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Working Group3

BIOSPHERE-ATMOSPHERE INTERACTIONS

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THE PHYSICAL HYDROCLIMATE SYSTEM OF THE AMAZON

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CHAPTER 5: THE PHYSICAL HYDROCLIMATE SYSTEM OF THE AMAZON

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ACRONYMS AND ABBREVIATIONS

ANA: Agência Nacional das Águas

ARME: Amazon Region Micrometeorological Experiment

BH: Bolivian High

CAO: Corriente de los Andes Orientales

CAPE: Convective Available Potential Energy

CHIRPS: Climate Hazards Group InfraRed Precipitation with Station data

CL: Chaco Low

CP-EN: Central Pacific El Niño

EN: El Niño (warm phase of ENSO)

ENSO: El Nino-Southern Oscillation


INMET: Instituto Nacional de Meteorologia

ITCZ: Intertropical Convergence Zone

LFC: Level of Free Convection

LN: La Niña (cold phase of ENSO)

m.a.s.l.: meters above sea level

MCS: Mesoscale Convective Systems

OLR: Outgoing Longwave Radiation

PB: Planetary boundary layer

SACZ: South Atlantic Convergence Zone
SALLJ: South American Low-Level Jet

SAMS: South American Monsoon System

SENAMHI: Servicio Nacional de Meteorología e Hidrología del Perú

SST: Sea Surface Temperature

TNA: Tropical Northern Atlantic Index

TRMM: Tropical Rainfall Measurement Mission

UTC: Universal Time Coordinated

VIIRS: Visible Infrared Imaging Radiometer Suite. Sensor onboard the Suomi-NPP Satellite (Suomi-National Polar-orbiting Partnership)
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KEY MESSAGES

- Given its tropical location enclosed by the Andes, its huge spatial extent (~7 million km$^2$) and forest cover, the Amazon River basin is one of the most critical elements of the Earth’s climate system. It is the largest and most intense land-based convective center, exerting a strong influence on atmospheric dynamics and circulation patterns both within and outside the tropics. It produces rainfall that results in the largest river discharges on Earth at 209,000 m$^3$/s, corresponding to 16-20% of the total world river discharge.

- The Amazon Basin is mainly characterized by lowlands with a warm and rainy climate. The upper part of the basin includes the eastern slope of the Andes, characterized by a wide variety of mountain climates (cloud forest, Páramos, Yungas, Punas, etc.).

- The El Niño-Southern Oscillation (ENSO) is the main cause of interannual variability in rainfall. EN is typically (but not exclusively) accompanied by droughts in the Amazon region, with recent severe droughts producing low river water levels, a high risk of forest fires, and impacts on natural river ecosystems. In addition to ENSO, Atlantic and Pacific SST variability influence the climate of the Amazon at interannual and interdecadal time-scale, including extreme events.

ABSTRACT

The physical hydroclimate system of the Amazon operates on several spatial and temporal scales. Large-scale processes, including solar forcing, control the main seasonal patterns of atmospheric circulation, rainfall, river discharge, and flooding. Persistent patterns of sea surface temperature control the main modes of interannual and interdecadal variability. Mesoscale processes cause other localized circulations. The abundant rainfall in the Amazon basin is a consequence of intense radiative heating; low-level convergence of oceanic water vapor; and permanent injection of water vapor into the atmosphere by the rainforest itself, aided by the mechanical uplifting of air by the Andes. Land surface processes partition precipitation into evapotranspiration, surface runoff, and deep drainage to the groundwater. The Amazon river system drains the surface and groundwater components of this abundant rainfall, forming the world’s largest watershed and feeding the world’s largest river. The Amazon has a discharge five times larger than the Congo, the world’s second-largest river. This large flow is highly seasonal and the river floods a large floodplain, with beneficial ecological and biogeochemical implications. Extreme flood and drought events are also associated with intense interannual rainfall variability, which, in turn, influence forest fires and biogeochemical cycles.

Keywords: Amazon water balance, extreme events
GRAPHICAL ABSTRACT

Incoming water vapor from the Atlantic Ocean

Amazon discharge into the Atlantic Ocean: 209,000 m³ s⁻¹

Outgoing water vapor across the Andes

long-term precipitation
2190 mm yr⁻¹ ± 7%

long-term evapotranspiration
1220 mm yr⁻¹ ± 15%

Outgoing water vapor to Central Brazil and the La Plata basin
5.1. INTRODUCTION

The Amazon is one of the three permanent centers of convection in the intertropical zone (along with Central Africa and Southeast Asia) – i.e., one of the main centers of ascending air that transports energy from land to the atmosphere. It is also the strongest of these three land-based convective centers, exerting strong influences on atmospheric circulation both within and outside the tropics. As one of the main drivers of the Hadley-Walker circulations, the Amazon is a critical energy source to the atmosphere, removing latent heat from the surface by evapotranspiration and releasing it to the atmosphere through condensation and cloud formation. This convective center's strength is due mainly to some of its geographical characteristics, including its large size, position near the equator, and the Andes mountains located downwind in the basin. As explained throughout this chapter, the rainforest also contributes to strengthening this convective center. The low albedo of the rainforest increases the absorbed net radiation, and the constant flux of water vapor to the atmosphere from rainforest evapotranspiration boosts the mean convection fields. At the same time, it smooths the seasonal and interannual variability of convection and rainfall in the region.

The region’s abundant convection and rainfall, along with the basin's large size, produce the world’s largest river, flanked by a complex network of channels and floodplains that transport sediments, carbon, and other nutrients. Intense seasonality and interannual variability are also dominant factors for local riverine communities who may have their towns either flooded or completely isolated depending on the mood of this river system – dictated by the modes of interannual climate variability of rainfall (Marengo and Espinoza, 2016).

Table 1 presents a synthesis of several estimates of the Amazon River basin's long-term water balance. Long-term estimates of precipitation (P) show little variability across studies, with a median value of ~2190 mm/yr±7%. The long-term mean runoff (R) is estimated at 1100 mm/yr±15%, which yields a median runoff coefficient (C) of 0.51±0.08.

Estimates of evapotranspiration (ET) have much higher uncertainties by comparison, with median values of ~1250 mm/yr±50%. This imbalance is likely because most high estimates of ET (>1500 mm/yr) are derived from reanalysis data, which (by design) do not conserve mass over the long-term. If these high values are excluded, the median
value of ET is closer to 1220 mm/yr±15%, with a median evaporative fraction (EF) of 0.54±0.07. Since the identity C+EF=1 must be achieved over the long term, it is reasonable to estimate that precipitation is split evenly between evapotranspiration and runoff.

**Table 1.** Long-term water balance of the Amazon river basin according to several studies. Studies marked by an asterisk (*) include the Tocantins river basin. Precipitation (P), evapotranspiration (ET), runoff (R), and the imbalance (P – ET – R) are expressed in mm/yr. The runoff coefficient (C) and evaporative fraction (EF) are dimensionless variables.

<table>
<thead>
<tr>
<th>Studies</th>
<th>Period</th>
<th>P</th>
<th>R</th>
<th>ET</th>
<th>C (R/P)</th>
<th>EF (ET/P)</th>
<th>Imbalance (P-E-R)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Salazar (2004)</td>
<td>1961-1990</td>
<td>218</td>
<td>940</td>
<td>124</td>
<td>8.043</td>
<td>0.57</td>
<td>1</td>
</tr>
<tr>
<td>Marengo (2004)*</td>
<td>1970-1999</td>
<td>211</td>
<td>105</td>
<td>157</td>
<td>9.050</td>
<td>0.74</td>
<td>-512</td>
</tr>
<tr>
<td>Getirana et al. (2014)</td>
<td>1989-2008</td>
<td>220</td>
<td>118</td>
<td>103</td>
<td>3.054</td>
<td>0.47</td>
<td>-13</td>
</tr>
<tr>
<td>Carmona (2015)</td>
<td>1982-2008</td>
<td>226</td>
<td>116</td>
<td>118</td>
<td>9.051</td>
<td>0.52</td>
<td>-86</td>
</tr>
<tr>
<td>Builes-Jaramillo and Poveda (2018)</td>
<td>1984-2007</td>
<td>222</td>
<td>965</td>
<td>124</td>
<td>8.043</td>
<td>0.56</td>
<td>12</td>
</tr>
</tbody>
</table>

This chapter reviews the main features and the main large-scale and mesoscale mechanisms that cause the mean Amazon climate, its interannual and interdecadal
variability, and extreme drought and flood events (Sections 5.2 and 5.3). The effects of extreme events on vegetation dynamics are discussed in Section 5.3. Next, the chapter describes the partitioning of precipitation into evapotranspiration (Section 5.4), runoff, flow seasonality, and floodplain dynamics (Section 5.5). Finally, the floodplain’s role in the biogeochemical cycles is discussed in Section 5.6.

This chapter’s description of the Amazon’s physical hydroclimate system also serves as an introduction to the biosphere-atmosphere interactions discussed in Chapters 6 and 7, and to the Amazon climate change discussed in Chapter 20. Chapter 6 discusses the influence of the physical hydroclimate system on biogeochemical cycles, whereas Chapter 7 presents the rainforest's role in the water and energy exchange of this coupled biosphere-atmosphere system. Chapter 20 presents the long-term variability and changes in temperature and hydrometeorology in the Amazon.

5.2. MAIN FEATURES OF THE AMAZON CLIMATE

5.2.1. Spatial distribution of climate variables

5.2.1.1. Air temperature

Due to high, relatively constant incoming solar radiation, air temperature in the Amazon is practically isothermal, with only a small variation throughout the year except in the southern part (Rondônia, Mato Grosso, Bolivian Amazon, and Southern Peruvian Amazon). Annual averages show very high temperatures in the central equatorial region, with means that exceed 27-29ºC. The seasonal thermal amplitude is 1-2ºC, and average values range from 24ºC to 26ºC. The city of Belém (PA) has a maximum monthly average temperature of 26.5ºC in November, and a minimum of 25.4ºC in March, while Manaus (AM) has its temperature extremes in September (27.9ºC) and April (25.8ºC). In austral winter, the cold air masses that produce frosts in the South and Southeast of Brazil can also cool the southern and western Amazon, with significant air temperature drops (Ricarte and Herdies 2014, Viana and Herdies 2018). Near the Andes, the maximum monthly mean temperature in Santa Cruz de la Sierra, Bolivia, reaches 26.1ºC in September and 20ºC in June. Despite small seasonal fluctuations,
large temperature oscillations (high amplitude) are typical of the diurnal cycle in this region.

5.2.1.2. Atmospheric circulation

The mean atmospheric circulation in the Amazon is forced by the annual cycle of solar radiation. The atmospheric circulation's main features are described here, while the solar forcing is described in Section 5.3.2.

Near the Amazon delta, maximum rainfall is observed during austral summer-fall, and dry conditions prevail during wintertime (Figure 1). This is due to the alternating warming of the two hemispheres and to the annual cycle associated with the seasonal meridional migration of the Intertropical Convergence Zone (ITCZ) (Vera et al. 2006a). The trade winds coming from the tropical North and South Atlantic converge along the ITCZ and are associated with subtropical anticyclones in the North and South Atlantic.

**Figure 1.** Schematic of the main climatological features in South America. The blue and red lines represent June-July-August (JJA) and December-January-February (DJF), respectively. The annual cycle of rainfall (bars) is shown for stations located in various
sections of the Amazon region (in mm), indicated by dots. Low-level circulation features: CL, Chaco Low; BH, Bolivian High; ITCZ, Intertropical Convergence Zone; MCS, mesoscale convective system; SACZ, South Atlantic Convergence Zone; SALLJ, South American low-level jet. Sources of rainfall data: INMET and ANA (Brazil), SENAMHI (Peru), SENAMHI (Bolivia) and INAMHI (Ecuador). The figure is adapted from Figure 1 of Cai et al. (2020). Climatology of the period 1961-2010.

Monsoonal rain over the Amazon basin during austral summer provides moisture to establish an active South Atlantic Convergence Zone (SACZ; Figure 1). The SACZ is characterized by a convective band that extends northwest-southeast from the Amazon Basin to the subtropical South Atlantic Ocean. It is identifiable by persistent cloudiness and frequently configured in the austral summertime (Ambrizzi and Ferraz 2015). The SACZ’s northern edge merges with the Atlantic ITCZ (Cai et al. 2020). Diabatic heating in the Amazon basin contributes to the formation of the Bolivian High (BH) in the upper atmosphere (Lenters and Cook 1997). At the regional scale, moisture transport in and out of the Amazon basin is critical for the rainfall regime, particularly during the wet season. The moisture from the Amazon is exported out of the region, transported via the South American Low-Level Jet (SALLJ) east of the Andes, interacting with the Chaco Low (CL) and contributing to precipitation over the La Plata basin by intensifying mesoscale convective systems (Marengo et al., 2004, Drumond et al. 2008, 2014; Arraut et al. 2012; Vera et al. 2006b, Liebmann and Mechoso 2011, Jones and Carvalho 2018, Gimeno et al. 2016, 2019, Jones 2019, Cai et al. 2020).

5.2.1.3. Rainfall

Because it extends into both hemispheres, the Amazon is characterized by several rainfall regimes due to the alternating warming of each hemisphere. During a ‘normal’ year, rainfall in the region shows opposing phases between the northern and southern tropics, with a rainy season in austral winter in the north and austral summer in the south. In the southern Amazon, rainfall peaks during austral summer; in the central Amazon and near the Amazon delta, it peaks in austral autumn; and north of the Equator, it peaks in austral winter (Figure 1). The northwest equatorial region experiences low rainfall seasonality, with wet conditions throughout the year. For more details about rainfall regimes in the Amazon basin, see Figueroa and Nobre (1990), Rao

The onset and demise of the rainy season in the Amazon varies gradually from south to north. The end of the rainy season is more regular than its onset. The rainy season in the southern Amazon ends in April, while in the northern part, it ends in September. SST anomalies in the Pacific or Tropical Atlantic play a dynamic role in controlling the beginning and end of the rainy season (Liebmann and Marengo 2001, Liebmann et al. 2007; Arias et al. 2015).

5.2.2. The role of ENSO and other large-scale mechanisms

5.2.1.1. ENSO

The El Niño-Southern Oscillation (ENSO) is the main cause of global interannual variability in the water and energy budgets. ENSO extremes represent a reversal of the typical SST patterns in the Tropical Pacific – El Niño (EN)/La Niña (LN), when there is warming/cooling in the eastern or central-eastern tropical Pacific. EN is typically (but not exclusively) accompanied by drought in the Amazon region. In general, recent severe droughts over the Amazon have resulted in low river water levels, a high risk of forest fires, and impacts on natural river ecosystems (Cai et al. 2020).

Changes to atmospheric circulation during EN and drought have been summarized by Builes-Jaramillo et al. (2018a) and Jimenez-Munoz et al. (2019): observed anomalies in the vertical distribution of zonal and meridional wind are consistent with SST anomalies. During drought and EN years, subsidence anomalies appear over areas with negative rainfall anomalies over the Amazon, with convection and intense rainfall over warm SST in the eastern Equatorial Pacific region. The upper-level convergence anomalies observed during drought years over tropical equatorial South America (east of the Andes) are consistent with low-level subsidence anomalies. This suggests anomalies in the upper and lower branches of the Hadley circulation over tropical South America east of the Andes, and of the Walker circulation over the equatorial Atlantic. The ascending branch of the Walker circulation over the eastern central Pacific is the main driver of the subsidence branch over the Amazon Basin east of the Andes, which extends all the way to the tropical Atlantic.
There are different “types” of EN depending on the location of maximum warm anomalies over the tropical Pacific: Eastern Pacific (EP) EN or Central Pacific (CP) EN (Takahashi et al. 2011). Because the Hadley and Walker circulations are affected differently during EP-EN and CP-EN episodes (Zheleznova and Gushchina 2017), they lead to different precipitation anomalies over South America (Tedeschi and Collins 2017; Sulca et al. 2018). Physical mechanisms behind the different patterns of rainfall deficits during CP- and EP-ENs and warm Tropical Northern Atlantic Index (TNA) events are described in Jimenez-Munoz et al. (2019). EP-EN years were detected in 1983 and 1998, whereas CP-EN occurred in 2010 and 2016 (Sulca et al., 2018; Gu and Adler 2019, Gloor et al., 2013, 2018).

5.2.2.2. PDO, AMO, MJO

In addition to ENSO, there are two other modes of interannual and interdecadal variability with teleconnections that influence the climate of the Amazon: The Pacific Decadal Oscillation (PDO) and the Atlantic Multidecadal Oscillation (AMO). They represent changes in the organization of air-sea interactions that vary at decadal scales and affect the sea surface, inducing later circulation and rainfall changes in the Amazon. For a detailed definition of these modes of variability, please see the Glossary.

Consistent with the ENSO (EN) positive phase, the PDO and AMO’s positive phases matched the intensification of negative rainfall anomalies in the Amazon towards the end of 2015, during the 2015-16 EN event (Aragão et al. 2018). This finding is consistent with previous work (Kayano and Capistrano, 2014) showing that the Atlantic Multidecadal Oscillation (AMO) and the ENSO influence South American rainfall at the end of the year, before the peak of EN.

Positive phases of the PDO are associated with an increase in precipitation in the central and northern parts of the basin and a decrease in the southern regions (Gloor et al., 2013). Andreoli and Kayano (2005) show that EN effects on rainfall over South America differ from those of the PDO phases in the Amazon. For example, they show negative precipitation anomalies for the warm PDO regime, consistent with the descending motion and cyclonic circulation over northern South America and the adjacent Atlantic sector. On the other hand, the relatively weaker circulation patterns in these sectors result in smaller magnitude precipitation anomalies in the Amazon for the cold PDO phase.
The intraseasonal variability is particularly important during the austral winter (Mayta et al., 2018). Previously, Souza and Ambrizzi (2006) found that the Madden–Julian Oscillation (MJO) is the main atmospheric mechanism influencing rainfall variability at intraseasonal timescales over the eastern Amazon and during the rainy season in northeast Brazil. During the drought of 2005, however, the intraseasonal oscillation was weaker than normal, favoring drought conditions in the region. The Tropical North Atlantic played a major role in this drought (Builes-Jaramillo et al., 2018b).

5.2.3. Extreme drought and flood events

In the last 15 years, the Amazon Basin has witnessed climate extremes, some of them characterized as ‘events of the century’: droughts in 2005, 2010, and 2015–16 and floods in 2009, 2012, 2014, and 2021. Historical records show previous droughts in 1926, 1964, 1980, 1983, and 1998 and floods in 1953, 1988, 1989, and 1999. These events have been linked to modes of natural climate variability (EN, warm TNA anomalies) with strong impacts on natural and human systems. Some of the Amazon's main cities were flooded during flood years or isolated by extremely low river levels during droughts. The number of fires increased during drought years, releasing carbon and smoke and soot into the atmosphere and affecting the local population (Marengo and Espinoza 2016, Gatti et al. 2014, Aragão et al. 2018, Jimenez-Munoz et al. 2016, 2019). The year 1999 and other wet years (1988-89, 2007-2008, 2011-2012) were LN years (see Chapter 22). It is worth mentioning that droughts and floods are not synchronous and do not affect the whole basin in the same way, as seen in Figures 2 and 5.3.

Overall, droughts affect the north-central Amazon, but the spatial pattern differs from one EN event to another and even from one drought case to another (Figure 2). Droughts in the Amazon have been related to EN events, such as in 1912, 1926, 1983, 1997–1998, and 2015-16 (e.g., Aceituno 1988; Williams et al. 2005, Coelho et al. 2013, Marengo et al. 2018, Jimenez-Munoz et al. 2018, 2019). However, the 1964 and 2005 severe droughts were exceptions, indicating TNA's active influence on those extremes (Marengo et al. 2008, Zeng et al., 2008, Builes-Jaramillo et al., 2018b). The 2010 extreme drought was related to the successive occurrences of an El Niño in austral summer and a very warm TNA in the boreal spring and summer (Espinoza et al. 2011; Marengo et al. 2011, Lewis et al. 2011, Gatti et al. 2014, Andreoli et al. 2012). Figures
2 and 3 show seasonal rainfall anomalies in South America for drought and wet years, respectively. In each case, whether EN or not, the geographical distribution of droughts may differ, affecting the southeastern, central, or northern Amazon differentially, and thus impacting the region's hydrology.

**Figure 2.** Spatial patterns of precipitation anomalies during seasons DJF, MAM, JJA, and SON for drought years in the Amazon. These are for different strong EN and TNA warming. Precipitation anomalies were obtained from the CHIRPSv2.0 dataset using the reference period 1981-2010. A black contour marks the Amazon Basin. Adapted from Jimenez-Muñoz et al (2019; ©RMetS).
5.2.4. Andean-Amazon hydrometeorology and variability

This section focuses on the western Amazon, including the Andean part of the Amazon basin. The region encompasses the upper Madeira basin in Bolivia, Peru and Brazil; the Amazonas-Solimões basin in Peru and Ecuador; and the Japurá-Caquetá basin in Colombia and Brazil. This region presents a wide variety of mountain climates, including humid conditions in the cloud forests, Páramos, and Yungas, and dry conditions in the highland Punas.
5.2.4.1. Seasonal patterns

Seasonal rainfall cycles in the upper part of the Andean-Amazon basins of Colombia and Ecuador follow a unimodal regime with a wet season during the boreal summer (Laraque et al. 2007; Arias et al. 2020). In these basins, river discharge peaks around May-July (e.g., Napo and Caquetá rivers in Figure 7) – a pattern associated with the intensification of westward moisture advection from the equatorial Amazon basin and orographic uplift forced by the Andean topography during boreal summer (Rollenbeck and Bendix 2011; Campozano et al. 2016). The Andean-Amazon basins of Ecuador exhibit a bimodal annual cycle of precipitation, with peak discharge observed around March-April and October-November in the upper part of the Napo, Pastaza, and Santiago basins (Campozano et al. 2018) (e.g., Reventador station in Figure 1). Consequently, the lowlands of these intra-Andean basins follow a bimodal annual cycle of discharge with peaks around June-July and October-November (Laraque et al. 2007). In these regions, less rainfall during boreal summer is associated with atmospheric subsidence that inhibits convective activity (Campozano et al., 2016; Segura et al., 2019).

In the southern tropical Andean-Amazon basins (mainly south of 8˚S), the dry season occurs in June-August and the rainy season in December-March, linked to the mature phase of the South American Monsoon System (SAMS) and the meridional movement of the ITCZ. River discharges over these basins show unimodal cycles peaking around January and March (e.g., Beni, Ucayali and Huallaga rivers in Figure 7; and Santa Cruz and San Gabán stations in Figure 1) (Espinoza et al. 2011; Lavado-Casimiro et al. 2012; Molina-Carpio et al. 2017). Rainfall seasonality is particularly strong in the upper and drier part of the Andean-Amazon basins (usually higher than 3000 m.a.s.l.), where around 75% of total annual rainfall is observed between November and March (~100 mm/month), driven by upward moisture transport from the Amazon toward the mountains (Garreaud et al. 2009). Easterly winds in the upper troposphere (200-300 hPa) also favor moisture fluxes from the Amazon to the Andes at different time scales (Garreaud et al., 2009; Segura et al., 2020).

Most of the Amazon’s Andean tributaries drain to two main rivers: the upper Madeira river (mainly from the Bolivian and southern Peruvian Amazon) and the Amazonas-Solimões river (mostly from the Peruvian and Ecuadorian Amazon) (Figure 7). At the
Porto Velho station, the basin of the upper Madeira river basin spans 975,500 km², of which 23% are in the Andes. Mean annual discharge at Porto Velho is estimated at 18,300 m³/s, with peak values around 36,000 m³/s from March-April and the lowest discharge around ~5000 m³/s from September-October (Molina-Carpio et al. 2017) (Figure 7). At the Tabatinga station, the Amazonas-Solimões river basin spans 890,300 km², of which ~ 40% are in the Andes. The mean annual discharge at Tabatinga is estimated at 38,000 m³/s, with peak values around 51,000 m³/s from April-May and the lowest discharge around ~20,000 m³/s in September (Lavado-Casimiro et al. 2012) (Figure 7).

5.2.4.2. Interannual variability and extremes

In the Andean-Amazon region, a rainfall deficit (excess) during austral summer is frequently associated with El Niño (La Niña) events (Poveda et al. 2006; Espinoza et al. 2011). However, different patterns occur in the upper and lower parts of Andean-Amazon basins (Arango-Rueda and Poveda 2019). Recent studies have also reported different precipitation anomalies for the Central-Pacific and Eastern-Pacific El Niño types (Lavado-Casimiro and Espinoza 2014; Sulca et al. 2018; Navarro-Monterroza 2019). In general, the Central-Pacific El Niño (La Niña) is associated with rainfall deficits (excesses) in the upper part of the basin (Andean regions of Colombia, Ecuador, and Peru). These anomalies are weaker during Eastern-Pacific El Niño (La Niña) events. In contrast, in the upper Madeira basin rainfall anomalies are stronger during the Eastern-Pacific El Niño.

Rainfall anomalies over the Andean Amazon Basin range from ±0.5 to ±2.0 mm/day (Sulca et al., 2018; Jiménez-Muñoz et al., [2019] 2021). During the austral autumn, winter, and spring, rainfall anomalies over the Andean-Amazon region are mainly related to SST variability in the TNA, which is the main source of atmospheric moisture for the Andean-Amazon region (Arias et al. 2015; Hoyos et al. 2017; Poveda et al. 2020). Warm TNA anomalies are associated with increased precipitation in Colombia and Venezuela, related to enhanced atmospheric water vapor transport from the tropical Atlantic and the Caribbean Sea toward northern South America (e.g., Arias et al., 2020). In the Andean-Amazon regions of Ecuador, Peru, and Bolivia, warm conditions in the TNA are related to rainfall deficits, associated with a reduction in moisture advection from the Atlantic Ocean and enhanced atmospheric subsidence over the central and

As a result of rainfall anomalies, extreme hydrological events in the Andean-Amazon basins have been associated either with El Niño/La Niña events or with SST anomalies in the TNA. The very unusual wet austral summer period of 2014, originating on the eastern slopes of the Peruvian and Bolivian Andes, was associated with warm anomalies in the western Pacific-Indian Ocean and over the subtropical South Atlantic Ocean (Espinoza et al. 2014). Wet conditions in the Bolivian Amazon during the 2014 austral summer were superimposed on flood waves from the main sub-basins, producing major floods in this region that same year (Ovando et al., 2016). This was also related to long-term atmospheric blocking systems during January and February of 2014 over southeastern Brazil, which ultimately caused the drought over São Paulo during the austral summer of 2014. In the higher part of the Amazon Basins' inter-Andean rivers, floods are frequently triggered by intense storms and/or rapid glacier melting during the austral spring-summer (Huggel et al. 2015).

5.3. THE AMAZON CONVECTION AND MESOSCALE CIRCULATIONS

5.3.1. Nature of the Amazon convection

Atmospheric deep convection is typical in the tropics in association with the ascending branch of the Hadley-Walker cells. Upward motion extends from near the surface to above the 500 hPa level, reaching the level of free convection (LFC) where buoyant convection begins. At the large-scale (> 1000 km), seasonal changes in the thermal contrast between tropical South America and the Atlantic Ocean modulate wind circulation, which supplies the available energy and moist instability over the Amazon Basin (Vera et al. 2006a). These features provide the convective available potential energy (CAPE), the gross moist instability, and the rising motion – the essential mechanisms to produce deep atmospheric convection (Garstang et al. 1994; Cohen et al. 1995; Zhou and Lau 1998). At regional (100-1000 km) to local scales (< 100 km), Amazon convection is also related to the land surface wet-bulb temperature, generally above 22°C (Eltahir and Pal 1996), which is closely determined by surface humidity and sensible and latent fluxes from the local land surface (Fu et al. 1999).
Deep atmospheric convection contributes to about 80% of the total annual precipitation in the Amazon Basin, while only 20% of yearly rainfall is associated with local systems (Greco et al. 1990). Seasonal changes in Amazon convection are related to changes in the moistening of the planetary boundary layer (PBL) and changes in the temperature at the top of the PBL (Fu et al. 1999; Liebmann and Marengo 2001). However, in the northwestern Amazon, deep convection is particularly intense year-round because the warmer land surface provides highly unstable atmospheric profiles. In addition, the concave shape of the Andes induces a low-level convergence over the northwestern Amazon Basin, which is related to high annual rainfall (>3000 mm) in this region (Figueroa and Nobre 1990; Espinoza et al. 2009b). Because deep convection over the Amazon is related to a strong release of latent heat, the Amazon Basin is an important source of energy. Through the equatorial Kelvin and Rossby waves and their interactions with the orography, the Amazon modulates the main regional structures of the atmospheric circulation in South America (Silva Dias et al. 1983; Figueroa et al. 1995; Junquas et al. 2015).

5.3.2. Solar forcing

Following the seasonal migration of the solar radiation maximum, the major heating zone migrates from northernmost South America (including the northern Amazon Basin) in austral winter to the central and southern Amazon in austral summer (Horel et al. 1989). Consequently, convective activity and rainfall enhancement show a seasonal displacement following the heating zone migration (see Section 5.2.1). Figure 4 shows the spatial and temporal evolution of the outgoing longwave radiation (OLR) in tropical South America, closely related to solar forcing and the development of deep convection.

The alternating warming of the two hemispheres modulates the seasonal displacement of the ITCZ, including its Amazonian part (Figure 1) and the ascendant branch of the Hadley-Walker cells, which is associated with maximum rainfall over the equatorial Amazon Basin. Over this region, solar radiation peaks at the equinoxes (Figure 4), and the northeastern Amazon Basin displays the maximum precipitation in the austral autumn, with peaks in April and May. However, in some western equatorial Amazon regions, the wet season occurs during austral fall and spring (see Section 5.2.1). In austral spring, surface heating by solar radiation is highest over the central and southern Amazon (south of 5°S), where deep convection appears. By late November, deep
convection happens over most of the Amazon Basin, mainly from 5°S to 20°S, but it is still absent over the eastern Amazon Basin and northeast Brazil (Horel et al. 1989; Zhou and Lau 1998).

**Figure 4.** (a) 1974–2019 mean annual values of outgoing longwave radiation (OLR, in W.m⁻²) over tropical South America. (b) Time-latitude diagram of the climatology of monthly OLR (1974–2019) averaged across a 10’ longitudinal strip centered on the black line over tropical South America shown in panel a. Adapted from Horel et al. (1989). Interpolated OLR data provided by the NOAA/OAR/ESRL PSL (https://psl.noaa.gov/; Liebman and Smith 1996).

At the peak of austral summer, following the southward migration of the sun, heating and convective activity moves toward the subtropical highlands. Rainfall peaks over the central Andes and the southern Amazon Basin during this season. The thermal contrast between the continent determines the SAMS configuration (Marengo et al., 2012). The mature phase of the SAMS (typically from late November to late February) exhibits four dominant features (Section 5.2.1 and **Figure 1**): (i) an anticyclone located over Bolivia at 200–300 hPa (the Bolivian High -BH); (ii) the occurrence of high surface temperatures over the Atlantic Ocean before the wet season begins in the southern Amazon; (iii) a northwest-southeast oriented band of maximum cloudiness over the
southeast of the continent, the SACZ; and (iv) the intensification of the SALLJ to the east of the Andes (See review in Espinoza et al. 2020).

5.3.3. Forest breeze and river breeze circulations

Forest and river breezes are mesoscale (10-100 km) circulations close to large rivers. They result from differences in the sensible and latent heat fluxes between the hot land and the cool water during daytime, which produces a horizontal pressure contrast. This mechanism enhances cloudiness over land during the day, while clear skies predominate over water. The opposite occurs during the night. In the Amazon Basin, convergence zones lead to enhanced rainfall over forests away from large rivers, and convective activity is reduced near rivers (e.g., Paiva et al. 2011; Figure 5).

Figure 5. Rainfall estimated by TRMM 3B42 between (a) 15 to 06 UTC; and (b) between 06 and 15 UTC. Adapted from Paiva et al. (2011). (c) Image of the VIIRS sensor (Visible/Infrared Imager Radiometer Suite) in true color corresponding to July
Several studies have described river breezes in the central Amazon, using both observed and modeling approaches (e.g., Ribeiro and Adis 1984; Garstang and Fitzjarrald 1999; Cutrim et al. 2000). Near the Amazon-Tapajós confluence (Figure 5), rain gauges close to large rivers show less convective rainfall in the afternoon. Still, this deficit is more than compensated by additional nocturnal rainfall (Fitzjarrald et al. 2008). Near Manaus, dos Santos et al. (2014) show that river breezes and their impact on moisture transport are more evident during the dry season. The authors show that winds away from the rivers are frequent in the morning and afternoon, transporting moist air from the rivers to the city of Manaus. In contrast, winds blowing towards rivers are mainly observed at night.

River breezes affect moisture transport (Silva Dias et al. 2004) and local rainfall patterns. Paiva et al. (2011) showed a marked reduction in rainfall over the Solimões-Amazon river and along most Amazon tributaries. Since meteorological stations are often sited near large rivers (where most Amazon cities are situated), rain gauge-derived estimates of Amazon rainfall may be biased by river breezes (Silva Dias et al. 2004; Paiva et al. 2011).

5.3.4. Sea breeze and coastal circulations

The sea breeze system occurs at coastal locations due to the propagation of cool marine air towards inland areas. This system is initiated when the land surface heats faster than the sea surface (generally under relatively clear sky conditions). The thermal contrast creates a pressure gradient force directed from sea to land, causing a shallow layer of marine air to move inland (Miller et al. 2003).

Over the easternmost Amazon Basin, the presence of numerous bays, rivers, lakes, and the Atlantic Ocean, create the ideal environment for the formation of local circulations, which modulate the regional weather and climate (Souza Filho 2005, Planchon et al. 2006; Germano and Oyama 2020). The main circulation patterns of the coastal and bay breezes over this region have been described elsewhere, based on observational and modeling studies (e.g., Silva-Dias et al. 2004; Germano et al. 2017; Wanzeler 2018). In
Belém (eastern Amazon Basin), the bay breeze starts in the morning and early afternoon. It is characterized by significant changes in wind direction from south to north (Matos and Cohen 2016) and is associated with the presence of stationary cloudiness. Rainfall peaks during the April-May season coincide with the sea breeze's maximum activity, which interacts with the Atlantic Ocean's trade winds to produce storm systems known as squall lines (Kousky 1980; Silva Dias 1987; Cohen et al. 1995).

Squall lines are multicellular storms that propagate inland in the Amazon Basin for over 1000 km at speeds of 50–60 km h\(^{-1}\) (Garstang et al. 1994; Greco et al. 1994). At the mesoscale, squall lines are characterized by advection of moisture produced by a sea breeze, a strong and deep low-level easterly jet, and a heat source in the central and western Amazon (Cohen et al. 1995). Strong jets tend to propagate the squall lines at higher speeds, with a longer lifetime and increased cloud development, forming thunderstorms with strong updrafts and downdrafts, as well as lightning. Downdrafts and lightning, in turn, cause disturbances that affect ecosystem dynamics, as described in Section 5.3.6.

5.3.5. Orographic induced circulations and spatial rainfall distribution in the Andean-Amazon region

The Andean-Amazon hydrometeorology is characterized by interactions between regional atmospheric circulation, lowland-highland temperature contrast, and the complex Andean topography (e.g., Houze 2012; Roe 2005; Barry 2008). In addition, regional atmospheric circulation over South America is directly influenced by the Andean orography, particularly at low-levels (Figueroa et al. 1995). In the Andean-Amazon region, the SALLJ and the Llanos Jet (or Corriente de los Andes Orientales – CAO) are strongly controlled by the presence of the Andes, which acts like a barrier to the west, and the Amazon Basin to the east (e.g., Marengo et al. 2004; Jiménez-Sánchez et al. 2019). These LLJs are key elements of the South American atmospheric circulation because they transport vast quantities of moisture along with large meridional distances throughout the east of the Andes. Indeed, the CAO's easterly flow reaches the eastern piedmont of the Andes as the northernmost leg of the SALLJ (Espinoza et al., 2020; Poveda et al., 2020).
At the local scale, Andean orography can influence the atmospheric circulation through mechanical and thermal processes. The diurnal cycle of insolation generates thermally driven winds, such as anabatic (warm upslope) and katabatic (cold downslope) winds due to radiative warming of the surface during the day and radiative cooling during the late afternoon and night, respectively (e.g., Wallace and Hobbs 2006; Junquas et al. 2018). In addition, katabatic winds from the Andean highlands could trigger mesoscale convective systems (MCS) over the Andean-Amazon transition region (Trachte et al. 2010a,b; Kumar et al. 2020). Over this region, large and medium MCS are generally related to wet episodes, enhanced by the orographic lifting of moisture advection from the SALLJ (e.g., Giovannettone and Barros 2009; Romatschke and Houze 2013). Consequently, the mountainous precipitation diurnal cycle is associated with complex characteristics related to local atmospheric circulations (Poveda et al., 2005; Junquas et al., 2018). For example, on the eastern slopes of the tropical Andes, the highest precipitation rates are observed at nighttime due to downslope wind and moisture transport (Figures 5a and b). Observational and modeling studies have shown that inter-Andean valleys also generate mechanical channelization of the moisture flux, which could contribute to moisture and rainfall over the tropical Andes, where glaciers, agriculture and food security depend on precipitation. This includes regions such as La Paz, Cuzco, and the Mantaro valleys, among many others (Egger et al. 2005; Junquas et al. 2018; Saavedra et al. 2020). Convective activity forced by the Andes also generates sudden reversals of the river stage in the western Amazon (e.g., near Iquitos, Peru), where riparian agriculture is closely related to the annual hydrological cycle (Figueroa et al., 2020).

Interactions between large-scale atmospheric circulation and the orographic circulations described above contribute to the high spatial variability of precipitation over the Andes-Amazon region. Studies have described a complex relationship between altitude and rainfall, which produces a strong spatial rainfall gradient associated with the windward or leeward exposure of the rain station to the dominant moist wind (Bookhagen and Strecker 2008; Espinoza et al. 2009b, Rollenbeck and Bendix 2011). The highest rainfall rates in the Amazon Basin (6000–7000 mm/year) are generally observed at about 400–2000 m.a.s.l in the Amazon Basin of Colombia, Ecuador, Peru, and Bolivia (Poveda et al. 2014; Espinoza et al. 2015; Chavez and Takahashi 2017) (e.g., San Gabán station in Figure 1). As a result of these rainfall characteristics, the
Andean basins show the highest runoff per unit area of the Amazon River basin (Moquet et al. 2011; Builes-Jaramillo and Poveda 2018), and Andean rivers drain sediments, pollutants, and nutrients downstream to the Amazon lowlands (McClain and Naiman 2008; Vauchel et al. 2017). In turn, the Amazon lowlands export water vapor and nutrients to the Andes through the moisture-laden trade winds, which is part of a strong interaction between the Amazon-Andes hydroclimatic system (e.g. Staal et al., 2018; Weng et al., 2018, Espinoza et al., 2020).

5.3.6 The role of extreme weather events on ecosystem dynamics

At least two types of extreme weather events affect ecosystem dynamics and the natural carbon cycle. First, severe storms associated with squall lines can propagate strong downdrafts (Fujita 1990, 1981, Garstang et al. 1998) that cause forest blowdowns (Nelson 1994, Garstang et al. 1998, Negrón-Juárez et al. 2010, Espírito-Santo et al. 2010), affecting forest structure and species composition (Marra et al. 2014, Rifai et al. 2016, Magnabosco Marra et al. 2018, Chambers et al. 2009). Second, lightning is a frequent disturbance mechanism that can propagate fire and kill trees directly (Gora et al. 2020, Yanoviak et al. 2020, McDowell et al. 2018, Foster, Knight, and Franklin 1998). The frequency of lightning is positively associated with the density of large trees and biomass stocks in tropical forests (Gora et al. 2020). In the Amazon, this is important in the southern and eastern transition zones between forests and savannas, but also in Roraima state (Gora et al. 2020).

3.6.1. Severe storms, blowdowns, and impacts on forest ecosystem dynamics

Wind is a major cause of disturbance in forests worldwide, with impacts ranging from minor loss of leaves to widespread tree-mortality (Mitchell 2013). In the Amazon, convective storms can generate strong downdraft winds and extreme rainfall (e.g., 26-41 m s\(^{-1}\) and 30 mm h\(^{-1}\), respectively) (Garstang et al., 1998; Fujita et al., 1990; Negrón-Juárez et al., 2010) that can fell forest patches ranging in size from <2 ha (Negrón-Juárez et al., 2011) to >3,000 ha (Nelson et al., 1994). Large blowdowns can be associated with squall lines (Negrón-Juárez et al., 2010; Araujo et al., 2017). Forest blowdowns can be detected with remote sensing imagery because they create a large contrast in geometric and reflectance patterns between images acquired before and after the event (Figure 6a).
Blowdowns occur across the Amazon Basin, with the highest frequency in the Northwest region (Nelson et al., 1994; Negrón-Juárez et al., 2018; Espírito-Santo et al., 2010). In the Central Amazon, blowdowns mostly occur during the transition from dry to rainy seasons (Negrón-Juárez et al., 2017). The size distribution of blowdowns follows a power-law (Negrón-Juárez et al., 2018; Chambers et al. 2009), resulting in a mosaic of forest patches at different successional stages (Chambers et al. 2013). Because of their greater frequency, relatively small-sized patches dominate the landscape.

Tree damage and mortality occur when wind and rain loads exceed the mechanical stability of trees, leading to snapping and uprooting (Ribeiro et al., 2016; Peterson et al., 2019). In the Amazon, winds, and rain interact with different forest types that may harbor more than 280 tree species in a single hectare (de Oliveira et al., 1999). In these heterogeneous forests, storm mortality can be controlled by biotic and abiotic factors (e.g., within species and across topography), with severely damaged areas experiencing up to 90% of tree mortality (Magnabosco et al., 2014; Rifai et al., 2016) (Figure 6b). The forest can lose its typical closed-canopy structure and accumulate large amounts of wood debris on the forest floor (Figure 6c). This gradient of gap sizes and resource/niche availability has relevant consequences for regional patterns of forest dynamics, biodiversity, and biogeochemical cycles.

Figure 6. Forest blowdown (total area of ca. 91 ha) in 2011 in Central Amazon, Brazil. Blowdowns can be identified on satellite imagery by geometric and spectral features such as diffuse shape and high short-wave infrared reflectance, indicating non-photosynthetic vegetation (NPV) resulting from widespread tree damage and mortality (A). The severity of the associated tree-mortality can be estimated using normalized ΔNPV (year of the blowdown – previous year) combined with field-measured tree
mortality (B). Edge of the blowdown/old-growth forest less than six months after disturbance, with toppled, survivor, and resprouting trees (C).

Tree mortality can be selective and depends on species traits and individual characteristics (Ribeiro et al., 2016; Magnabosco et al., 2014; Rifai et al., 2016). Snapping and uprooting of large individual trees can topple neighbors, altering the number and size distribution of trees and reducing stand biomass. Mortality rates among surviving trees are higher in the first years following the event, slowing biomass recovery. Resprouting and growth of survivor trees contributes little to biomass recovery, which can take decades (Magnabosco Marra et al. 2018). Recovery trajectories differ with the severity of mortality. However, even low severities trigger secondary succession, with substantial species turnover and dynamics distinct from those observed in small treefall gaps and human forest clearing (Chambers et al. 2009b; Magnabosco Marra et al., 2014, 2018). Soil organic carbon can also increase as a function of blowdown severity due to the decomposing organic matter available from wood debris (dos Santos et al., 2016).

Blowdowns can also promote tree diversity by providing niches to a diverse cohort of species that differ widely in their requirements and recruitment strategies (Magnabosco et al., 2014; Chambers et al., 2009). Nonetheless, altered functional composition indicates that blowdowns may affect the resilience of biomass stocks by favoring soft-wooded species with shorter life spans, which are also more vulnerable to future disturbances (Magnabosco et al., 2018; Trumbore et al., 2015). The impacts of blowdowns can be more pronounced in secondary and fragmented forests with altered composition and structure, and a relatively higher proportion of exposed edges (Silvério et al., 2019; Schwartz et al., 2017). That aspect is critical since these account for much of the remaining forests in Amazon (Brando et al., 2014; Hansen et al., 2013).

Research has focused on detecting blowdowns and quantifying their local to regional impacts on species composition, forest structure and dynamics. However, the effects of blowdowns on forest functioning at the landscape scale are still poorly understood. Assessing the return frequency of disturbances and the recovery rates of biomass and functional composition in different regions is critical. Climate change projections indicate that the frequency and intensity of convective storms could increase in the Amazon (Negrón-Juárez et al., 2017; McDowell et al., 2018; IPCC. Climate Change
Determining the possible thresholds of disturbance severity under these shifting disturbance regimes is thus critical, since it will affect the future vulnerability and resilience of the Amazon forest (Trumbore et al., 2015; Turner et al., 2010). The effects of forest blowdowns on other taxa remain unassessed in the Amazon.

**5.3.6.2 Lightning, natural fires and impacts on vegetation structure and biome distribution**

Lightning is an impressive and common phenomenon in the Amazon due to the meteorological systems that occur there, such as the squall lines and the SACZ. Natural fires can happen when electrical storms develop in conditions where vegetation is dry, especially when cloud-to-ground lightning is accompanied by little precipitation (conventionally ≤2.5 mm) (Viegas, 2012; Nauslar et al., 2013). This phenomenon, known as “dry lightning” or “dry thunderstorm”, also happens when the rain evaporates before reaching the ground, if a storm moves quickly or if cloud-to-ground lightning occurs outside the region where precipitation occurs (Dowdy and Mills, 2012).

Natural causes have been reported as an important ignition in the Cerrado, mainly due to cloud-to-ground lightning during the transition between dry and rainy seasons (Ramos-Neto and Pivello, 2000). There is still no conclusive information on the proportion of human causes and natural, but natural fires are believed to be around 1-2% of total fires (Alvarado et al., 2018).

The transition between the Amazon and Cerrado in Brazil has the largest area of contact between forest and savanna in the tropics, and these biomes differ fundamentally in their structural characteristics and species composition (Torello-Raventos et al., 2013). In this transition, rainfall seasonality and fire disturbances have an important ecological role on the vegetation structure and composition due to influences on the ecological and biogeochemical processes of vegetation directly affecting the Net Primary Production and respiration that, over time, lead to changes in composition and structure of vegetation (Alves et al., 1997). Fires change plants’ phenology and physiology and
modify competition among trees and lower canopy plants such as grasses, shrubs, and lianas. Depending on its frequency and intensity, fire occurrence may increase trees' mortality and transform an undisturbed forest into a disturbed and flammable one (House et al., 2003; Hirota et al., 2010; Hoffmann et al., 2012). Tree species associated with forest or savanna vegetation differ in numerous physiological characteristics, such as fire survivorship (Hoffmann et al., 2009; Ratnam et al., 2011) and their wood and foliar characteristics (Gotsch et al., 2010).

Couto-Santos et al. (2014) demonstrated the effects of climate variability and fire occurrence on forest-savanna boundaries dynamics in Roraima, in the northern part of the Brazilian Amazon. In wet years, the forest advanced over the savannas, while in years with lower rainfall, the forest receded, and the savanna expanded due to the increased frequency of drought and fire.

5.4. EVAPOTRANSPIRATION

When rainwater reaches the rainforest's land surface, most of it infiltrates into the soil, increasing soil moisture. About 50% of the rainfall returns to the atmosphere as evapotranspiration (ET: plant transpiration plus water evaporation from free water surfaces and bare soil; Table 1). The remainder supplies the groundwater pool, which ultimately contributes to the formation of the Amazon Basin’s streams and rivers. This section discusses the seasonal patterns of ET and their controlling mechanisms. The role of ET as a source of water to the atmosphere, and consequently for the processes of rain formation, is discussed in Sections 7.1 and 7.2.

An early attempt to characterize Amazonian ET was made during the Amazon Region Micrometeorological Experiment (ARME), a British-Brazilian experiment. Starting in 1983, this campaign made several micrometeorological measurements at the Ducke Reserve, about 30 km northeast of Manaus. Using ARME’s data and the Penman-Monteith equation, Shuttleworth (1988) showed a small seasonality in ET, with peaks in March and September that coincided with net radiation (Rn) extremes. The study also found that actual ET rates were nearly equal to potential ET rates throughout the year, suggesting plenty of water availability even during the dry periods.

Starting in the late 1990s, during the Large-Scale Biosphere-Atmosphere project (LBA), a network of intensive eddy-covariance (EC) measurements was set up throughout the
lowland Amazon to quantify surface energy, water, and carbon fluxes under different land covers (Keller et al. 2004). Data analysis from the EC flux towers revealed different ET seasonality depending on the study site. Most of the sites showed a seasonal pattern similar to that observed at Manaus during ARME – i.e., ET in phase with Rn, maintaining either a constant flux or showing a slight increase during the dry period compared with the rainy season (Costa et al., 2004; Hutyra et al. 2005; Juárez et al. 2007; da Rocha et al. 2004; Sommer et al. 2003; Souza-Filho et al. 2005; Vourlitis et al. 2002). A few studies, mostly located in the Southwestern Amazon (Aguiar et al., 2006) or at the transition between Amazon forests and cerrado savannas (Borma et al., 2009), observed higher ET in the rainy season compared with the dry season.

Syntheses of flux tower observations across the Amazon (Costa et al. 2010; Hasler and Avisssar 2007; Juárez et al. 2007); comparisons of the Amazon with other biomes (da Rocha et al. 2009); and a pan-tropical analysis (Fisher et al. 2009) helped elucidate the seasonal and spatial variability of Amazonian ET. Hasler and Avisssar (2007) found strong seasonality in ET for the stations near the equator (2°S-3°S), with ET increasing during dry periods (June-September) and decreasing during wet periods (December-March), both correlated and in phase with Rn. In stations located further south (9°S-11°S), ET and Rn did not present clear seasonality. These studies found the best correlations between ET and Rn at these sites during wet periods, but no correlation during dry periods. The authors attributed this response to water stress during dry periods, especially at the drier southern sites.

Negron-Juarez et al. (2007) analyzed ten LBA sites and concluded that all of them had higher ET during the dry period than during the rainy period. Fisher et al. (2009) analyzed 21 pan-tropical sites and observed an increase in ET in the dry period compared to the rainy period, with Rn explaining 87% of monthly ET variance. Da Rocha et al. (2009) analyzed ET data from EC flux towers at seven sites, four of them located in the northern Amazon Basin and three in the Cerrado (semideciduous forest, transitional forest floodplain, and cerrado). They observed that the seven sites analyzed could be divided into two functional groups in terms of ET seasonality. The southernmost sites, generally drier and with a long dry season, showed decreased ET in the dry period compared to the rainy period. Minimum ET values of 2.5 mm/day were observed in transitional forests, and a minimum of 1 mm/day was observed in the cerrado sites. The northern and more humid sites, with dry season length under four
months, showed the opposite pattern, with increased ET in the dry season and maximum values of around 4 mm/day. ET, Rn, and vapor pressure deficit (VPD) were positively correlated at these sites, suggesting that atmospheric conditions exert control over ET. However, it is important to consider that the most seasonal sites studied by da Rocha et al. (2009) had a predominance of deciduous and semi-deciduous vegetation. In these sites, the falling leaves in the dry period may have exercised important controls over ET, together with climatic conditions.

Costa et al. (2010) analyzed three evergreen rainforest wet equatorial sites (2°S-3°S) and two seasonally dry rainforest sites (at about 11°S). They observed that, in general, dry season ET is greater than rainy season ET. Following previous studies, they found that Rn was the main controlling factor of ET in wetter sites, followed by VPD and aerodynamic resistance. They identified different controlling factors of ET in wet and seasonally dry sites. While ET seasonality in humid equatorial forests was controlled only by environmental factors (i.e., abiotic controls), in seasonally dry forests ET was controlled by biotic parameters (e.g. stomatal conductance, gs), with surface conductance varying by a factor of two between seasons.

Observational studies generally agree on the seasonal pattern of ET in the Amazon rainforest, where ET is strongly dependent on net radiation (Rn) for seasonally humid forests. In the early 2000s, however, most models still simulated ET as being in phase with precipitation (Bonan, 1998; Werth and Avissar, 2004; Dickinson et al. 2006), suggesting that water availability limits ET. Around 2010, the LBA Data-Model Intercomparison Project (LBA-DMIP) compared the results of 21 land surface and terrestrial ecosystem models to the comprehensive observational dataset from the LBA network of flux towers to evaluate how well the new generation of models could reproduce the Amazon rainforest and Cerrado functions (de Gonçalves et al., 2013). As part of this project, Christoffersen et al. (2014) concluded that models have improved in their capacity to simulate the magnitude and seasonality of ET in equatorial tropical forests, having eliminated most dry-season water limitation. Their performance diverges in transitional forests, where seasonal water deficits are greater, but mostly capture the observed seasonal depressions in ET seen in the Cerrado. Many models depended only on deep roots or groundwater to mitigate dry season water deficits. Some models were able to match the observed ET seasonality, although they simulated no seasonality in stomatal conductance (gs). Some of these deficiencies can be improved by parameter
tuning, but in most models these findings highlight the need for continuous process
development (Christoffersen et al., 2014).

In summary, ET is controlled by the balance between water demand imposed by the
atmosphere (aboveground conditions) and the water supply in the soil (belowground
conditions). Both are considered abiotic controls (Costa et al., 2010) or ecohydrological
mechanisms (Christoffersen et al., 2014). By opening and closing stomata, plants may
exercise important additional controls over evapotranspiration fluxes through stomatal
canopy conductance (Costa et al. 2010; Christoffersen et al. 2014), resulting in a
These biotic (Costa et al., 2010) or ecophysiological (Christoffersen et al., 2014) control
mechanisms over ET and their importance in the context of regional climate will be
discussed in detail in Section 7.2.2

5.5. MAIN CHARACTERISTICS OF THE SURFACE HYDROLOGICAL
SYSTEMS IN THE AMAZON

The Amazon River basin (including the Tocantins River as a tributary) drains about 7
million km² and discharges about 16-20% of all global river inputs to the oceans
(Richey et al. 1989; see also Box 5.1). This vast hydrological system is formed by the
Andes, the Guiana and Brazilian shields, and the Amazon plain (Sorribas et al. 2016).
As a consequence of the seasonal rainfall cycle (Section 5.2.2), the main stem Amazon
River and its tributaries exhibit high and low river levels a few months after
the preceding wet and dry seasons. In general, rivers in the southern Amazon Basin
(e.g., Solimões, Madeira, Xingu, Tapajós, Tocantins-Araguaia) peak from April–May,
whereas rivers in the northern Amazon (e.g., Japura-Caquetá, Rio Negro) peak from
May–June (Espinoza et al. 2009a, b, Marengo and Espinoza et al. 2016). At annual time
scales, the hydrological contribution of southern and northern rivers is roughly
equivalent due to much higher total rainfall in the smaller northern basins compared to
the larger southern basins.
Box 5.1 – How Large is the Amazon River

"Born in the lofty, snow-clad Andes, the Amazon flows four thousand kilometers until it confronts the Atlantic at the equator. The Amazon is not only the world’s longest river; it carries more water than any other river – more than ten times that of the Mississippi, for example. One-fifth of all the water flowing off the face of the earth passes through the Amazon’s mouth. Such is the force of the Amazon as it clashes with the Atlantic that it pushes a vast plume of freshwater for hundreds of kilometers into the sea. Five centuries ago a Spanish explorer traveling up the coast of Brazil noted that at a certain point the sea tasted fresh, even though his ship was out of sight of land. Pinzón dubbed that spot the sweet sea (mar dulce), which historians and geographers take to be the mouth of the river, named after women warriors in Greek mythology... The Southern Equatorial Current pushes this turbid plume, which reaches some 400 kilometers long and between 100 and 200 kilometers wide, in a northwesterly direction up the coast of Amapá and the neighboring Guianas. Because it is lighter, the freshwater overrides the salty oceans and dilutes and muddies the surface for up to one million square miles." (Quoted from Smith 2002)

Most people know that the Amazon River is the largest river of the world. What most people do not realize is just how large it really is. This figure compares the world’s 10 largest rivers by discharge, showing the remarkable difference between the Amazon and all other rivers. The Amazon discharges about five times more water to the ocean than the world’s second largest river, the Congo. The magnitude of the difference is so striking that the Amazon’s largest tributary, the Madeira – discharging about 50,000 m$^3$/s to the main stem – would rank second among the world’s largest rivers if considered independently.
5.5.1. Seasonality of discharge

As noted above, the discharge of the mainstem Amazon River and its tributaries integrates hydrological fluctuations occurring upstream. These hydrological dynamics show a strong seasonal cycle that lags behind the rainfall cycle by a few months (See Section 5.2.2), with significant variations in the timing and magnitude of discharge across the Amazon’s tributary watersheds (Sorribas et al. 2016). The southern and western reaches of the Amazon River usually flood first, peaking between March and May. In the central Amazon, river levels are controlled by contributions from northern and southern tributaries, generally peaking in June (Figure 7). Long-term discharge measurements recorded near the central Amazon city of Óbidos, for example, indicate a peak discharge approaching ~250,000 m$^3$s$^{-1}$ during the high-water period in June, and a minimum discharge of ~100,000 m$^3$s$^{-1}$ during the low-water period in November (Goulding et al. 2003).

Because the northern headwaters of the Amazon are near the equator, their water levels fall between October and February, even as the Amazon River is rising due to contributions from the large southern tributaries. Small coastal watersheds of the northern Amazon (e.g., the Araguari) are also influenced by ocean tides in their lower reaches. In contrast, most of the Amazon River’s southern tributaries reach their highest levels in March or April (at points >300 km upstream from their mouths) and their lowest levels between August and October (Goulding et al. 2003). For example, discharge at Itaituba in the Tapajós River peaks at ~23,000 m$^3$s$^{-1}$ in March and reaches its minimum (~5000 m$^3$s$^{-1}$) in October (Figure 7). To its west, the Purús River at Arumã-Jusante shows even more pronounced variability, with a peak discharge of 11,000 m$^3$s$^{-1}$ in April and a minimum discharge of ~1000 m$^3$s$^{-1}$ in September (Coe et al. 2008). The lower sections of these southern tributaries are heavily influenced by a backwater effect of the Amazon River itself, rising and falling in response to changes in the main stem (Sorribas et al., 2016).
Figure 7. Annual cycle of river discharge for stations located in the main tributaries of the Amazon (in m$^3$ s$^{-1}$). The location of the main hydrological stations of the Amazon Basin is shown with red dots. Names of stations (and rivers) are indicated in each sub-panel. Elevation is shown in colors from <100 (blue) to >4000 (white) m.a.s.l. Sources of river level/discharge data: INMET and ANA (Brazil), SENAMHI (Peru), SENAMHI (Bolivia), INAMHI (Ecuador), SNO-HYBAM. The digital elevation model is provided by Shuttle Radar Topography Mission (SRTM) CGIAR CSI resampled at 250 m. Source of hydrological data: Laraque et al (2007), Espinoza et al (2009b, 2011, 2019) and Molina-Carpio et al (2017) based SNO-HYBAM.

5.5.2. Seasonality of floodplain dynamics

Fluctuations in rainfall and river discharge drive pronounced seasonal changes in the water level of large Amazon rivers, causing them to overflow their banks into adjacent floodplains. On a local scale, flooding can also result directly from rainfall in areas with poorly drained soils or rising groundwater levels, as in the case of the Llanos de Mojos.
in Bolivia. The periodic rise and fall of water levels – often referred to as the seasonal flood pulse – connects rivers and their floodplains during part of the year (rivers rise between November and June, and recede between June and November), resulting in heterogeneous habitat structure, rapid recycling of nutrients and organic matter, and high rates of biological production (Junk et al. 2012). The Amazon River and its large tributaries are characterized by a monomodal flood pattern with an average amplitude of 10 m near Manaus, ranging from 2 to 18 m depending on the location and year (Melack and Coe 2013). The greatest annual river-level fluctuations occur in the southwestern Amazon, especially the Madeira, Purus, and Juruá Rivers, while the smallest changes happen in the east. Small (low-order) streams in the Amazon lowlands exhibit complex hydraulics, with backwater effects resulting in a less predictable polymodal hydrological regime (Piedade et al. 2001).

The characteristic vegetation in these flooded regions is strongly influenced by hydrological dynamics, including maximum inundation extent, flood amplitude, and the duration of the low- and high-water phases of the flood pulse. On average, the lowland rivers of the Amazon are flooded for 6-7 months out of the year, with southern tributaries flooding from January-May and northern tributaries from June to August. Conversely, the southern Amazon undergoes a pronounced dry season from August to December, which generally coincides with the low-water period. In the north, floods can last until September (Goulding et al. 2003). Seasonally inundated wetlands thus cover an extensive (17%) area of the lowland Amazon – estimated at 8.4×10^5 km² of the region <500 m above sea level (Hess et al. 2015). About 44% of the wetland area is located in the Madeira and Negro watersheds, the Amazon’s two largest tributaries (Figure 2). The Marañon sub-basin has the highest proportion of total area as wetland (20 %), followed by the Madeira (19 %) and Içá-Putumayo (17 %). The Tapajós (5 %) and Xingu (8 %) sub-basins have the lowest proportion of wetland (Hess et al. 2015).

5.6. THE ROLE OF RIVERS IN BIOGEOCHEMICAL CYCLES

Rivers and related aquatic systems are key ecosystems in the Amazon region. The region’s underlying geology and landscape structure determine land-water connections via hydrological flow paths that influence river flow and chemistry. In disturbed systems, hydrological dynamics are strongly influenced by the type and intensity of land
use, which may alter rates of runoff, infiltration of water in soils, and water chemistry. Castello and Macedo (2015), considering river systems of different orders, stressed that soil attributes (chemical, physical, and biological) and land use are the main drivers of river biogeochemistry and metabolism. In small catchments, deforestation may increase inputs of nutrients, phosphorus, and carbon to aquatic environments, dramatically changing their natural functions. For instance, studies in small catchments identified extensive growth of an aquatic herbaceous species, leading to a high concentration of dissolved organic matter and, consequently, higher decomposition and respiration rates (Deegan et al., 2011).

The cascade from small to larger river systems depends on the extent of deforestation, soil type, and topography. Rivers are important providers of dissolved organic matter and nutrients to the ocean. This organic matter's chemical characteristics are key in defining its role in the coastal ocean’s metabolism. The Amazon River plume has a global influence. Recent data shows that 50 to 76% of the dissolved organic matter carried by Amazon waters to the ocean is stable (Medeiros et al., 2015), contributing to long-term storage of terrigenous carbon and potentially adding to the deep ocean carbon pool.

The biogeochemistry of carbon in aquatic systems involves production, transformations, and connections to terrestrial systems in environments ranging from small rivers to large river-floodplains. Small rivers, which are well connected to the surrounding watershed, are strongly influenced by riparian vegetation and biota. In the case of large rivers and their flood plains, on the other hand, the processes of carbon, nitrogen, and other nutrients are intensively modulated within the aquatic system (see also Section 6.2.2).

Changes in river flow and the frequency of floods and droughts are connected to changing climate patterns (Section 5.2), as are aquatic biogeochemical cycles. Martinelli et al. (2010) showed a decrease in the concentration of nitrogen species (dissolved inorganic and organic nitrogen (DIN and DON)) in aquatic systems in the Amazon with increasing river flow, but also noted the effects of changing land use and increasing population density (>10 people/km²) in the region. One important driver of nutrient flow to aquatic systems is the soil parent material and chemistry. On weathered, heavily leached tropical soils, vegetation cover is a key component in the nitrogen and
carbon cycles (Chapter 6). Nitrogen leaching to aquatic systems from “terra firme” may vary from 3 to 6 kg N-NO₃/ha/year with stream exports of around 4 kg-N/ha/yr (Wilcke et al. 2013). In contrast, in flooded areas where N is exported as dissolved NO₃ and NH₄, N exports can reach up to 12 kg-N/ha/yr. Lesack & Melack (1996) analyzed the impact of deforestation on nitrogen export to the aquatic system, finding an export of 2.7 kg N-NO₃/ha/yr for upland forests along the floodplain. After partial deforestation in the same area, measurements identified a 40% increase in nitrogen export in stream water, reaching 3.6 kg N-NO₃/ha/yr (Williams and Melack, 1997).

In contrast, dissolved phosphorus export is typically low. Values reviewed by Buscardo et al. (2016) indicate dissolved phosphorus export in streams ranging from 0.01 kg/ha/yr in a terra-firme forest (Leopoldo et al. 1987) to 0.006 kg P/ha/yr in an upland forest bordering a floodplain lake (Lesack and Melack 1996). Exports were an order of magnitude higher in a lower montane forest in Ecuador, reaching 0.6 kg/ha/yr (Wilcke et al. 2008).

5.7. CONCLUSIONS

The Amazon’s rainfall, river flow, and flood regime exhibit considerable variability at seasonal, interannual and interdecadal scales, with extreme flood and drought events becoming more common in the last two decades. Seasonal variability is mainly controlled by solar forcing. ENSO events are a major cause of interannual variation in rainfall, flow, and floodplain extent in the Amazon Basin. Central-Pacific El Niños (La Niñas) are related to rainfall deficits (excesses) over the upper part of the basin (Andean region of Colombia, Ecuador, and Peru), but these anomalies are weaker during Eastern-Pacific El Niño (La Niña) events. During Eastern-Pacific El Niño events, rainfall anomalies are stronger in the Madeira basin. The interannual modes of variability are modulated by interdecadal modes of the nearby oceans, such as the Pacific Decadal Oscillation and the Atlantic Multidecadal Oscillation. Moreover, extreme rainfall and flooding events are not necessarily associated with ENSO events.

Interactions between large-scale atmospheric circulation and orographic induced circulations result in high spatial variability of precipitation over the Amazon-Andean region, which may reach 7000 mm/year – the highest rainfall levels seen anywhere in the Amazon Basin. As a result of these interactions, the Andean basins also show the
largest runoff per unit area, and Andean rivers deliver sediments, pollutants, and nutrients downstream to the Amazon lowlands.

RECOMMENDATIONS

● The main processes of the Amazon hydroclimate system (convection, mesoscale circulations, land surface processes) are associated with the rainforest's presence. Preserving and restoring the Amazon forest is essential to the maintenance of these globally and regionally important processes.
● It is still unknown which factors are driving the recent acceleration in interannual climate variability, particularly given the interactions among deforestation, changes in atmospheric greenhouse gas concentrations, and natural modes of climate variability. Further research is needed to attribute the causes of this acceleration and to reduce uncertainties in that attribution.
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CORE GLOSSARY

Atlantic Multidecadal Oscillation (AMO): Climate cycle that affects the sea surface temperature (SST) of the North Atlantic Ocean.

Madden-Julian Oscillation (MJO): The main fluctuation in tropical weather on intraseasonal timescales. It is an eastward-moving pulse of clouds and rainfall near the equator that typically recurs every 30 to 60 days.

Pacific Decadal Oscillation (PDO): Leading Principal Component of monthly SST anomalies in the North Pacific Ocean.

Tropical Northern Atlantic Index (TNA): Anomaly of the average monthly SST from 5.5°N to 23.5°N and 15°W to 57.5°W, based on the climatology from 1971-2000.