the AM and TNA regions at interannual timescales, specifically during extreme events of precipitation in the AM. The influence of AM during such events may be mediated by a common driver such as the ENSO system. To verify the influence of ENSO in the two-way feedback mechanisms between AM and TNA, we compare the time evolution of the variables during four selected extreme events, defined in Table 1 of Marengo and Espinoza (2016): the flood of 1999 related to La Niña, the drought of 2010 related with El Niño, and two other events unrelated to ENSO: the 2009 flood, and the 2005 drought. Fig. 3-12 shows the time evolution of the relevant variables for the chosen extreme years. The evolution of the mechanism during the droughts (Fig. 3-12a) indicates that in both episodes AP exhibits a strong negative anomaly in January, followed by negative anomalies in G and W, and warmer than normal SSTs; both droughts showing differences in the magnitude and behavior of the precipitation anomalies, which implies that the mechanism operates differently: for the non ENSO-related drought of 2005 (Fig. 3-12a, left) there is a delay of two months between the peaks of AP and G, and of one month between W and S. In the case of the ENSO-related drought of 2010 (Fig. 3-12a, right) the decrease in G and W occurs simultaneously, and with a 2-month delay between W and S. It is worth noticing that G exhibits abrupt changes in March 2010, which may be related to the influence of ENSO in the
surface pressure of both regions due to an anomalous Walker circulation in the Amazon [Lewis et al., 2011a; Marengo et al., 2011].

Figure 3-11: Monthly mean values of GPCC precipitation (AP), surface atmospheric pressure gradient between AM and TNA (G), zonal winds (W), and SSTs (S), during extreme dry years (red lines): 1963, 1980, 1983, 1997, 1998, 2005, and 2010, and during extreme wet years (blue lines): 1989, 1999, and 2009 in AM. Negative zonal winds represent a reduction of the winds flowing from the TNA to the AM. Light blue shading represents the temporal progression of the mechanism during dry years from the beginning of the year (peak of the AM wet season).
Figure 3-12: Average GPCC precipitation (AP), surface atmospheric pressure gradient (G), zonal winds (W), and SSTs (S), for the droughts of 2005 and 2010 (a), and the floods of 1999 and 2009 (b) in AM. ENSO-related extremes depicted in the panels of the right in (a) and (b), and non-ENSO related extremes depicted in the left panels in (a) and (b). Negative zonal winds represent a reduction of the winds from the TNA towards the AM.

For the flood years (1999 and 2009) the peaks of AP, G and W occur simultaneously (Fig. 3-12a and Fig. 3-12b), and the whole mechanism is fully fledged from January to February. This result is explained given that from November onwards the Amazon River basin is experiencing the wet season, and thus positive precipitation anomalies are present even before January. For
the 2009 flood (Fig. 3-12b, left) one may expect that the AP peak of March will trigger the mechanism, but the ocean is experiencing the passage of the ITCZ, so there is another forcing in S that influences also the evaporative cooling in the ocean. These results may indicate that although the mechanism is well developed during the occurrence of El Niño and La Niña episodes, it is also present when the extreme convection events in the AM are not influenced by ENSO. There are two peaks of anomalous precipitation during the months of September 2010 (Fig. 3-12a) and September 1999 (Fig. 3-12b) that may be related with the triggering of the mechanism. During 2010 the time evolution of the variables is not conclusive, and during 1999 the mechanism appears to be weak and to take three months from the P anomaly to the SST anomaly. An extensive analysis was already presented in the discussion of Fig. 3-10.

3.5 Concluding remarks

Convective processes in the Amazon are characterized by a strong seasonality, where climate systems such as ENSO (interannual) and the migration of the ITCZ (seasonal) are key players. The moisture flux through the Amazon River basin and further south in the continent is driven by the well-known development of the SAMS system and the important source of moisture from the TNA to the AM. Thus the availability and dynamics of moisture in the Amazon are of utmost importance for the stability of the whole hydroclimate in South America. There is an ongoing warming of the TNA SSTs, reported as a part of a decadal cycle of the Atlantic that has a strong relation with droughts in the Amazon [Lewis et al., 2011a; Marengo et al., 2011]. Also, in recent years there have been reports on how the surface and atmospheric branches of the hydrological cycle in the Amazon River basin exhibit a positive trend toward wetter conditions [Gloor et al., 2013] and how the dry season tends to lengthen in duration [Fu et al., 2013].

Therefore, the identified two-way feedback mechanisms between the AM and TNA regions are of paramount importance towards understanding the dynamics of particular of extreme events at interannual timescales over both regions. Our results point out a significant influence of the AM hydrology over the TNA SSTs, and unveil that such influence is complex and has to be mediated by the surface atmospheric pressure gradient between both regions and by the zonal winds (G and W). We have also found that the identified feedback processes develops from 0 to 2 months after convection anomalies in the AM, and that dry conditions in the AM have the potential to increase TNA’s SSTs.

Using a nonlinear technique, we have gone further in the analysis of the teleconnections between the AM and the TNA, putting forward the hypothesis that convective negative (positive) anomalies in the AM can trigger an increase (decrease) of the TNA SST’s not necessarily mediated by ENSO. Our results allow us to conclude the AM is not a passive spectator of large scale
ocean-atmospheric phenomena, but a key player in the tropical climate at interannual time scales. We have also found that the AM influence over the TNA at interannual timescales goes beyond the IAS region (Misra and DiNapoli, 2012), covering an even greater extent of the Tropical Atlantic Ocean. The dynamical analysis strongly supports our results from the recurrence analysis regarding the coupling of the studied variables in time.

There is evidence suggesting that the Amazon interannual hydrological variability is getting more extreme [Fu et al., 2013; Gloor et al., 2013; Yin et al., 2014; Zou et al., 2015]. According to our results, a continuous warming of the TNA SST’s may be amplified by a drier Amazon, and may affect the region’s ocean-atmosphere dynamics and diverse mechanisms and processes such as tropical easterly waves, tropical storms, which might induce more frequent and intense hurricanes over the TNA and the Caribbean Sea, that in turn could cause positive feedbacks into the continent, thus accelerating desiccation and vegetation dieback of Amazonia, which in turn could lead to a tipping point in the Amazonian hydroclimate [Lenton et al., 2008], with significant consequences at continental and global scales.

As suggested by our results, future anomalies in rainfall over the Amazon River basin possibly caused by climate change, climate variability, deforestation and land use/land change might thus be related to anomalous future stages of TNA’s SSTs. Our results also suggest that the Amazon plays a key role in the TNA warming, reinforcing the feedback and triggering more severe droughts, such as the well-studied two one-in-a-century record breaking events of 2005 and 2010 in the Amazon.

Although the recurrence measure reported here has proven to be useful in the study of lagged dependencies, there are still shortcomings that can be addressed like the selection of parameters or the identification of the type of interaction between the variables in order to detect positive or negative dependence. Other dynamical systems measures as the transfer entropy is an alternative to quantify causality between variables excluding the presence of an extra common driver, although it requires longer time series and greater computational resources.
4. Atmosphere-land Bridge between the Pacific and Tropical North Atlantic SST’s through the Amazon River basin during the 2005 and 2010 droughts

Alejandro Builes-Jaramillo, Antônio M.T. Ramos and Germán Poveda

Abstract: In the present work we study the role of the Amazon River basin (AM) as a land-atmosphere bridge between the Niño 3.0 region in the Pacific Ocean and the Tropical North Atlantic. We use a novel approach of causal inference between complex systems called the Recurrence Measure of Conditional Dependence (RMCD) based on the recurrence plots theory. We selected two anomalous droughts in the Amazon River basin, one mainly attributed to the warming of the Tropical North Atlantic (2005) and the other to a warm phase of ENSO (2010). We have compared the RMCD analyses during the two extreme droughts to find evidences of distinctive behavior in the information transfer between the two oceanic regions. During the 2005 drought we found indications of a strong connection between the Tropical North Atlantic and the Amazon River basin and not so with the Pacific, confirming previous studies. During 2010 the influence of the Pacific over the Amazon River basin was found to be significant for 5 to 7-month lags, and also the Amazon River basin exerted a significant influence on the Tropical North Atlantic, thus indicating that the proposed land-atmosphere bridge was active during the 2010 El Niño. We studied also the direct influence of the tropical Pacific over the Tropical North Atlantic during 2010 and found a significant causal relationship between the two oceans. RMCD proves to be remarkably consistent regarding the information transfer from the tropical Pacific to the AM six major sub-basins, but also for the Andes and the low lying Amazonia, the influence being stronger between 5 to 7-month lags in 2010 and also Andes receives the information transfer from the Pacific one month before Amazonia. During the 2005 drought, two results differ from the entire Amazon River basin: (i) there is a two month temporal gap in the information transfer from the tropical Pacific to each one of the regions, and (ii) the Andes receives the information transfer from the Pacific one month earlier than Amazonia. Our results confirm that the land-atmosphere bridge operating over the AM is an active hydroclimate mechanism at interannual timescales, and the RMCD analysis may be an ancillary resort to complement early warning systems.

Key words: Amazonia, Recurrence Plots, Hydrometeorology, Causality, ENSO, RMCD
4.1 Introduction

The influence of El Niño–Southern Oscillation (ENSO) events in the tropical Pacific on the global hydroclimatic variability is well documented [Cane, 2005; Sarachik and Cane, 2010; Ward et al., 2014]. Most regional and local impacts associated with ENSO are mediated by both direct influences and ocean-atmosphere teleconnections that drive continental, regional and local hydroclimate processes and weather conditions, related with droughts (floods) in regions as far as the Yangtze basin or South Africa [Cayan et al., 1999; Wang et al., 2000; Pozo-Vázquez et al., 2001; Zhang et al., 2007; Yu et al., 2012; Davey et al., 2014], and changing patterns in winter (summer) temperatures in oceans such as the North and Tropical Atlantic, Caribbean and Antarctic [Diaz et al., 2001; Turner, 2004; Yuan, 2004; Wang, 2005].

With respect to northern South America and the Amazon River basin, diverse studies have linked ENSO with extreme hydroclimatic events [Ropelewski and Halpert, 1987, 1996; Aceituno, 1988, 1989; Kiladis and Diaz, 1989; Halpert and Ropelewski, 1992; Marengo and Hastenrath, 1993; Poveda and Mesa, 1997; Hastenrath, 2000b; Poveda and Salazar, 2004; Andreoli and Kayano, 2005; Poveda et al., 2006, 2010; Araújo Gonzalez et al., 2007; Garreaud et al., 2009; Valverde and Marengo, 2011; Barreiro and Díaz, 2011; Tedeschi et al., 2014, 2016; Shimizu et al., 2016; Marengo and Espinoza, 2016]. Fig. 4-1 shows the regions that have been found to be under strong El Niño influence during the DJF and JJA seasons.

![Figure 4-1: Geographical locations where El Niño events may impacts while the northern summer (DJF) and winter (JJA). Image retrieved from NOAA](http://bit.ly/2h18XLM)
Chapter 4

The connections between the Pacific Ocean (PAC) and the Tropical North Atlantic (TNA) SST’s during ENSO have been a research topic for several decades, and there is evidence of Atlantic SST anomalies change during El Niño events. These alterations have been typically explained through a vertical stabilization of the tropical troposphere that may induce such feedbacks with a delay related with weaker trade winds and fluxes over the Atlantic due to an anomalous Walker circulation [Saravanan and Chang, 2000; Giannini et al., 2001], and also through an “atmospheric bridge” [Ngar-Cheung and Mary, 1996; Latif, 2001; Roy and Reason, 2001]. Furthermore, Poveda and Mesa [1997] showed evidence that the PAC to TNA connection occur through a “land-atmosphere” bridge, whereby hydrological processes operation over northern South American, mainly occurring in the Amazon River basin, play a key role in teleconnecting both oceanic regions during El Niño events.

The impacts of ENSO in the Amazon River basin are well documented, and are mainly related with interannual anomalies in precipitation triggered by a displacement of the Walker circulation over South America, that may lead to droughts (floods) all over the basin [Ropelewski and Halpert, 1987; Marengo and Hastenrath, 1993; Coe et al., 2002; Poveda and Salazar, 2004; Drumond and Ambrizzi, 2006; Poveda et al., 2006; Araújo Gonzalez et al., 2007; Davidson et al., 2012; Espinoza et al., 2012]. The increase of anomalous extreme events that may be induced by climate change [Marengo and Espinoza, 2016] could be related with a possible change in the Amazon ecosystems and in the equilibrium state of the Amazon from forest to savannah [Holmgren et al., 2001; Nepstad et al., 2004, 2008; Salazar et al., 2007, 2016; Nobre and Borma, 2009; Li et al., 2011; Swann et al., 2015]. At seasonal time scales the main forcing mechanisms of the Amazon River basin hydroclimatology are the TNA, the Intertropical Convergence Zone (ITCZ), and the South Atlantic Convergence Zone (SACZ), influencing the availability of moisture, surface air temperatures and extreme hydrological events [Hastenrath and Heller, 1977; Aceituno, 1988; Marengo, 1992; Zeng, 1999; Poveda et al., 2006, 2014; Nobre et al., 2009; Yoon and Zeng, 2010; Yin et al., 2014; Arias et al., 2015; Zemp et al., 2017]. The importance of the TNA on hydroclimatic processes over the Amazon River basin are well documented [Servain, 1991; Nobre and Sukla, 1996; Souza et al., 2000; Hastenrath, 2000a; Bosilovich et al., 2002; Ronchail et al., 2002; Labat et al., 2004; Good et al., 2008b; Yoon and Zeng, 2010; Moraes-Arraut et al., 2011; Gimeno et al., 2012; Arias et al., 2015]

On the other hand, owing to its size (more than 6.2 million km2), cross equatorial location, land-cover types and hydroclimatic dynamics, the Amazon River basin has been considered as a green ocean and a hotspot of Earth’s climate dynamics [Lenton et al., 2008; Nobre and Borma, 2009]. The strong role of soil moisture and evapotranspiration on precipitation recycling over the Amazon River basin has been known for decades [Lettau et al., 1979; Salati and Nobre, 1991; Eltahir and Bras, 1994; Zemp et al., 2014; Rocha et al.,]
and therefore the region is not just a passive spectator of the forcing from large-scale ocean-atmospheric phenomena. For instance, there is evidence that deforestation in Amazonia severely reduces rainfall in the lower U.S. Midwest during the spring and summer seasons and in the upper U.S. Midwest during the winter and spring [Avissar and Werth, 2004]. Furthermore, land surface and hydroclimatic processes taking place in the Amazon River basin have been shown to affect the Tropical North Atlantic and Caribbean Sea SSTs [Wang and Fu, 2007b; Misra and DiNapoli, 2012]. Those studies found evidence that convection anomalies in the Amazon induce the displacement of the ITCZ and SST’s in the Caribbean. Wang and Fu showed evidence of the presence of coupled convective waves emerging from Amazonia and traveling over the TNA to the African coast.

The study by Poveda and Mesa, [1997] put forward and provided statistical evidence of the existence of a land-atmosphere bridge acting over northern tropical South America that connects SST anomalies over the Tropical Pacific and Tropical North Atlantic Oceans, through a suite of physical mechanisms discussed and depicted by Poveda et al., [2006]. More recently, Builes-Jaramillo et al. [2017, submitted] provided further evidence about the existence of such a land-atmosphere bridge mechanism over the Amazon River basin connecting the tropical Pacific and TNA, and explain the functioning of a two-way feedback physical mechanism between the TNA SSTs and the Amazon River basin hydrology during El Niño events, and also during neutral years. The two-way interactions between the Amazon River basin and the Tropical North Atlantic are mainly driven by changes in the surface pressure difference between the two regions, in turn induced by anomalies in land surface-atmospheric processes and convection in the Amazon, the mechanism acting from the continent to the TNA, during El Niño events and neutral years.

In the present work, we aim to further understand the dynamics of such land-atmospheric bridge linking the Tropical Pacific (PAC) and Tropical North Atlantic (TNA) oceans, with particular emphasis on the two “droughts of the century” occurred in the Amazon River basin during 2005 and 2010 [Marengo et al., 2008, 2011; Chen et al., 2009; Espinoza et al., 2011; Frolking et al., 2011; Lewis et al., 2011b]. To that end, we use tools from non-linear dynamical systems and information theory in the search for evidence about the transfer of information from PAC to TNA SSTs. Our study will investigate the linkages between both oceanic regions with and without considering the presence of the Amazon River basin (AM) to shed light about the role of the Amazon River basin hydrology in the land-atmosphere bridge mechanism. We also aim to study the transfer of information from PAC to the six main sub-basins of the Amazon (Madeira, Solimoes, Tapajos, Xingu, Purus and Negro), as well as over the Andes and the low-lying portion of the Amazon River basin (Amazonia, hereafter). Our study aims to provide further insights about the physical processes involved in the dynamics of the land-atmosphere bridge connecting two of the most important oceanic basins for the tropical climate [Ham et al., 2013]. Towards those ends, we use a novel approach from
complex and non-linear dynamical systems, based on the framework of recurrence plots and the notion of Recurrence Measure of Conditional Dependence (RMCD), which allows inferring causality among dynamic variables [Ramos et al., 2017].

The paper is organized as follows: in section 2 we present the methods and data used for the analysis, in section 3 we show the results and discussions dealing with the Pacific to Atlantic feedback and the Atlantic to Amazon feedback (3.1), the information transfer from PAC to TNA through AM (3.2), information transfer from TNA to AM (3.3), direct information transfer from PAC to TNA, (3.4), information transfer from PAC to the major AM sub-basins and (3.5) information transfer from PAC to the Andes and Amazonia regions. Finally, in section 4 we provide concluding remarks of the present study.

4.2 Methods and Datasets

4.2.1 Recurrence Measure of Conditional Dependence

Some dynamical systems, including climate, present a recurrent behavior in the phase space, which constitutes a fundamental property of the systems [Marwan et al., 2007b]. This property can be easily visualized by the so-called Recurrence Plot (RP) [Eckmann et al., 1987]. In order to construct an RP from a time series, \( X = \{x_i: i=1, 2, ..., N\} \), it is necessary to represent the \( m \)-dimensional phase space of the underlying system \( X \). In case of a single time series the dynamics has to be artificially reconstructed using the time delay embedding technique [Packard et al., 1980; Takens, 1981], whereby the phase trajectories \( \vec{x}_i \) are defined as:

\[
\vec{x}_i = (x_i + x_{i+\omega}, ..., x_{i+(m-1)\omega}) \in \mathbb{R}^m \quad (4-1)
\]

where \( m \) is the embedding dimension and \( \omega \) is the time delay embedding. To determine the embedding dimension, \( m \), we use the method of false neighbors [Kennel et al., 1992], and to determine the embedding delay, \( \omega \), we use the mutual information function procedure [Fraser and Swinney, 1986; Marwan, 2011]. Once the parameters are set, the RP can be estimated as the pair-wise proximity test such that,

\[
R_{i,j}^X = \Theta(\varepsilon - \|\vec{x}_i - \vec{x}_j\|) \quad i, j = 1, ... N', \quad (4-2)
\]

where \( N' = N-(m-1)\omega \) is the number of phase space vectors, \( \varepsilon \) is the threshold defined for the proximity between the phase space vectors, \( \|\vec{x}_i - \vec{x}_j\| \) is the spatial distance between vectors in phase space, and \( \Theta(\ ) \) is the Heaviside function: \( \Theta(<0) = 0, (\Theta \geq 0) = 1 \). The plot of the \( R^X \) recurrence binary matrix provides the RP. The probability that one system recurs to a certain
state $\vec{x}_i$ is equal to the column-average of the recurrence matrix [Marwan et al., 2013]:

$$p(\vec{x}_i) = \frac{1}{N} \sum_{j=1}^{N} R^X_{i,j}$$  (4-3)

Joint recurrence plots (JRP) are used to study the possible influence between two physically different systems [Romano et al., 2004; Marwan et al., 2007b], as they provide a measure of the simultaneous recurrence. The JRP matrix is defined as the Hadamard product of the RPs of systems $X$ and $Y$:

$$J R^X_{i,j} = \Theta(\varepsilon - \|\vec{x}_i - \vec{x}_j\|) \times \Theta(\varepsilon - \|\vec{y}_i - \vec{y}_j\|) i,j = 1, \ldots, N'. \quad (4-4)$$

The probability of finding a simultaneous recurrence at time $i$ in both systems $X$ and $Y$ is equal to the column-average of the JRP matrix:

$$p(\vec{x}_i, \vec{y}_i) = \frac{1}{N} \sum_{j=1}^{N} J R^X_{i,j}. \quad (4-5)$$

It is possible to calculate the conditional probability of the system $X$ recurring conditioned to $Y$ in a given time $i$ such that:

$$p(\vec{x}_i | \vec{y}_i) = p(\vec{x}_i, \vec{y}_i) / p(\vec{y}_i) = \frac{1}{N} (\sum_{j=1}^{N} J R^X_{i,j} / \sum_{j=1}^{N} J R^Y_{i,j}). \quad (4-6)$$

The concept of Transfer Entropy [Schreiber, 2000; Runge et al., 2014] is used to assess the recurrence relation between two variables by excluding the past self-influence of the driven variable and thus inferring causality. This expansion in the concept of conditional recurrence is called Recurrence Measure of Conditional Dependence (RMCD), and it quantifies the recurrence between systems $X$ and $Y$ given $Z$, which has proven to be accurate in paradigmatic models such as coupled ARMA models, coupled Logistic Maps and coupled Lorenz systems [Ramos et al., 2017], defined as:

$$RMCD (X, Y | Z) = \frac{1}{N'} [ p(\vec{x}_i, \vec{y}_i, \vec{z}_i) \log \left( \frac{p(\vec{x}_i, \vec{y}_i | \vec{z}_i)}{p(\vec{x}_i | \vec{z}_i) p(\vec{y}_i | \vec{z}_i)} \right) ] \quad (4-7)$$

With $RMCD$ is possible to quantify the causal dependence of system $X$ on system $Y$ based on the joint recurrence between the past of the driver system $X^T$ and the present of the driven system $Y$, discarding the past contributions $Y^T$ such that:

$$RMCD (X^T, Y | Y^T) = \frac{1}{N'} \sum_{j=1}^{N} \left[ \frac{1}{N'} \sum_{j=1}^{N} J R^X_{i,j} \cdot J R^Y_{i,j} \log \left( \frac{\sum_{j=1}^{N} J R^X_{i,j} J R^Y_{i,j}}{\sum_{j=1}^{N} J R^X_{i,j} \sum_{j=1}^{N} J R^Y_{i,j}} \right) \right] \quad (4-8)$$

Where $\tau$ represents the lag by which the system is shifted back in the past, and analogously to other recurrence based measures [Goswami et al., 2013]
$RMCD$ is nonnegative, in particular $RMCD=0$ when $Y^T=X^T$ or $Y^T=Y$, and when $X^T$, $Y$ and $Y^T$ are mutually independent.

### 4.2.2 Significance testing

For a finite hydro-climatic time series we have to rely on a null hypothesis test to define the statistical significance of the $RMCD$ measure in determining the possible causal relation between systems $X$ and $Y$. The null hypothesis assumes that all trajectories in the embedding space are independent realizations of the system, with different initial conditions. The statistical significance is tested then using a twin surrogate hypothesis [Thiel et al., 2006, 2008], which produces a number of surrogates or copies defined as $Nsurr$ with the same dynamical properties as the original sample, but with a different recurrence structure. The 99% percentile of the distribution of the surrogates $RMCD$ values is defined as the confidence interval.

The $RMCD$ value of the original set of time series is then compared with its respective confidence interval (threshold) at all lags $\tau$. If the RMCD is higher than the confidence interval, we reject the null hypothesis, i.e. the variables are not independent with respect to the surrogate test with significance 0.01. Otherwise, the hypothesis is accepted, meaning that the variables are independent in the recurrence sense. Therefore, rejection of the null hypothesis for a particular lag $\tau$ indicates a causal interaction at the time scale $\tau$. Finally, we perform a multiple comparison analysis (M.C.A.) between all lags investigated using the Dunn-Sidák test [Šidák, 1967]. The significance for comparison, $a=0.001$, yields a family-wise error rate around 0.03.

### 4.2.3 Data

We defined three main geographical regions to study the role of the land-atmosphere bridge over the Amazon River basin (AM) in connecting PAC and TNA (Fig 4-2). The regions are NIÑO 3.0 ($90^\circ$W-$150^\circ$W and 5°S- 5°N ), TNA ($75^\circ$W to $10^\circ$W and 5°S to 29°N) and the Amazon River basin that comprises 146 sub-catchments as defined by the Observation Service SO-HYBAM (formerly Environmental Research Observatory ORE-HYBAM available at http : //www.ore – hybam.org/). To those three main regions we added the six major sub-basins of the Amazon River: Madeira, Solimoes, Tapajos, Xingu, Purus and Negro, as defined by ORE-HYBAM, as well the Andes and Amazonian regions of AM.
Daily precipitation data for the whole AM and its main sub-basins, and over the Andes-Amazonia regions were obtained from the Tropical Rainfall Measuring Mission (TRMM) whose product 3B42 provides satellite measured precipitation corrected with rain gauge information \cite{Huffman, 1997; Huffman et al., 2007, 2010}. Daily time series are averaged for the regions within the Amazon River basin, and daily anomalies were computed with respect to the climatology from 1998 to 2014. Daily SST data were obtained from the NOAA OI SST High Resolution Dataset with spatial resolution of $0.25^\circ\times0.25^\circ$ and spanning from 1985 to 2014 \cite{Reynolds et al., 2007}. Daily time series were averaged over the NIÑO 3 and TNA regions. The SST anomalies were obtained removing the seasonal effect, in turn estimated from the climatology from 1982 to 2014. Daily river flows time series at Óbidos (last river gauging station before the Amazon River delta), was obtained from SO-HYBAM and spans from 1968 to 2011. River flows anomalies were computed with respect to the 1968 to 2010 climatology. We performed a 3-day filter to averaged time series in order to remove the synoptic effect present at daily time scales.

### 4.3 Results and discussion

To gain further insights about the role of the land-atmosphere bridge connecting the PAC and TNA regions through the Amazon River basin, we focused our analysis on two once-in-a-century droughts in the Amazon River basin in 5 years \cite{Lewis et al., 2011b}, one mainly driven by the Atlantic Ocean (2005) and the second one related with an El Niño episode (2010) \cite{Marengo et al., 2008, 2012; Zeng et al., 2008; Lewis et al., 2011b; Espinoza et al., 2012; Sena et al., 2012}. We tested up to 210 day lags (7 months) to evaluate the influence of SSTs on precipitation over land. The rationale behind the use of those two contrasting years is to evaluate with one experiment not only the possibility of a non-linear causal connection between both oceanic regions through the Amazon River basin, but also the capability of the \textit{RMCD} measure to pinpoint the differential influence over the Amazon River basin during the 2005 and 2010 droughts. For estimation purposes, we used the following embedding parameters: $m = 3$, $d = 15$, and $\epsilon$ is 20\% the size of the phase space, based on a criterion of maximizing the RMCD value as proposed.
and used in Ramos et al., [2017] to study the coupling between PAC and Southwest Amazonia.

This section is organized in four subsections presenting results and analysis about the transfer of information: (1) from the Tropical Pacific to Amazon River basin to Tropical North Atlantic (PAC→AM→TNA), and from the Tropical North Atlantic to the Amazon River basin (TNA→AM); (2) from PAC→TNA without the mediation of the Amazon land-atmosphere bridge; (3) the PAC to the main sub-basins of AM: and (4) from the PAC to the Andes and Amazonia. Results will be presented with a schematic diagram illustrating the significance of the results. Fig. 4-3 shows two panels with the results obtained with the RMCD analysis; the upper panel shows in red the time evolution of the RMCD values for each one of the lags (days), and in grey line the confidence interval constructed with the surrogates. The bottom panel shows the results of the significance test, with red dots depicting the lags where the RMCD values cross the confidence interval (rejecting the null hypothesis, thus denoting transfer of information and causality), and with blue dots representing the significant lags according to the Dunn-Sidák M.C.A test. The example in Fig. 4-3 shows the information transfer between PAC and AM according to the RMCD analysis for the 2010 drought; red dots denote the lags of significant information transfer, for this particular example in 55 out of 210 lags analyzed (26%), and blue dots denote the lags for which the M.C.A test indicate a positive result, 17 out of those 55 lags (30%). For the sake of simplicity, the results section will be presented only with the panel showing the significant results (Fig. 4-3, bottom).

Figure 4-3: Example of the results obtained with the RMCD analysis. The upper panel shows the RMCD values (red), and the confidence interval (grey). The red dots in the bottom panel denote those lags for which the RMCD values cross the uncertainty threshold, for the relation between the Pacific Ocean SST anomalies and rainfall anomalies over the Amazon River basin during 2010, thus rejecting the null hypothesis which denotes a significant transfer of information or causality. The blue dots that represent the lags with a significant value of the M.C.A test.
### 4.3.1 Information Transfer from the Pacific Ocean to the Tropical North Atlantic through the Amazon River basin

In this section we investigate the role of the Amazon River basin as a land-atmosphere bridge connecting in connecting the PAC and TNA during the 2005 and 2010 droughts. First, we study the pathway between PAC and AM, and second the pathway between AM and TNA. Fig. 4-4b shows results of the RMCD analysis on the information transfer from PAC to AM during 2005 and 2010. During 2005 there is significant information transfer from PAC to AM between 30 and 40 day-lags (1 to 2-month), and from 150 and 210 day-lags (5 to 7-months), although the significant RMCD behave intermittently (Fig 4-4b, top), during 12% of the lags. During the 2010 drought the influence of PAC over AM is more intense and almost steady from 150 to 210 day-lags (5 to 7-months) (Fig 4-4b, bottom), with crossings at 26% of the lags. The information transfer form PAC to AM in 2010 is not only more intense, but also more consistent, as it gets through both the confidence interval crossing and the M.C.A., as evidenced in Fig. 4-4b (bottom panel) by a larger number of blue dots.

Results of the M.C.A test show that 30% of the crossings pass both tests. Accordingly, during both 2005 and 2010 droughts there is information transfer from PAC SSTs to AM precipitation, the influence being stronger during the occurrence of the 2010 El Niño event, with a 5 to 7-months lag. The timing of the influence from PAC to AM is in concordance with the well-known dynamics El Niño, which starts in MAM and reaches its maximum extent and anomalies 5 to 7 months later in DJF, and also with the timing of the influence of the particular 2010 event on AM [Marengo et al., 2011; Espinoza et al., 2012].

Fig. 4-4c shows the results of the RMCD analysis for the information transfer from AM precipitation to TNA in 2005 and 2010. For 2005 (Fig 4-4c, top) there are some individual crossings during the first month’s lags, and significant crossings from between 180 to 210 day-lags (6 to 7 months), that might be related with the response of PAC to AM perturbations starting at 150 day-lags (5 month lag, Fig. 4-4b, top), and a propagation towards the TNA (Fig 4-4c, top), although the confidence bands are crossed only at 8% of the lags. The panorama is quite different during the 2010 drought, with is a strong indication of AM feedback over the TNA that spans from 5 to 180 day-lags (0 to 6 months), covering 30% of lags, and being more significant between 120 to 150 days (4 to 5 months), according to the M.C.A. test (Fig 4-4c, bottom).

Fig. 4-4d shows the results of the RMCD analysis for the information transfer from AM river flows to TNA SSTs in 2005 and 2010. The rationale behind the use of river flows is that this variable acts a physical and mathematical filter of the high frequency variability inherent to rainfall data, and therefore it summarizes all land-surface processes taking place within river basins. As a
matter of fact, the land-atmosphere bridge theory that links PAC and TNA was put forward by Poveda and Mesa, [1997] using Andean river flows. For 2005 there is no information transfer from AM river flows to TNA (Fig 4-4d, top), given the lack of RMCD crossings or positive results from any of the significance tests, whereas in the case of the 2010 drought the crossings of the confidence interval represents 30% of the lags and the information transfer is strong from 60 to 120 day-lags (2 to 4 month-lags), and have some intermittency in between 200 to 210 day-lags (the 7 month-lag, Fig. 4-4b bottom).

Our results indicate that during the 2010 El Niño event AM was more active in transferring information to the TNA. As for the 2010 drought, the influence of AM over TNA tends to be strengthened through an increase of air temperature in AM induced by the displacement of the Walker circulation in association with El Niño in PAC, and also activating diverse processes involved in the feedback mechanism recently proposed in chapters 2 and 3, whereby changes in convection in the AM may induce the warming of the TNA SSTs due to changes in the surface atmospheric pressure gradient, which in turn disrupts the patterns of moisture advection to the AM basin.
4.3.2 Information Transfer from the Tropical North Atlantic to the Amazon River basin

Fig. 4-5 shows the results of information transfer from the Tropical North Atlantic (TNA) to the Amazon River basin rainfall (AM) (panel a). For 2005 and 2010 the RMCD measure captures an evident effect of the TNA on AM throughout the entire year. Results regarding the crossing of the threshold are from 40 to 210-day lags (2 to 7-month lags) (Fig. 5-5b, red dots in top and bottom panels). These results agree with the well-known fact that the TNA is of utmost importance to influence the AM hydroclimatic regime [Hastenrath, 2000a; Marengo, 2006; Nobre et al., 2009; Yoon and Zeng, 2010; Moraes-Arraut et al., 2011; Gimeno et al., 2012; Poveda et al., 2014; Arias et al.,]
During the 2005 drought we observe more indications of significance than during 2010 along the studied lags, with less scattered red dots and crossings of the confidence interval covering 28% of the lags (Fig. 5-5b, top), while those found in 2010 only cover 16% of the lags, which means that the TNA→AM feedback was much stronger in 2005. The drought of 2005 has been attributed mainly to an anomalous warming of the TNA [Marengo et al., 2008; Zeng et al., 2008; Marengo and Espinoza, 2016], and our results confirm more information transfer during the 2005 drought than the 2010 one which has been attributed mainly to PAC, thus supporting the TNA origin of the 2005 AM drought.

Figure 4-5: Same as Figure 4-4 for the transfer of information from the Tropical North Atlantic (TNA) to Amazon River basin rainfall.

4.3.3 Information Transfer from the Tropical Pacific the Tropical North Atlantic without mediation of the Amazon River basin

In order to test the transfer of information from PAC to TNA without including the proposed land-atmosphere bridge acting on the Amazon River basin, we also carried out an RMCD analysis between the SSTs time series on both oceanic regions, whose results are shown in Fig. 4-6. When AM is removed, we observe a more active influence of the PAC over the TNA in 2010 (44% of lags) than in 2005 (10% of lags), although there is evidence of information transfer during both years. In 2010 the signal starts to be significant from 60 to 210-day lags (2 to 7-months lags) (Fig. 4-6b, bottom), while for 2005 the
signal is intermittent between 120 to 180-day lags (4 to 6-month lags) (Fig. 4-6b, up). According to the M.C.A. test, there is more confidence in 2010 than in 2005 about the influence of PAC on TNA.

Up to this point our results provide further support about the role of the land-atmosphere bridge acting on the Amazon River basin to enhance the influence of the tropical Pacific over the Tropical North Atlantic. Diverse studies have suggested a direct atmospheric pathway from PAC to TNA [Ngar-Cheung and Mary, 1996; Latif, 2001; Roy and Reason, 2001] at play during El Niño events (Fig. 4-5a). For ENSO neutral years as 2005 there is evidence of an influence from the Pacific Ocean to the Atlantic Ocean and South American climate [Berri and Bertossa, 2004; Kajtar et al., 2016], although it is not so easy to estimate and disentangle the Pacific’s from other influences as the one from the Indian Ocean. ENSO events force the TNA SSTs via the weakening of the northeasterly trade winds as a result of the vertical stabilization of the tropical atmosphere related to a tropic-wide warming [Giannini et al., 2001; Rodrigues et al., 2011]. A recent study has found evidence of an indirect path through the continent due to the influence that the AM has over the TNA, thus supporting the existence of the land-atmosphere bridge and also the presence of an AM→TNA physical mechanism [Chapter 3].

Figure 4-6: Same as Figure 4-4 for the influence of the Pacific Ocean to the Tropical North Atlantic.

Results presented until this point indicate that both processes have a significant influence on TNA’s interannual dynamics (crossings of the confidence interval and M.C.A. test): the direct atmospheric pathway from
Chapter 4

PAC→TNA and the indirect PAC→AM→TNA mediated by the Amazon River basin surface processes (precipitation and river flows).

4.3.4 Information Transfer from the Tropical Pacific to Major Amazon River Sub-Basins

This section quantifies the transfer of information from PAC to the major sub-basins of the Amazon River. It is well-known that different sub-basins exhibit distinctive hydrological patterns depending on their location, land cover and land use change (deforestation). There are well-known north-south hydrological differences in the Amazon River basin [Marengo, 2004, 2009; Carmona et al., 2016], as well in the behavior of the droughts and lengths of dry periods depending on location in the AM region [Fu et al., 2013; Yin et al., 2014; Zou et al., 2015; Espinoza et al., 2016]. In order to understand the way in which PAC may affect the major sub-basins of the AM we computed the RMCD between the PAC SSTs and precipitation in the six major AM sub-basins: Madeira, Solimoes, Tapajos, Xingu, Purus and Negro, whose results are presented in Fig. 4-7.

Fig. 4-7b shows the RMCD analysis for the transfer of information from PAC to each one of the sub-basins in 2005. Results show that the Xingu River basin receives a significant influence from PAC (10% of the crossings) at 40 and 60 day-lags (2 and 3-month lags) (Fig. 4-7b, sixth row). The Tapajos River basin (Fig. 4-7b fifth row) has an early influence of PAC in 2005 (11% of the lags) like the one found for Xingu but it lasts almost up to 180-day lags (month 6th). The Madeira River basin (Fig. 4-7, fourth row) has a strong signal of influence between 110 and 180-days lags (4 and 6-month lags) for 2005 (14% of the lags). The Purus River basin (Fig. 4-7b, third row) for the drought of 2005 starts around the 50-days lag (2-month lag) and goes intermittently until 210-day month lag (7-month lag) (15% of the lags). The Solimoes River basin (Fig. 4-7b, second row) receives a strong influence from PAC around 40 and 60day lags (2 and 3-month lags), and later on from 110 to 180-day lags (4 to 6-month lags) (14% of the lags). For the Negro River basin (Fig. 4-7b, first row) the influence of PAC is found from 120 to 180-day lags (4 to 6-month lags) (15% of the lags).

Fig. 4-7c shows the results of the RMCD analysis for the transfer of information from PAC to each one of the major AM sub-basins in 2010. A remarkable consistency of the influence of PAC on all sub-basin is observed throughout the lags for 2010. For instance, the Xingu, Tapajos, Madeira and Negro river basins receive the influence from 150 to 210-day lags (5 to 7 months-lags) (with 11%, 17%, 29% and 15% of lags, respectively). The influence on the Solimoes River basin also starts earlier that in the other basins, from 110 to 210-day lags (4 to 7-month lags) (24% of the lags). Finally, the Purus River basin receives a lagged influence from 140 to 180-day lags (4 to 6-month lags) (8% of lags). The early signals of coupling during 2010 and the percentage of crossings of the confidence interval in that same
event in the *Madeira* River basin are consistent with previous studies showing that the 2010 drought was especially strong over southwestern AM [Marengo *et al.*, 2011; Zou *et al.*, 2015].

These results on the RMCD measure allow us to conclude that during the 2010 El Niño event the influence of PAC on the major Amazon sub-basins was stronger from 5 to 7-month lags, regardless of the location of the sub-basin. This finding is in tune with the timing of ENSO dynamics and with our previous results presented in section 3.1. Also, PAC also had an influence on the major sub-basins during the 2005 drought, though with a different timing that cannot be generalized for all the basins as it was found for 2010. The particular drought of 2005 shows a distinctive feature reflected in a temporal gap in the PAC influence from the end of month 2 that lasted almost one month, perfectly visible for 5 of the 6 sub basins (Fig. 4-7b). The way and the timing in which PAC transfer information on to each of the major sub-basins of the Amazon River are mediated not only by location and land cover, but also by the influence on the Walker circulation and by the teleconnections from PAC to South America via wave trains [Marengo *et al.*, 2002; de Drumond and Ambrizzi, 2008].

Figure 4-7: Same as Figure 4-4 for the influence of the Pacific Ocean to the six major sub-basins in the Amazon River basin, (a) 2005 and (b) 2010.

Regarding the information transfer from PAC to TNA, one interesting feature about the RMCD results for the 2010 drought is the increasing pattern in the RMCD values that starts around the 150th day lag (Fig. 4-6b, red dots). On
the same token, Fig. 4-8 shows the evolution of the \( RMCD \) values in the 2010 event up to 210-day lags (7 months) for the six major AM sub-basins. For all sub-basins the values of \( RMCD \) start to cross the confidence interval around the 150-day lag, and for 5 of them (Xingu, Purus, Solimoes, Tapajos and Negro) the increasing trend starts around the 100-day lag, while for Madeira the trend starts around the 10-day lag. Other characteristic of the \( RMCD \) values is that after reaching the point where the confidence interval is crossed the metric gets stable (with no trend). Such increasing trends before reaching the confidence level in the \( RMCD \) values for the 2010 event suggest the need of further investigating the behavior of the measure in other ENSO events. Such results may be pretty valuable in order to propose \( RMCD \) as a robust early warning system ancillary measure in the Amazon River basin.

![Figure 4-8: Time evolution of the RMCD values during the 2010 drought from 0 to 210-day lags (black line), and confidence interval (red dashed line), regarding the information transfer from PAC to the major AM sub-basins. Crossings of the confidence interval are evident around 150th month-lag.](image)

**4.3.5 Information Transfer from the Tropical Pacific to the Andes and Amazonia regions**

In this section we study the information transfer from PAC to the Andes, defined as the portion of the Amazon River basin that comprises regions located above 500 m a.s.l., and to the low-lying Amazonia, whose results are shown in Fig. 4-9. During the 2005 drought (Fig. 4-9b), the influence of PAC on both regions has a distinctive timing in the second month and a second strong influence during the fourth and fifth months. This temporal gap in the
influence of PAC in 2005 is similar to the one found in the PAC to sub-basins analysis (section 3.3). The second period of influence starts earlier in the Andes, and it may be explained in terms of the geographical proximity of Andes to the Pacific Ocean. The influence found in 2005 covers 16% of the lags for Amazonia and 12% of the lags for the Andes.

On the other hand, for the 2010 drought (Fig. 4-9b) the influence of PAC shows up after the fourth month and more significantly from 5 to 7-month lags. The signal of significant information transfer over the Andes reaches up to 30% of the lags (40% of which passes also the M.C.A test); while for Amazonia it represents 17% of the lags. During the 2010 drought, PAC exerts a larger influence on the Andes, and the information transfer starts almost one month earlier than in Amazonia, such as in the 2005 drought. The Andes may represent only the 13% of the total AM basin, but the region is the most important source of surface water (runoff), nutrients and sediments to the whole basin [Poveda et al., 2006; McClain and Naiman, 2008; Espinoza et al., 2015]. Our results indicate that the Andes receive stronger influences from El Niño (30% of lags have a significant signal) starting one month earlier that in the low lying Amazonia. The cascading of the ENSO influence on both regions have been already reported and analyzed as one of the most distinctive features of the 2010 drought [Espinoza et al., 2011]. As it was found for the PAC to sub-basins analysis, there are increasing trends in the RMCD values for the information transfer from PAC to Andes and to Amazonia regions, that taper off around the 150th day lag (Fig. 4-10).

Figure 4-9: Same as Figure 4-4 for the information transfer from PAC to the Andes and Amazonia regions, during 2005(a) and 2010 (b).
Figure 4-10: Same as Figure 4-8 for the information transfer from the Pacific Ocean to the Andes and Amazonia regions.

4.4 Summary and Conclusions

Using a novel approach to quantify causality between hydro-climatic systems we have provided further support to the existence of a land-atmosphere bridge that connects the Tropical Pacific and the Tropical North Atlantic, and disentangled the different aspects of the role of the Amazon River basin in such a bridge during the two extreme droughts of 2005 and 2010. The bridge mechanism is captured with the causal analysis when using two different hydrological processes in the Amazon River basin (precipitation and river flows). Our results confirm the presence of information transfer from the AM basin to the TNA, which is intensified by the impact of the El Niño event of 2010 in the hydrological regime of the Amazon River basin.

The complex interaction between the oceans through the Amazon River basin was found to be significatively active during the 2010 El Niño event, whereas in 2005 is intermittent and was significant only through the precipitation in the AM. Our results also show an active transfer of information from the Amazon River basin to the Tropical North Atlantic, which confirms that Amazon hydrology feedback on the Tropical North Atlantic climate variability. With the causal measure we were able to disentangle the importance of PAC and TNA in the 2005 and 2010 historical droughts of the Amazon River basin. Our results
are in agreement with previous studies that pointed out TNA as a major influence in the 2005 drought and PAC in the 2010 one.

We have also analyzed the information transfer from the Tropical Pacific to the major AM sub-basins, and also to the Andes and Amazonia regions. Our results show that the information transfer from PAC to these regions varies regionally, and exhibits different timings for the 2005 drought event. With regards to the 2010 drought, the pattern of coupling between PAC and the regions inside the Amazon is characterized by a higher significance, given by an increase in crossings of the confidence interval and positive values of the M.C.A test, starting around the 150-day lag (5 months). This consistent signal during an El Niño event shows the classical timing of El Niño events reaching their peak 5 to 7 months after the beginning of the anomalous warming of the Pacific.

Results are showing increasing trends in the RMCD values in the presence of a strong climate forcing as El Niño. These kinds of patterns may suggest a further use of the causality measure to develop or to test early warning systems for extreme ENSO events. An early signal of information transfer between PAC and Andes may be useful for the communities that rely on water supply from the high lands of the basin, or prone to floods communities. In order to achieve such goal is necessary to use the measure to assess previous El Niño and La Niña events to assess the full capability of the causal measure in disentangling the influence exerted over the Amazon River basin.
5. General conclusions and future work

This thesis has explored the hydroclimatic interrelations between the Andes, the Amazon River basin and the Tropical North Atlantic. Particularly, we have estimated the separate and conjoint surface and atmospheric water budgets in the entire Amazon River basin, its major sub-catchments, as well as disentangling the Andean and low-lying portions of the Amazon River basin. Particularly, we provided new insights regarding the spatial patterns of the water budgets closure in the basin, and have quantified the two-way interactions and coupling existing between the Andean and low-lying regions of Amazonia.

With the new knowledge gained in this research about the surface and atmospheric branches of the hydrological cycle in the Amazon River basin we advance our understanding of the Amazon hydrology as a key player in the tropical hydroclimate by studying its complex relation with the Tropical North Atlantic and the Tropical Pacific Ocean. In consequence, this dissertation proposes a suitable mechanism to explain how the Amazon River basin hydroclimatology plays a key role in the modulation of the Tropical North Atlantic SST’s and provides further evidence about the role of the basin in the connection of the Tropical Pacific and Tropical North Atlantic at interannual timescales.

In Chapter 1 we advance our state of knowledge regarding the surface and atmospheric water balances in the Amazon River basin using five datasets of evapotranspiration, two of precipitation, two of runoff and one of vertically integrated moisture flux divergence. Our results provided a state of the art conjoint analysis of surface and atmospheric balances using the most comprehensive and extensive available data sets. One of the major contributions of this chapter is the analysis of the surface and atmospheric water budgets and their (im)balances at three different spatial scales: the entire Amazon River basin, its major sub-catchments and the Andes-Amazonia regions. We can conclude that the results for each spatial scale are highly dependent on the dataset used. For instance, when using observational datasets the surface water balance (SWB) for entire basin can be consider to be balanced, while only 50% of the major sub-basins are in balance; and as for the results when using the reanalysis dataset the percentage of sub-basins in balance reach up to 92%. This result suggests that for observational datasets there is higher uncertainty at the sub-catchment scale.

The sub-catchment analysis (in chapter 1) showed differences in the sign of the water balance closure according to the location of the catchment in the basin, presenting distinctive regional patterns. Also, the sign of the closure in the catchments is independent of the sign found for the closure in the entire
General conclusions and future work

basin. There is consistency in the regional patterns of the SWB when using observational datasets of evapotranspiration and high discrepancies when results are compared with the ones obtained with reanalysis dataset. In the case of the atmosphere water balance (AWB) we found no closure in any of the dataset and a pattern of positive residuals for all the regions studied. The previous results lead us to conclude that the modeling of convergence in reanalysis still lacks of a better representation of surface fluxes in the Amazon River basin that are highly influenced by forest activity and ecosystems.

Regional closures are highly dependent on the Evapotranspiration dataset used (observational or reanalysis), but for observational datasets of Evapotranspiration patterns of closure are similar in the sub-catchment scale. The previous results indicate that is still a lot of work in the field of direct measuring of Evapotranspiration that have to be accounted for. In the studied regions, satellite derived datasets of Evapotranspiration are in disagreement with the reanalysis dataset, justifying the need of improvement in the assimilation and modeling of this variable in regions highly influenced by biogeophysical flows as the ones from the Amazon and Andean forests.

We also propose a corrected form for the imbalance equation proposed by Marengo [2004] to introduce the rates of storage of precipitable and soil water in order to attempt a long-term analysis of the imbalance between water budgets. Results of the corrected imbalance exhibit a pattern of negative imbalance in the basin related to a deficit in the representation of convergence in reanalysis, that leads to a constant pattern of \( R > C \) during the entire year, that increases during the Amazon dry season. Finally, from the Andes-Amazonia water budgets analysis we found evidence of the importance that Andes has for the entire basin as a source of surface water, being the region with the most runoff for unit area.

Results shown in Chapter 1 provide new insights about surface and atmospheric water flows in the Amazon River basin, presenting a broad set of new results regarding the imbalances of both budgets and its regional patterns. Our results support the hypothesis of the importance that surface runoff from Andes has in the entire basin [Poveda et al., 2006; McClain and Naiman, 2008] and that the analysis by regions and sub-basins makes it possible to verify and to better understand the heterogeneity of the hydrological behavior of the basin, reported also by different authors [Marengo, 2004; Zhang et al., 2008; Zou et al., 2015; Carmona et al., 2016]. The recognition of the Andes as a key region to maintaining the stability of the Amazon River basin is an important stepping stone to propose the development of new efforts to collect high-quality information in the Andes, where information is not as reliable as the one available for the low-lying Amazon.

In chapter 2, we have explored the linear relationships between the hydroclimatology of the Amazon River (AM) basin (represented by the \( P-E \) index) and the Tropical North Atlantic (TNA) SSTs using lagged-correlations,
building up upon the results of *Poveda and Mesa* [1997]. We have found significant correlations between the two regions at seasonal and interannual timescales. In summary, the feedback relates the increase (decrease) of convective processes in the Amazon River basin with a decrease (increase) in the TNA SSTs. Seasonal and interannual analysis showed that the feedbacks between AM and TNA develops in a timeframe of 0 to 2 months. At seasonal timescales, more than 60% of the TNA responds to the influence of the AM, while this percentage is of around 45% at interannual timescales.

Analysis of the dynamical response over the TNA SSTs during the wet (dry) seasons in the AM shows a concomitant influence of the continental region over the oceanic one represented in an increase (decrease) of zonal winds and horizontally integrated water vapor transport that are therefore related to a cooling (warming) of the TNA SSTs. At interannual timescales, the dynamical analysis was carried out during extreme events of precipitation in the Amazon. The analysis of the time evolution of total winds and SSTs over the TNA two months before and after the extreme events confirmed the time lapse of the feedbacks to be within 0 to 2 months after the peak of convection anomalies over the AM River basin.

One of the main contributions of the linear analysis presented in Chapter 2 is the proposal of a physical mechanism that connects the AM hydrology and TNA SSTs through the lowest atmospheric levels. The mechanism connecting the two regions involves diverse processes, as follows: the Amazon driest (wettest) months/events are characterized by a decrease (increase) in rainfall, which in turn increases (decrease) atmospheric surface pressure in the AM region thus reducing (increasing) the pressure gradient between the TNA and AM regions, which in turn causes a slowing down (speeding up) of the zonal trade wind velocities over the TNA region. The weakening (strengthening) of the trade winds over the TNA is related with a reduction (increase) in evaporative cooling and a subsequent increase (reduction) of TNA SSTs. The analysis of the proposed mechanism refined into a pentad or daily timescale in order to have further insights about its exact time development is proposed for future research.

Results from the linear analyses presented in chapter 2 provided the stepping stone to propose a nonlinear approach to disentangle the physical mechanism drawn in chapter 3. We used a recurrence based methodology to evaluate the pairwise relationship between the variables involved in the mechanism connecting the AM and TNA at interannual timescales. One of the major contributions of the chapter was to provide evidence on how the AM hydrology influences future states of the TNA SSTs from 0 to 2 months in advance, with a 90% to 95% statistical confidence, confirming the statistical and dynamical results from chapter 2. At the interannual timescale there are forcings such as ENSO and the Atlantic Meridional Mode that have a strong influence on TNA SSTs and that may act together with the proposed mechanism.
In chapter 3 we verified that the proposed mechanism was set in place during several extreme dry and wet events over the Amazon River basin that indeed triggered the mechanism and the concomitant changes in SST’s over the TNA. For instance, we found that the influence of the AM hydrology over the TNA SSTs during January occurs in a timelapse of one month, almost in phase with the extreme events of precipitation. Another relevant result is the fact that extreme precipitation events over the AM may influence the TNA SSTs regardless of the occurrence of a major forcing such as ENSO. From chapters 2 and 3 there is extensive evidence of two-way feedbacks between the AM and the TNA at seasonal and interannual timescales, leading us to conclude about the importance that the Amazon River basin has for the SSTs stability in the TNA and therefore in other processes such as the formation of tropical storms, hurricanes and even for the uptake of moisture during the Amazon wet season to maintain processes such as the South American Low Level Jet.

The new insights introduced in the present study, regarding the physical mechanism that connects the Amazon River Basin and the Tropical North Atlantic, enhance the ongoing notion of the Amazon as a key player in the modulation of the TNA as already reported by other authors at seasonal and interannual timescales [Poveda and Mesa, 1997; Wang and Fu, 2007a; Yoon and Zeng, 2010; Misra and DiNapoli, 2012]. Our results confirm the importance of the Amazon River basin on the stability of the tropical climate, and open up new avenues of further research about how changes in land use and deforestation may disrupt the convection patterns in the Amazon their impacts on the TNA SSTs. Recent studies have found the relationship between deforestation and alterations in the uptake of moisture of the Amazon River basin from the TNA during the South American Monsoon System [Boers et al., 2017]. According to our results, the effects of reduced convection over the AM during the Monsoon can in turn contribute to an increase in the TNA SSTs. Distinguishing the future changes in convection associated to land use change in the Amazon may be the next step to model the changes in the physical mechanism exposed in this research.

This thesis has proposed a physical mechanism that connects AM and TNA, and found evidences of this mechanism acting at interannual timescales. Ever since the work of Poveda and Mesa [1997] we are aware of the hypothesis that AM may act as a bridge connecting the Tropical Pacific and Tropical North Atlantic oceans at interannual timescales, thus we keep on analyzing the complex interrelation between the three regions by means of a nonlinear analysis in chapter 4.

During the development of this thesis, we made part of a research team that made an important contribution in the field of complex systems and nonlinear analysis through the development of a new approach for the analysis of lagged causal information transfer between complex systems, in collaboration with researchers from the National Institute for Space Research in Brazil and from the Potsdam Institute for Climate Impact Research in Germany. The new measure called Recurrence Measure of Conditional Dependence (RMCD) is
based on the recurrence plots framework and was used to test the hypothesis of an atmosphere-land bridge through the Amazon River basin that connects the Tropical North Atlantic and Tropical Pacific.

Using the \textit{RMCD} measure we analyzed the two once-in-a-century droughts of 2005 and 2010 in the Amazon River basin [\textit{Marengo et al.}, 2008, 2011; \textit{Lewis et al.}, 2011b; \textit{Marengo and Espinoza}, 2016]. We choose those events with the aim of analyzing the patterns of information transfer from the Tropical Pacific Ocean to the Tropical North Atlantic through the Amazon River basin during an El Niño year (2010) and a ENSO-neutral year (2005). The measure was able to pinpoint significant information transfer from the Pacific Ocean to the Amazon River basin in both episodes. Also, we found a significant information transfer from the Amazon River basin to the Tropical North Atlantic in 30% of the lags analyzed during the El Niño related drought (2010), while for the neutral year drought (2005) this percentage was identified only in 8% of the lags. Thus, our results provide further evidence supporting the existence of a land-bridge between the two oceans through the Amazon River basin during El Niño event of 2010.

One of the most interesting results from chapter 4 is that with the \textit{RMCD} approach we were able to differentiate the macroclimatic forcings associated the two studied droughts. In the 2005 drought, we found a significant information transfer from the Tropical North Atlantic to the Amazon River basin, while for the event of 2010 the significant influence comes from the Pacific Ocean. Our results are in agreement with the work of \textit{Marengo and Espinoza} [2016], which accounts the forcings causing the reported droughts and floods in the Amazon River basin. The measure also exhibits differences in the timing of the information transfer for different regions within the Amazon. For instance, the Madeira River basin was found to get the information from the Pacific earlier than the rest of the main sub-basins in the Amazon. Such an early influence in southern Amazonia was also reported by \textit{Zou et al.}, [2015]. Other remarkable result is the presence of increasing trends in the \textit{RMCD} values during El Niño event in 2010, which might be used as an early warning system for droughts to the communities that depend on the Amazon.

Finally, it is important to acknowledge that the the causal analysis presented in chapter 4 is only for two particular extreme dry events. We consider that the measure may be used for the analysis of different El Niño, La Niña and neutral years in order to test for average patterns of the information transfer between the Pacific Ocean and the regions within the Amazon.
## A. Appendix A: Chapter 1

Table A1. Comparison between previous surface water budget studies at Óbidos in the Amazon. Adapted from [Carmona, 2015]

<table>
<thead>
<tr>
<th>Study</th>
<th>P (mm/year)</th>
<th>E (mm/year)</th>
<th>R (mm/year)</th>
<th>P-E-R (mm/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>[Baumgartner and Reichel, 1975]</td>
<td>2170</td>
<td>1185</td>
<td>985</td>
<td>0.0</td>
</tr>
<tr>
<td>[Villa Nova et al., 1976]</td>
<td>2005</td>
<td>1080</td>
<td>925</td>
<td>0.0</td>
</tr>
<tr>
<td>[Marques et al., 1980]</td>
<td>2083</td>
<td>1000</td>
<td>1083</td>
<td>0.0</td>
</tr>
<tr>
<td>[Jordan and Heuveldop, 1981]</td>
<td>3684</td>
<td>1985</td>
<td>1759</td>
<td>-60.0</td>
</tr>
<tr>
<td>[Leopoldo et al., 1982]</td>
<td>2076</td>
<td>1676</td>
<td>400</td>
<td>0.0</td>
</tr>
<tr>
<td>[Franken and Leopoldo, 1984]</td>
<td>2510</td>
<td>1641</td>
<td>869</td>
<td>0.0</td>
</tr>
<tr>
<td>[Shuttleworth, 1988]</td>
<td>2636</td>
<td>1329</td>
<td>1317</td>
<td>-10.0</td>
</tr>
<tr>
<td>[Vörösmarty et al., 1989]</td>
<td>2260</td>
<td>1250</td>
<td>1010</td>
<td>0.0</td>
</tr>
<tr>
<td>[Russell and Miller, 1990]</td>
<td>2010</td>
<td>1620</td>
<td>380</td>
<td>10.0</td>
</tr>
<tr>
<td>[Nizhizawa and Koike, 1992]</td>
<td>2300</td>
<td>1451</td>
<td>849</td>
<td>0.0</td>
</tr>
<tr>
<td>[Matsuyama, 1992]</td>
<td>2153</td>
<td>1139</td>
<td>849</td>
<td>165.0</td>
</tr>
<tr>
<td>[Vörösmarty et al., 1996]</td>
<td>2301</td>
<td>1221</td>
<td>1080</td>
<td>0.0</td>
</tr>
<tr>
<td>[Costa and Foley, 1998]</td>
<td>2160</td>
<td>1360</td>
<td>1106</td>
<td>-306.0</td>
</tr>
<tr>
<td>[Zeng, 1999]</td>
<td>2044</td>
<td>1679</td>
<td>1095</td>
<td>-730.0</td>
</tr>
</tbody>
</table>

Figure A1: Spatial distribution of mean annual values of C in the basin from Era-Interim
Figure A2: Long-term mean annual cycle of E, P and C from Era-Interim in the period 1979-2012

Figure A3: Log-log relationship between SWB closure (P-E-R) and drainage area for different sub-catchments, for each evapotranspiration E data set used: (a) ORCHIDEE, (b) MPI, (c) GLEAM, (d) MODIS, and (e) ERA-Interim.
Figure A4: Closure of the AWB in the Amazon basin sub-catchments using the Era-Interim source for P and for each of the E data sets used: a) ORCHIDEE, b) MPI, c) GLEAM, d) MODIS, and e) ERA. Green values are related to positive closure, and yellow to red values to negative closure.
Figure A5: Relationship between AWB closure (P-E-C) and sub-catchment area using the Era-Interim source for P and for each E dataset used: a) ORCHIDEE, b) MPI, c) GLEAM, d) MODIS, and e) ERA.
Figure A6: Log-log relationship between AWB closure, P-E-C and sub-catchment area for each E data set used: a) ORCHIDEE, b) MPI, c) GLEAM, d) MODIS, and e) ERA.

Figure A7: Comparison of the annual cycle of E derived from Era-Interim and the satellite-based data sources MODIS, GLEAM and MPI.
Appendix B: Chapter 2

Annual cycles and complementary CCF analyses

In Figure A1a is depicted the mean annual cycle of the two variables used to construct the $P-E$ index in the Amazonia, both variables are presented with the same units. When comparing the mean seasonal cycle of Precipitation and Evapotranspiration is clear that Precipitation is the most sensible variable for the $P-E$ index, the former reaching a maximum value less than 300 mm month$^{-1}$ and the later with a maximum of approximately 100 mm month$^{-1}$. Precipitation governs the dynamics of the AM hydro-climatology, while Evapotranspiration exhibits a stable behavior around the year. Precipitation reaches the minimum when Evapotranspiration is maximum in August at the final of the dry season. In figure B1b we present how the mean annual cycles of surface pressure gradient and the TNA SSTs are not in phase having a contrary sign behavior. At higher pressure gradients (higher TNA surface pressure) are found the lowest values of SST with a one month lag, and vice versa. From the annual cycle of both variables is evident that the Ocean reaches the maximum temperature between August and October, after the boreal summer season and related with the ITCZ migration reaching the austral summer season. Figure B1c presents the annual cycle of the pressure gradient between the two regions and the Precipitation over the basin, it is clear that both variables are in phase, reaching both minimum values in August at the end of dry season. The comparison of the cycles suggests that the decrease in surface pressure at the basin is related with the further reduction of precipitation in the region, thus influencing the gradient between the two regions.
Figure B1: **a.** Mean annual cycles of Precipitation and Evapotranspiration in the AM region. **b.** Mean annual cycles of surface pressure gradient (TNA-AM) and the SSTs over the TNA. **c.** Mean annual cycles of surface pressure gradient (TNA-AM) and the Precipitation over the AM.

Figure A2a presents the mean annual cycles of the absolute value of the zonal winds velocities over the TNA and the surface atmospheric pressure between the two studied regions. For the period August-September is evident the relationship between the minimum values of the two variables, in Figure A2b is presented the CCF of the above variables, and there is 0 lag positive correlation ($r=0.64$, $P>0.95$) suggesting that the winds and pressure gradients are evolving in phase. It is important to note the one month lagged influence of the TNA zonal winds over the pressure gradient, present in the August-September period that agrees with the one month lag influence of the P-E index over the TNA SST’s. Figure A2c the annual cycles of P-E index and TNA zonal winds which also present a clear relation between the two variables in the August-September period, in figure A2d we present the CCF analysis between the two variables and there is a positive correlation when P-E lead the zonal winds at one month lag ($r=0.23$, $P>0.95$).
Figure B2: **a.** Mean annual cycles of TNA zonal wind velocities and TNA-AM surface pressure gradient. **b.** Lagged cross-correlations between time series of monthly mean values of the TNA zonal wind and TNA-AM pressure gradient, blue dashed lines represents the 95% confidence bounds. **c.** Mean annual cycles of the TNA zonal wind and P-E index. **d.** Lagged cross-correlations between time series of monthly mean values of the TNA zonal wind and P-E index, blue dashed lines represents the 95% confidence bounds.

Figure A3a presents the mean annual cycles of the surface atmospheric pressure gradient TNA-AM and the TNA SSTs, as noted in figure S1b there is a clear precedence of the SST maximum values by the minimum values of the pressure gradient at the August-September period, in figure A3b we present the CC between these two variables, there is a strong negative correlation when the pressure gradient lead the SST at one month lag ($r=0.82$, $P>0.95$). These results suggest that when the pressure gradient reaches its minimum values in August, the ocean experiences a progressive warming that reaches its maximum in September.
Figure B3: a. Mean annual cycles of TNA SST’s and TNA-AM surface pressure gradient. b. Lagged cross-correlations between time series of monthly mean values of the TNA SST's and TNA-AM pressure gradient, blue dashed lines represents the 95% confidence bounds.

**Extreme convection events in the Amazon River basin**

In Fig. B4 to B7 we present the time evolution of the total wind anomalies and SSTs in the TNA from two months before and two months after four different extreme convection events in the Amazon River basin: i) September 1999 (precipitation 22% above average), ii) June 2005 (precipitation 20% below average), iii) January 2009 (precipitation 9% above average), and iv) March 2010 (precipitation 25% below average).

Figure B4 shows that during the two months previous to September 1999 total winds were directed towards east, and that SSTs exhibit positive anomalies reaching up to 1°C. During September of 1999 precipitation is 22% above average in the AM, the central regions of AM were experiencing the higher values of precipitation, and the total winds are shift from to a westerly direction from 0°N to 15°N. There is a reduction in SSTs that reaches a maximum value of 0.7°C on the region between 5°N to 20°N. During the two months after the extreme decrease in precipitation in October-November 1999, total winds do not exhibit clear direction towards the AM and TNA SSTs exhibits negative anomalies in the region between 10°N to 30°N. Figure B5 shows that from April to August 2005 total winds in the TNA have a dominant easterly direction, and the TNA SSTs exhibits positive anomalies that reach up to 1.3°C in the region from 0°N to 25°N during the period April-June. During June of 2005 precipitation is 20% below average in the AM and the total winds exhibit an increase of around 3 m s⁻¹ near the coasts of South America. Fig. B5 an increase in the region with positive SST anomalies to a region between 5°S to 30°N from July to August 2005. Although positive anomalies in SSTs after June are similar to the ones observed in April and May, the spatial extent of the positive anomalies is increased after the precipitation extreme event in the AM.
Figure B6 shows that during the two months previous to January 2009 total winds do not exhibit a clear-cut defined direction, and that SSTs exhibit positive anomalies with maximum values of 1.0°C from 0°N to 20°N. During January of 2009 precipitation is 9% above average in the AM, the northwest regions of AM were experiencing the higher values of precipitation, and the total winds were increased in the westerly direction from 0°N to 30°N, Fig. B6 also shows that total winds are increased from 0°N to 10°N up to 5 m s-1 and that they are directed towards the continent between 5°N and 20°N. There is an increase in SST’s that reaches a maximum value of 0.8°C on the region between 5°N to 15°N and -50°E to -30°E, and there is also a zone of reduced SSTs that reaches a minimum of 0.4°C between 15°N to 30°N and -35°E to -20°E. During the two months after the extreme increase in precipitation, in February 2009, total winds exhibit a clear direction towards the AM and TNA SSTs experience negative anomalies that reach values of -0.6°C from 0°N to 30°N and from -40°E to -20°E, and in March 2005 total winds do not exhibit a clear direction and TNA SSTs negative anomalies moves towards northern TNA and reach values of -0.8°C.

Figure B7 shows that during the two months previous to March 2010 total winds exhibit a well-defined easterly direction, and that TNA SSTs exhibit positive anomalies ranging from 0.7°C to 1.9°C. During March 2010 precipitation is 25% below average in the AM; the total winds maintain their westerly direction in the TNA. There is an increase in SST’s that reaches maximum values of 2.2°C. During the two months after the extreme decrease in precipitation, in April 2010, total winds exhibits a clear-cut increase in the easterly direction and reach values of 4 m s-1 and TNA SSTs maintain positive anomalies ranging from 1.3°C to 2.0°C, and in March 2005 total winds do not exhibit a clear direction and TNA SSTs negative anomalies are maintained. Two months after the precipitation decrease in the AM the SST gradient in the TNA is decreased, meaning that there is a more uniform warming of the ocean.
Figure B4: Time evolution of total wind anomalies and SST’s two months before and two months after the extreme month of precipitation in the Amazon in September 1999. First row extreme precipitation anomalies in September of 1999, middle row total wind anomalies from July 1999 to November 1999, bottom row SST’s anomalies from July 1999 to November 1999.
Figure B5: Time evolution of total wind anomalies and SST’s two months before and two months after the extreme month of precipitation in the Amazon in June 2005. First row extreme precipitation anomalies in June of 2005, middle row total wind anomalies from April 2005 to August 2005, bottom row SST’s anomalies from April 2005 to August 2005.
Figure B6: Time evolution of total wind anomalies and SST’s two months before and two months after the extreme month of precipitation in the Amazon in January 2009. First row extreme precipitation anomalies in January of 2009, middle row total wind anomalies from November 2008 to March 2009, bottom row SST’s anomalies from November 2008 to March 2009.
Figure B7: Time evolution of total wind anomalies and SST’s two months before and two months after the extreme month of precipitation in the Amazon in March 2010. First row extreme precipitation anomalies in March of 2010, middle row total wind anomalies from January 2010 to May 2010, bottom row SST’s anomalies from January 2010 to May 2010.
Appendix C: Chapter 3

From the mutual information analysis we selected the lag value that corresponds to the one where mutual information steep changes to a slower decrease in its steep as the optimum value for $\tau$ from Fig. C1 we observe that for the P-E index $\tau = 1$, also Fig. C1 shows that the value of $m$ that corresponds to the lowest percent of false neighbors in the reconstructed phase-space is 4. The same procedure was used for all the time series (not shown).

Figure C1: Mutual information with respect to time delay $\tau$ and the false nearest neighbors with respect to the dimension $m$ for the AM P-E index.

The lagged dependence analysis for a 1% of the recurrence rate maintains the same results previously described for the feedback mechanism between the AM and TNA regions, and in some of the lags increasing the confidence of the results, this increase in confidence for some of the lags may be a result of the finer threshold that cleans spurious proximities between the trajectories and that may reduce the effect of noise in the time series (Fig. C2 a, b and c).
Figure C2: Recurrence lagged dependence between the four variables according to the direction defined in the mechanism of feedback. The arrow between the names of the variables denotes which is the leading variable, the gray area represents the 90% confidence area, the blue dashed lines represent the 95% confidence intervals and the red line represents the calculated RMD between the variables. The recurrence threshold $\epsilon$ was based on a fixed 1% of the recurrence rate $RR = 1/N^2 \sum_{i,j} r_{ij}$ for all the time series.

We compute a complementary recurrence analysis of the two-way mechanism, and analyze the dependence between the variables in both trajectories. In Fig. C3 we illustrate how the two-way feedback mechanism operates among the variables involved in process (such plot is furthered explained in Fig. 3-6 of the manuscript). The proposed two-way mechanism is supported by the significant correlations between the variables for lags from -2 to 2 months. Such results confirm that the mechanism acts as a two-way process, such that the TNA STTs affects the AM hydrology and vice versa. The
Recurrence results support the well-known influence of the TNA SSTs on the surface winds $S \rightarrow W$ [Chung et al., 2002] is presented in Fig. C3.

Figure C3: Recurrence lagged dependence between $G \rightarrow P$, $W \rightarrow G$ and $S \rightarrow W$. The arrow denotes the direction of influence between variables, the gray area represents the 90% confidence area, the blue dashed lines denote the 95% confidence intervals, and the red line represents the calculated $RMD$ between variables. The recurrence threshold $\varepsilon$ was based on a fixed 5% of the recurrence rate $RR = 1/N^2 \sum_{i,j} R_{ij}$ for all the time series, as in Fig. 3-6.

In Fig. C4 we present the two-way relations between the variables $G$ and $S$, as well as $P$ and $W$. Once the surface winds are affected by the TNA SSTs, then the pressure gradient ($G$) between the TNA and AM is affected. Cooling or heating of the TNA SSTs are also related with changes in $G$. The relation $S \leftarrow G$ can be seen also as a two-way relationship where the SSTs drive the atmospheric pressure gradient for several months (0 to 1, and 4 to 8) while the pressure gradient drives the SSTs during a period of 2 months (0-2). According to the recurrence analysis $P$ does not seem drive $W$, and therefore any connection between those two variables over the AM and the TNA is to be mediated by other variables in a nonlinear way as evidenced in the recurrence analysis. The influence of $W$ over $P$ is significant during the entire year.
confirming the well-established fact that there is a direct influence of zonal winds in the transport of moisture from the ocean to the continent to influence convective process in the AM [Yoon and Zeng, 2010; Moraes-Arraut et al., 2011; Poveda et al., 2014]

Figure C4: Recurrence lagged dependences supporting the two way influences between $G < - > S$ (left panel) and $P < - > W$ (right panel). The arrow denotes the direction of influence between the variables, the gray area represents the 90% confidence area, the blue dashed lines denote the 95% confidence intervals, and the red line represents the calculated RMD between variables. The recurrence threshold $\epsilon$ was based on a fixed 5% of the recurrence rate $RR = 1/N^2 \sum_{i,j} R_{ij}$ for all the time series, as in Fig. 3-6.
References


Brando, P. M., S. J. Goetz, A. Baccini, D. C. Nepstad, P. S. a Beck, and M. C. Christman (2010), Seasonal and interannual variability of climate and


Davidson, E. a et al. (2012), The Amazon basin in transition., *Nature*, 481(7381), 321–8, doi:10.1038/nature10717.

References


Keller, M., M. Bustamante, J. Gash, and P. Silva Dias (2009), Amazonia and Global Change.


References


Salazar, J. F. (2004), Balances hidrológicos y estimación de caudales extremos en la amazonia, Universidad Nacional de Colombia.


