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# Seasonal types in homogeneous rainfall regions of the Amazon basin

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## Abstract

Due to its size and geographical features, different average annual rainfall regimes co-exist in the Amazon basin, with distinct year-to-year variability dependent on regions within the basin. In this study, we define and explain the seasonal regional types of annual regimes, i.e., years with similar seasonal anomalies. Our work is based on a 205 rain gauge network distributed over 5 Amazonian countries, spanning a period over 30 years. Using a spectral clustering method, we identified 7 sub-regions within the basin in which annual rainfall regimes are spatially homogenous. For each sub-domain we estimate specific parameters that characterize the rainy season (onset and demise dates, sign and duration of rainfall anomalies). Finally, using spectral analysis we identify between 2 and 4 “seasonal type” of precipitation in these seven sub-domains. Most of these seasonal types are in phase with the large-scale atmospheric circulation, which explain the temporal link with rainfall anomalies. The seasonal types result of the superposition of interannual and intra-seasonal variability whose factors are then difficult to identify and attribute. Part of the rainfall anomalies characterizing seasonal types is related to the interannual variability of the sea surface temperature in the Atlantic or the Pacific oceans, especially in the North-East and South-East part of the Amazon Basin, while in other parts strong intra-seasonal and local factors have a larger impact. The same sign and duration of anomalies do not concomitantly affect the various regions of the Amazon Basin, confirming that one mode of variability does not homogeneously affect precipitation in different parts of the basin.

**Key words:** precipitation regimes, seasonal type, rain gauge network, spectral clustering, sea surface temperature

## I. INTRODUCTION

The Amazon basin (AB) is very often reduced to the mere image of evergreen forest continuously showered with large amounts of rain. However, the diversity of environments and precipitation regimes are complex and stems in particular from the size of the Amazonian watershed, which extends from the Andes to the Atlantic covering more than six million km<sup>2</sup>. Located in the intertropical zone between 5°N and 20°S, the AB regional climate is regulated by the South American monsoon, the displacement of the Intertropical convergence zone (ITCZ), and on their seasonality (Vera *et al.*, 2006). Average precipitation and rainfall regimes are highly contrasted (see, e.g., Figueroa and Nobre, 1990; Marengo, 1992; Espinoza Villar *et al.*, 2009 ; among others). According to the Köppen climate classification, the basin falls within a type A warm climate, with subtypes Af (tropical wet without a dry season) in the west and along the Negro River, and Am (tropical monsoon) and Aw (tropical savanna with dry winter) in the north, northeast, south and southeast. In the equatorial Amazon-Andes transition zone, several

48 seasonal rainfall regimes are identified in few kilometers due to the interaction between regional  
49 atmospheric circulation and the Andean topography (Laraque *et al.*, 2007; Espinoza *et al.*, 2009; Segura  
50 *et al.*, 2019). Between 50 and 70% of rainfall in the Amazon basin comes from the Atlantic via moisture  
51 advection, variable according to the season, through the ITCZ (Costa and Foley 1999; Durieux 2002;  
52 Wang and Fu 2002; Arraut and Satyamurty 2009; 2012; Satyamurty *et al.* 2013). Moisture flows enter  
53 the basin mainly to the east at around 45°W and to the north. Convective systems are able to reach the  
54 Andes thanks to contribution of synoptic circulation and evapotranspiration from the Amazon rainforest  
55 (Simões Reibota *et al.* 2010 ; Segura *et al.*, 2020). The Amazon rainforest feeds the atmosphere with  
56 moisture through evapotranspiration, and it is estimated that between 33% and 56% of precipitation  
57 comes from evapotranspiration (Molion 1975; Marques *et al.* 1977; Salati *et al.* 1978; Makarieva *et al.*  
58 2007; Satyamurty *et al.* 2013; Rocha *et al.* 2015). However, southern and south-western Amazonia  
59 differ, with a greater contribution from local recycling by vegetation in the former region than in the  
60 latter, which in turn depends on evapotranspiration in other parts of the basin (Staal *et al.*, 2018). Some  
61 of the moisture in the Amazon basin is also transported southwards towards the subtropical regions of  
62 southern South America via the Low Level Jet east of the Andes (Salio *et al.* 2002; Arraut and  
63 Satyamurty 2009) and then is also a source of moisture for the Plata basin. The seasonal precipitation  
64 regimes in the Amazon basin are however highly variable from one year to another when compared to  
65 the historical mean seasonal regime. Therefore, a high variability in the different parameters of the rainy  
66 season can be observed, for instance in its duration and the onset and demise of the rainiest period, the  
67 occurrence of dry spells, or in the intensity of intra-seasonal rainfall (e.g., (Moron, Camberlin and  
68 Robertson, 2013; Arvor *et al.*, 2017; Funatsu *et al.*, 2021).

69 On an interannual time scale, rainfall variability is strongly related to the ocean-atmosphere coupling  
70 modulated by anomalies of the sea surface temperatures (SST). The oceanic phenomenon most often  
71 implicated in South American precipitation variability is the El Niño Southern Oscillation (ENSO) in  
72 the Pacific, in which cold (La Niña) or warm (El Niño) phases result in rainfall excesses or deficits,  
73 depending on the region of the subcontinent (Kousky, Kagano and Cavalcanti, 1984; Aceituno, 1988;  
74 Marengo, 1992; Marengo and Espinoza, 2016; Sulca *et al.*, 2017 ; (Towner *et al.*, 2020) ; among others).  
75 However, the impact of ENSO is variable and depends on its duration, intensity, location in the eastern  
76 or central Pacific (Takahashi *et al.*, 2011; Andreoli *et al.*, 2016; Tedeschi *et al.*, 2016; Jimenez *et al.*,  
77 2021), and its association with temperature anomalies in the tropical North and/or South Atlantic. In  
78 addition, the role of the Atlantic on the South American rainfall is increasingly highlighted. Indeed, the  
79 warming of the tropical North Atlantic is related to a lengthening of the dry season in southern Amazonia  
80 and to severe droughts in 2005 and 2010 in the Amazon basin (Marengo *et al.*, 2008; Zeng *et al.*, 2008;  
81 Marengo and Espinoza, 2016; Arias *et al.*, 2020). Moreover, Espinoza *et al.* (2014) showed that the  
82 exceptional flood of 2014 in the southwestern AB was due to warm conditions in the Indo-Pacific Ocean  
83 and a strong warm gradient between the tropical and sub-tropical South Atlantic Ocean.

84 On the other hand, at higher frequency, other atmospheric modes than ENSO or coupling with Atlantic  
85 SST anomalies can cause strong precipitation variability. For instance, Mayta *et al.* (2019) showed that  
86 around 25% of the intraseasonal rainfall extremes in the Amazon basin are related to the eastwards  
87 propagating equatorially confined Madden Julian Oscillation (MJO; Madden and Julian, 1971). Other  
88 high-frequency mechanisms such as Rossby wave trains in the Southern Hemisphere have been  
89 proposed to explain the observed intraseasonal and synoptic rainfall variability. According to these  
90 results, extratropical wind intrusions into the southern Amazon and cross-equatorial wind regimes  
91 contribute to the weather patterns variability over the Amazon basin and associated rainfall (Wang and  
92 Fu, 2002; Paccini *et al.*, 2017), including the onset and demise of the wet season in southern Amazon  
93 (Fu *et al.*, 2013; Arias *et al.*, 2015; Espinoza *et al.*, 2021).

94 Several previous studies have focused on rainfall patterns, their spatial and temporal diversity in the AB  
95 and their link with SST anomalies, or with the high frequency phenomena described above, most often

96 with an emphasis on the study of extreme events. However, few studies have been dedicated to analyzing  
97 systematically the intraseasonal regional consistency of precipitation anomalies based on long-term in-  
98 situ rainfall stations.

99 The present work contributes to the analysis of intra-seasonal precipitation variability. To this end, we  
100 define a “seasonal type” that is cluster of years with similar intra-seasonal rainfall variability in a  
101 regional scale. Furthermore, we aim at explaining the modulation of these seasonal types through the  
102 comparison of rainfall anomalies with those of i) cloud cover - in order to better identify the major  
103 convection regions -, ii) the IWV circulation which favor or restrict precipitation, and iii) SST, especially  
104 in the tropical and subtropical Pacific and Atlantic oceans

105 The paper is organized as follows: The second section presents the data used in this study. Section 3  
106 introduces the methods used to distinguish first, homogeneous precipitation regions, then, seasonal types  
107 and relate the latter to elements of atmospheric circulation and Pacific and Atlantic SST. Finally, section  
108 4 presents the results and discusses them before a final section with conclusion and final remarks .

## 109 **II. Data**

### 110 **a) Daily Rain Gauge Data**

111 Rainfall stations offer long (more than 30 years) and relatively reliable time series making it possible to  
112 analyze the temporal variability of rainfall on a small scale, despite their sparseness in the AB and  
113 possible measurement errors. Rainfall from models derived satellite data also have limitations, on the  
114 one hand the time series rarely cover three decades and although progress has been made since the first  
115 generation satellite (Fassoni-Andrade et al., 2021), rainfall estimates are still difficult to calibrate.  
116 Indeed, rainfall estimations from satellite models often tend to be overestimated. As a matter of fact,  
117 even a satellite product from TRMM 3B42, which has one of the longest estimated series and among  
118 the best quality, tend to overestimate the mean annual precipitation (Michot et al., 2018) and  
119 overestimate precipitations particularly during the rainiest season in south, north, and northeast of the  
120 AB (Michot et al., 2018) and in west and center regions during dry season (Zulkafli et al., 2014; Zubieta  
121 et al., 2015; Michot et al., 2018). Another major issue for this work is that it is not totally reliable to  
122 entirely reproduce the seasonal differences of some regions of the AB like in the Marañón and the  
123 Ucayali subcatchments. For this reason we chose to use observed data, from a network of 205 rain  
124 gauges (Figure 1) at daily scale, for which advanced quality control has been implemented (more details  
125 can be found in Michot et al., 2019). The data from the stations cover the period from 1981 to 2013 and  
126 were obtained from the National Water Agency (ANA) and the National Meteorological Institute  
127 (INMET) in Brazil, the National Meteorological and Hydrological Institute (INAMHI) in Ecuador, the  
128 Hydrological Meteorological and Environmental Studies Institute (IDEAM) in Colombia, and the  
129 respective National Meteorological and Hydrological Services (SENAMHI) in Peru and Bolivia.  
130 Unfortunately, no such data could be collected in Venezuela, and in the southwest part of the Amazon  
131 basin (between 10°S-15°S and 76°W-69°W) because data from stations in these regions did not meet  
132 the criteria for the database used for this work (more details can be found in Michot et al., 2019). The  
133 rain gauge-based data is able to account for the diversity of mean annual rainfall even though it misses  
134 the rainfall hotspots in the south and southwest of the Amazon-Andes transition zone (Espinoza et al.,  
135 2015 ; Chavez and Takahashi, 2017), as shown by the data from the Tropical Rainfall Measurement  
136 Mission (TRMM) satellite product (Figure 1).

### 137 **b) Outgoing Longwave Radiation and integrated water vapor flux**

138 Outgoing Longwave Radiation (OLR) was used as a proxy of deep convective clouds and was obtained  
139 from the National Center for Atmospheric Research of the National Ocean and Atmospheric  
140 Administration (NCAR/NOAA; Liebmann and Smith, 1996). Integrated water vapor flux (IWV)  
141 reanalysis (Kalnay *et al.*, 1996) that can provide indications about the pressure fields, is from the  
142 National Center for Environmental Prediction (NCEP-NCAR ). The IWV is computed from the specific  
143 humidity and the horizontal wind (zonal -  $u$ ) and meridional ( $v$ ) from the ground to 300 h Pa (Peixoto  
144 and Oort, 1992; Satyamurty, da Costa and Manzi, 2013) as:

$$145 \quad F_w = \int qV p/g \quad (1)$$

146 where  $q$  is the humidity,  $V$  is the wind vector,  $p$  is the pressure, and  $g$  is the acceleration due to gravity.  
147 This is a vertically integrated value indicates whether atmospheric motions act to decrease (for  
148 divergence) or increase (for convergence) the vertical integral of moisture, over a given time period.  
149 High positive values of this parameter (i.e. large moisture convergence) can be related to precipitation  
150 intensification and floods.

### 151 c) Sea Surface Temperature data

152 The NOAA-CPC global monthly SST data at  $2^\circ \times 2^\circ$  horizontal grid resolution (Smith and Reynolds,  
153 2004) are used to analyze the variability of surface temperatures in the tropical Pacific and Atlantic  
154 oceans. The types of El Niño (EN ) and La Niña (LN) events to which we refer to describe the relation  
155 between rainfall and oceanic anomalies, are detailed in Takahashi *et al.* (2011), Andreoli *et al.* (2016),  
156 Tedeschi *et al.* (2016), Pillai *et al.* (2021).

## 157 III. Methods

### 158 a) The spectral clustering method

159 The spectral clustering method is used in order to i) classify the stations according to their annual regime  
160 and define homogeneous rainfall regime regions in the AB, and ii) classify the years within a region to  
161 detect the seasonal types.

162 The main principle of spectral clustering is to represent all rain gauges in separate nodes of a connected  
163 graph whose vertexes express the similarity between two nodes. The spectral analysis of this graph  
164 (i.e analysis of the matrices that represent connections inside the graph) enables it to isolate its main  
165 consistent groups. To compute the connection between two nodes into this graph, the basic solution  
166 consists in computing the usual Euclidian distance. However, for clusters where the separation is tricky  
167 and non-linear, we exploit the kernel trick. It consists in projecting data in a space other than the usual  
168 one (i.e. represented by time series of rainfall) where the separation between clusters is linear. Under  
169 some specific properties (see Camps-Valls and Bruzzone, 2009 for a complete theory of kernels) this  
170 projection can simply be done by changing the way one computes the connection between nodes. In  
171 practice, this is done using a Gaussian kernel where the connection between two rain gauges  $x_1$  and  $x_2$   
172 is given by:

$$173 \quad K(x_1, x_2) = e^{\frac{-\|x_1 - x_2\|}{\sigma^2}} \quad (2)$$

174 where  $\sigma$  is a parameter to fix. This kernel enables to efficiently separate highly non-linear clusters. By  
175 a non-linear separation, we refer to the idea that the boundaries that separate different clusters of rain

176 gauges are not linear in nature. Unlike linear separation, which assumes that clusters can be divided by  
177 straight lines or hyperplanes, non-linear separation acknowledges that clusters might be intertwined or  
178 have complex, curvy, or irregular boundaries that cannot be accurately represented by linear models.  
179 Non-linear separation is particularly important when dealing with data in high-dimensional spaces (as  
180 in our situation), where linear boundaries may not adequately capture the underlying cluster structure.  
181 The non-linear spectral clustering technique aims to uncover hidden patterns and structures of  
182 precipitations in the data, even when they are not easily distinguishable by simple linear separation.

## 183 **b) Determination of homogeneous regions based on their rainfall regimes**

184 Mean historical precipitation regimes are often computed based on a multi-decadal time series, but  
185 spatially averaging these series do not discriminate well between homogeneous and transition areas  
186 where regimes are less well individualized. Indeed, some regions may have the same year-to year  
187 variability of the seasonal regime and thus have a stronger spatial consistency, while in other regions  
188 the regime is more variable from one year to another. Our approach therefore aims to identify the areas  
189 where regimes are similar year after year. In order to do this, we first use the spectral clustering method  
190 to describe year by year the spatial distribution of rainfall regimes, from August to July of the following  
191 year, as most of the rainy seasons or wetter periods are centered on the austral summer. The geographical  
192 coordinates of the stations are not considered to avoid forcing spatial homogeneity. Indeed, even if  
193 spatial coordinates can be valuable in a clustering problem involving spatial data, in this particular  
194 situation where the goal is to obtain homogeneous groups of precipitation whatever the location, it is  
195 preferable not to use spatial coordinates to avoid introducing a bias that might cluster nearby stations  
196 together. This procedure leads to one classification per year. An open issue of data clustering for which  
197 no sound solution yet exists is the determination of the optimal number of clusters. In this study, we  
198 rely on the intra/inter inertia to obtain reliable clusters. A good clustering should indeed exhibit  
199 homogeneous data within each cluster and a large variability between different clusters (for example  
200 between the mean of each cluster). This notion of homogeneity can be quantified using the standard  
201 deviation of the data: the average of the standard deviations of the time series within each cluster gives  
202 the intra-class variance and the standard deviation between the centers of each cluster gives inter-class  
203 variance. Therefore, the ratio between the inertia among clusters and the internal inertia should be  
204 maximal. Applying this approach to each year of the series, the optimal number of clusters obtained is  
205 13 per year. Small clusters (less than five rain gauges) are gathered with the nearest cluster (considering  
206 the spatial distance and interclass variance), in this case the number of annual clusters is reduced 10 to  
207 6.

208 Finally, to analyze the pairing of the stations, i.e. to define which stations were most often part of the  
209 same cluster, we mapped the weight of the link between the stations through the time series thanks to  
210 the 'Cartography' package of the R-cran software (<http://rgeomatic.hypotheses.org/659>). This allowed  
211 us to visualize the sectors within which the annual regimes were most often similar. Conversely, stations  
212 with more random behavior were determined as belonging to a transition region.

## 213 **c) Detection of seasonal type in a homogeneous region**

214 Since the interannual variability of daily precipitation is very high, the intercomparison of annual  
215 patterns was achieved by performing a Fourier transform to smooth the data, removing frequencies  
216 higher than 60 days. On the one hand this frequency allows to observe, if it is the case, alternating  
217 rainfall anomalies during the season, while higher frequencies show only one anomaly direction. On the  
218 other hand, this time step may correspond to intra-seasonal atmospheric variability modes. Thus, a

219 smoothing of 60 days allowed us to keep a daily time step with a reduced noise so that we could observe  
220 more adequately the evolution of the annual regime of precipitation.  
221 From these smoothed annual regimes, we identified in each region the clusters of years that have had a  
222 similar rainfall regime, which correspond to our seasonal types, using the previously described spectral  
223 clustering method. The maximum number of seasonal types is arbitrarily set at four but without imposing  
224 any minimum number. The years are as before, defined from August to July of the following year, with  
225 the exception of the Northern Region where the rainy season is aligned to the boreal summer (Rao and  
226 Hada, 1990), i.e. from June to September. The number of years within these seasonal types is very  
227 heterogeneous, a greater homogeneity between them is obtained by eliminating the least representative  
228 years in each type. Years are eliminated when their distance from the mean regime is greater than a  
229 threshold which is the sum of the mean and the standard deviation of all distances from the historical  
230 mean seasonal regime. The choice of the number of representative years, seven, was defined by the  
231 median (Stavig and Gibbons, 1977) of the number of years of all AB seasonal types. One limitation of  
232 this approach is that we cannot speak in terms of magnitude of the anomalies because of the different  
233 number of years between the seasonal types compared. Our results therefore only allow us to analyze  
234 the duration and sign of the rainfall anomalies.

#### 235 **d) The seasonal parameters chosen to characterize different seasonal types**

236 The onset and demise dates of the rainy season are crucial parameters for agricultural planning, for  
237 example for determining sowing dates and whether more than one crop may be planted and harvested.  
238 In equatorial regions, for which there is no real dry season, dates delineating a wetter period were  
239 calculated.

240 In this study, we used the Anomalous Accumulation method developed by (Liebmann *et al.*, 2007)  
241 which is based on a threshold easily calculable for each region because it is the daily average of the  
242 series. Since the average is recalculated for each series, and here for each region, this method takes into  
243 account the local climatology (Liebmann *et al.*, 2007; Arvor *et al.*, 2014, 2017; Michot, Dubreuil and  
244 Ronchail, 2014; Dubreuil *et al.*, 2017) and allows us to catch the dates of the rainy season and their  
245 shifts. The Cumulative Anomaly Method has the great advantage of systematically providing dates at  
246 the beginning and end of the rainy season or a wetter period.

247 An adaptation has been realized to define the onset and demise of rainfall in the seasonal types in order  
248 to take into account a shift in their dates in relation to those of the average regional regime. The deviation  
249 from the average is not calculated in relation to the daily average of their own series but in relation to  
250 that of the region. This makes it possible to identify a deviation from the average regime of the region  
251 and not simply to identify the dates corresponding to a seasonal type.

252 Since it is difficult to determine thresholds for rainfall quantity the assessment of the onset and demise  
253 of the rainiest season is estimated using pentads as in Fu *et al.* (2013). The dates are therefore considered  
254 early or late when the shift exceeds one pentad, i.e. five days.

#### 255 **e) Relationships between rainfall and oceano-atmospheric variables**

256 Composite of seasonal anomalies of OLR and monthly anomalies IWV are the difference between the  
257 mean value of the representative years belonging to a seasonal type and the mean value of the  
258 climatological period 1981–2013 (from August y-1 to July y) and only the significant anomalies values  
259 (higher than 2 standard deviation; (Espinoza *et al.*, 2012; Takahashi *et al.*, 2021) are considered. The  
260 precipitation anomalies are normalized, that is, divided by the standard deviation of the whole study  
261 period. Rainfall regimes and seasonal types in the AB, discussions

## 262 a) Homogeneous precipitation regions versus transition regions

263 Figure 2 shows the frequency of the annual pairing of the stations for the period 1981-2013. A spatial  
264 distinction becomes clearer when stations are paired 50% of the time (Figure 2 b) i.e. when they are in  
265 the same cluster half of the time. When the threshold increased to 60% (Figure 2 c), there are isolated  
266 stations assigned to a transition sector as they have roughly one chance in two to belong to one group or  
267 another. Beyond a threshold of 60% the number of linked stations falls rapidly while the number of  
268 isolated stations increases sharply. This decrease can have several explanations, including poor data  
269 coverage. Since the spatial variability of precipitation on an intra- and inter-annual scale is significant,  
270 the further apart the stations are, the less likely it is that their annual regime has been the same, and that  
271 they are therefore grouped together several times in the same annual cluster. This explains why stations  
272 with few neighbors are linked fewer times to others and are therefore isolated. However, the northeast,  
273 north, northwest and southeast of the basin form organized groups up to a threshold of 70% (Figure 2  
274 d), indicating a greater spatial consistence of precipitation regimes in these regions. The final  
275 differentiation of the homogeneous regions results from the division of the groups of well individualized  
276 stations from there from 60%. It is important to note that these groups do not mean that all the stations  
277 are systematically linked together.

278 The regional classification (Figure 3) is also comparable to that provided by previous studies (Figuroa  
279 and Nobre, 1990; Espinoza Villar *et al.*, 2009; Barbosa Santos, Sérgio Lucio and Silva, 2015; Delahaye  
280 *et al.*, 2015). Tropical regions with dry season (Northeast, North, Southeast, South) are correctly  
281 distinguished from those with an equatorial climate (without dry season; Rio Negro, western Amazon).  
282 The homogeneous climatic regions based on the cluster analysis reflects well the AB intra-seasonal  
283 spatial variability, especially between the northwest and southeast of the Amazon basin, which is  
284 coherent with the eastwards propagation of the MJO described by Mayta *et al.* (2018).

285 In more detail, a distinction among the tropical regions is well made between (i) the North-East, where  
286 the maximum rainfall is observed between March and April, during the southernmost position of the  
287 Atlantic ITCZ (Espinoza *et al.* 2009) ; (ii) the North, with a rainy season centered in the austral winter,  
288 (iii) the South-East and the South, are grouped because of their wet season in austral season but are  
289 differentiated from each other due to their different amount of rainfall. Indeed, the incursion of extra-  
290 tropical cold fronts into the Bolivian Amazon in the austral winter (Garreaud, 2000; Seluchi and  
291 Marengo, 2000), results in greater winter rainfall in the south than in the southeast (as a proportion of  
292 annual rainfall). However, in the south, the variability of precipitation between different sectors of  
293 Bolivia (Roche *et al.*, 1990; Ronchail and Gallaire, 2006) associated with the small number of rainfall  
294 stations available for this study, leads to a fragmentation into annual clusters composed of very few  
295 stations. These were then grouped in the clusters that were spatially closest. The Andes and the Bolivian  
296 plain are thus united in the Southern region, despite the fact that rainfall variability also exists between  
297 these two sectors, as reflected in the boxplots associated with this regime in Figure 3.

298 In the equatorial Amazon-Andes transition zone the stations are gathered in a single region whereas a  
299 north-south split and a more pronounced bi-modal precipitation regime would have been expected  
300 (Laraque *et al.*, 2007 ; Segura *et al.*, 2019). The unification into a single region may be related on the  
301 one hand to the fact that geographic coordinates are not taken into account in the clustering analysis (the  
302 north and south are therefore not differentiated), and the shapes of the annual distributions of the rainfall  
303 series of this region being close enough to integrate them into a common cluster. On the other hand,  
304 according to the criteria applied in our approach, the division into homogeneous regions is based on the  
305 spatial continuity of the links between the stations (Figure 2), which remains continuous at 60% and  
306 even 70%, although the links are less important between the north and the south of this Andean region.  
307 This north-south continuous homogeneity agrees with the impact of the MJO convective core



308 propagation from the tropical Pacific to the northwestern South America, particularly identified in the  
309 Amazon-Andes transition zones of Peru and Ecuador (Recalde-Coronel *et al.*, 2020).  
310 The South and Andes regions have the highest heterogeneity rain gauges values as implied by the scatter  
311 observed on the boxplots and the highest AB coefficient of variations values (Figure 3). The full  
312 diversity of local regimes, more rainfall in lowland than on the Altiplano, may not be represented,  
313 however, this more complex reality could only have been captured by creating a large number of regions.  
314 A better division of these areas could in principle be achieved using satellite products such as TRMM  
315 3B42 version 7 (Arvor *et al.*, 2014; Michot *et al.*, 2018), however a long enough (three-decades) time  
316 series is not yet available.  
317 In the northwest AB the regimes are relatively homogeneous throughout the year, however the  
318 distinction between the Rio Negro region and the West region is due to a different timing of the maxima,  
319 comprised between April and June for the Rio Negro region and between February and April for the  
320 Western region, consistent with the quarterly rainfall percentages observed by Espinoza *et al.*(2009) in  
321 the basin.  
322 It is also regrettable that no stations were available in the south-western part of the basin (between 10°S-  
323 15°S and 76°W-69°W), where it has recently been shown that the annual variability of precipitation in  
324 the southern tropical Andes is linked to a change in the dominant mode of atmospheric circulation  
325 (Segura *et al.*, 2020). For the past fifteen years, this region has been experiencing heavier precipitation  
326 between December and February, certainly due to the strengthening of the meridional circulation  
327 between the tropical North Atlantic and tropical western South America.

## 328 **b) Description of the seasonal types**

329 The cluster analysis allowed to define subregions within the basin in which the annual rainfall cycle is  
330 spatially homogeneous. We explore now the seasonal types for each of these regions. Figure 4 shows  
331 for each region the historical mean seasonal regime (black line), the detected seasonal types (blue line),  
332 the precipitation anomalies calculated with respect to the mean seasonal regime of the region and  
333 standardized by the standard deviation of the climatology of the region (solid black) and finally, the  
334 dates of the beginning and end of the rainy season for the mean seasonal regimes (grey vertical lines)  
335 and for the seasonal types (orange vertical lines).

336 All regions, except the West, are characterized by 4 seasonal types. Some clusters include only one year  
337 (Figures 4b2 and 4c1) and are thus excluded as they correspond solely to an exceptional year.  
338 Furthermore, the Western region is totally excluded from the remaining of the analysis since only the  
339 anomalies of the year 1981-82 could be distinguished in a significant way whereas all the other years of  
340 the series are grouped in a single cluster. One of the reasons for this lower detection may be related to  
341 the average regime with constant precipitation during the year and that does not vary sufficiently from  
342 a year to another to be separated by the spectral clustering method and to the exceptional 1981-82 rainy  
343 season.

344 Rainy season dates are easily detected in the North-East, South-East, South and North where the  
345 seasonality is more pronounced (Figure 4a, 4b, 4c and 4d) and correspond to previous works (Kousky,  
346 1988; Gan, Rao and Moscati, 2005; Vera *et al.*, 2006 a; Liebmann *et al.*, 2007). They are also consistent  
347 with the profile of the curves representing the seasonal types of the region. In the Rio Negro, although  
348 the regime is equatorial (no dry season), the onset of rainier periods detected in February is coherent  
349 with the seasonal regime. However, the demises in July seems to further correspond to the end of the  
350 defined hydrological year rather than a climatological feature.

351 In the Andes, the average onsets and demises dates define a long rainy season due to the bi-modal regime  
352 of the region. These dates are not obtained for the seasonal regime with deficient rainfall throughout the  
353 year (Figure 4f1 and 4f3).

354 Table 1 summarizes the season dates (row) and the sign of the rainfall anomalies (column) of the  
355 different seasonal types. The seasonal types mainly highlight 3 kinds of rainfall anomalies i) deficit  
356 during the whole regime ii) excess during the whole regime iii) alternating sign of the anomaly. The  
357 latter is the most frequent, corresponding to 16 out of 21 seasonal types; the rainfall anomalies often last  
358 between 30 and 60 days. It is remarkable that there are not many common years between the seasonal  
359 types of the different regions and that for the years common to the different regions, the anomalies are  
360 not the same or out of phase; there is no coherence of the seasonal types neither in time nor in space (see  
361 Figure 4).

362 Regarding the parameters of season dates, we note that the onset of the seasonal types have higher  
363 temporal variability (later or earlier) than demise dates, particularly in the South and Rio Negro. This  
364 higher variability in the onset dates was also found by various authors (Marengo *et al.*, 2001; Fu *et al.*,  
365 2013; Nobre *et al.*, 2016; Arvor *et al.*, 2017) for the southern Amazon region. In general, an onset shift  
366 is not necessarily associated with a demise shift. When it occurs, an early (late) onset is most often  
367 associated with a late (early) demise and therefore with a lengthening (shortening) of the rainy season,  
368 as is the case for the seasonal type 4 in the North-East with rainfall excess during the whole regime ;  
369 for type 1 in the North with alternating signs of the rainfall anomaly; for type 4 in the South with  
370 alternating signs of the anomaly; for type 4 in the Andes with alternating signs of the anomaly. On the  
371 contrary, a late onset is most often associated with an early demise and therefore with a shortening of  
372 the rainy season, as is the case for type 1 in the North-East and type 2 in the South with deficit rainfall  
373 during the whole regime for the both types. The relationship between the date of onset and the extension  
374 of the wet season agrees with the results of Fu *et al.*, (2013) and Arvor *et al.* (2017) for the southern AB.  
375 The largest shifts are detected in the Rio Negro region, with an early onset of 31 days for the seasonal  
376 type 4 and a delay of 60 days for the seasonal type 1, while in the South-East, although three of the four  
377 seasonal type show onset or demise shifts, the average number of days is less important. Another  
378 important information is that a type of onset and/or demise shift is not specifically associated with a  
379 particular type of rainfall anomalies during the rainy season.

380 These predominant rainfall anomaly features (Figure 4 and table 1) already highlight that seasonal types  
381 with alternating anomalies are marked by the superposition of intra-seasonal and interannual variability  
382 which are difficult to disentangle and attribute to explicit drivers with large-scale or local processes. On  
383 the contrary the types with only excess or deficit of rainfall may be associated with large-scale and  
384 interannual variability drivers, for example, as a response to the interannual variability of SST.

## 385 **IV. Relationships between seasonal types and SST and** 386 **atmospheric circulation**

387 We provide a detailed observed atmospheric ocean conditions for selected seasonal types in the North-  
388 East and the South-East regions. The North-East region is strongly influenced by Pacific and Atlantic  
389 conditions (Liebmann and Marengo, 2001; Gan, Kousky and Ropelewski, 2004) whereas the South-East  
390 region is influenced by the South Atlantic Convergence Zone (SACZ), which, in turn, is partially  
391 modulated by Atlantic anomalies (Liebmann *et al.*, 1999; Doyle and Barros, 2002; Taschetto and  
392 Wainer, 2008). In these regions, the seasonal type 1, 3 and 4 in the North-East and the seasonal type 3  
393 in the South-East, are associated with specific sea surface temperature (SSTA) conditions.

394 [Figure S1 in supplementary material shows average conditions for OLR and moisture transport](#)  
395 [circulation, for easier comparison with the anomalies shown below.](#)

### 396 **a) North-East region**

397 The North-East is the Amazon region closest to the Atlantic ocean, and there were often observed  
398 precipitation anomalies related to cloud cover, IWV and SST anomalies (Fu *et al.*, 2001; Marengo, 2004;  
399 Tedeschi *et al.* 2016). Our analysis confirm that for all seasonal types a preponderant link between the  
400 ITCZ activity and position, and the precipitation in the region can be observed. For example, during  
401 December-February rainfall deficits of the seasonal type 1 (Figure 4a1) with 5 years and a late onset  
402 season, positive OLR anomalies are observed in Figures 5 i, j, k indicating a moderate migration of the  
403 ITCZ towards the south of the tropical Atlantic and over northeastern South America. This is related to  
404 a reduction of IWV over the AB in December and January due to weakening of the trade winds over the  
405 equatorial Atlantic, while the IWV east of the AB and south of south America strengthens (Figure 5 i,  
406 j, k). Indeed, very heavy precipitation days (that is, daily precipitation above 20 mm) have been found  
407 to be strongly and significantly anti-correlated to the vertically integrated VIMF divergence and to the  
408 upper-level (300-100 hPa) divergence, and significantly correlated with low-level (1000-850 hPa) zonal  
409 winds in northern AB (Funatsu *et al.* 2021). During the second rain deficit period (March to June), the  
410 composites in Figure 5 i, m, n, o show that OLR anomalies remain positive on the equatorial Atlantic  
411 while negative OLR anomalies are seen further north indicating few activity of the ITCZ or maybe its  
412 early northward migration. This early displacement could be due both to the strengthening of the  
413 southeasterly trade winds and weakening of the northeasterlies. In addition, divergence of the IWV from  
414 March over northeastern Brazil (Figure 5 l), and from May over northern AB (Figure 5 n) is observed,  
415 and the Low Level Jet (LLJ; Vera *et al.*, 2006 b) intensified in the south. As Marengo *et al.* (2012)  
416 pointed out, the relative position of the ITCZ during the rainy season plays an important role on the  
417 variability of rainfall in the AB. The atmospheric anomalies are related to El Niño events that occur 5  
418 times out of 6 years (Table 2 and Figure 5 a, b, c). These oceanic conditions alter the circulation of  
419 Walker and Hadley cells and lead to subsidence over the northeast Amazon. During the El Niño phase,  
420 the ITCZ is positioned further north and is less active on the AB (Hastenrath and Heller, 1977; Nobre  
421 and Moura, 1984; Nobre and Shukla, 1996).

422 Conversely, La Niña events lead to an intensification of the ascending branch of the Walker cell over  
423 the AB and promote convection (Barichivich *et al.*, 2018; Towner *et al.* 2020; Espinoza *et al.*, 2022).  
424 La Niña phases are related to a southward migration and more active ITCZ, driving rainfall excess on  
425 the north-east region of the AB (Marengo *et al.*, 2012). In this study, the excess rainfall during all the  
426 year and a longer rainy season of the seasonal types 3 and 4 (Figure 4a3 and 4a4) seem to be highly  
427 related to La Niña (Table 2) but also to north and south Atlantic anomalies. The combination of sea  
428 surface temperature anomalies however differs between the two seasonal type. In type 3, with 2 years,  
429 a shorter La Niña phase is associated with negative SSTA in the north Atlantic (NATL) and positive  
430 SSTA in the south Atlantic (SATL), whereas the La Niña phase of type 4 lasts longer and is only  
431 combined with positive SSTA of SATL.

432 In more detail, the La Niña phases of the seasonal type 3 are of the eastern and central type (Table 2;  
433 Andreoli *et al.*, 2016; Tedeschi *et al.* 2016; Pillai, Ramu and Nair, 2021). Thus, the increase in cloudiness  
434 in northeastern Brazil (Figure 6 i to o) is consistent with the precipitation excesses restricted to this  
435 region and observed in May by Tedeschi *et al.* (2016) during La Niña east type. According to them, the  
436 strengthening of the northeasterly trade winds and the weakening of the LLJ are more characteristic of  
437 a central La Niña, which is frequently related to an increased moisture convergence in the northwestern  
438 Amazon basin and to extreme floods in the Amazon River (Espinoza *et al.*, 2013; Marengo *et al.*, 2013).  
439 On the other hand, the negative SST in NATL (Table 2 and Figure 6 b and c), reinforced by the strong  
440 northerly trade winds, as well as positive SSTA in SATL (Table 2, Figure 6 a and Figure S2 in  
441 Supplementary material), participate to the strong migration and maintenance southward of the ITCZ,  
442 as also observed by Wang (2004) and thus in the precipitation deficit in April then later on a late rainy  
443 season. La Niña and negative SSTA in the NATL leading to excess rainfall in the northeastern AB are  
444 consistent with descriptions by various authors (Marengo and Hastenrath, 1993; Hastenrath, 2001;

445 Ronchail *et al.*, 2002; Souza, Ambrizzi and Coelho, 2004), but, the precipitation excesses could be  
446 further enhanced by the presence of warm SATL anomalies as observed by Nobre and Shukla (1996)  
447 and Fontaine *et al.* (1998). Indeed, similar to Huang and Shukla (2005) we observe that positive SATL  
448 SSTA are associated with a weakening of the subtropical high and associated trade winds in the South  
449 Atlantic (Figure 6 i to l). Furthermore, the year of 2008-09, which experienced exceptional flood in the  
450 AB and which is part of this seasonal type, is also described by Marengo *et al.* (2012) and Espinoza *et*  
451 *al.* (2022) with La Niña and positive SSTA in the SATL, a strong shift in the ITCZ, an intensified Walker  
452 circulation enhanced IWV from NATL, and increased precipitation in the northern and western AB.  
453 The seasonal type 4, with 7 years, seems to be highly modulated by long La Niña events (Figure 7 a to  
454 c) from August-1 to May, four of them being Central La Nina (Table 2). Cloudiness stretching over  
455 northeastern part of the Brazil and AB is consistent with the rainfall excesses described by Tedeschi *et*  
456 *al.* (2016) during central La Niña, and a stronger convergence over the northern, northeastern, and  
457 central AB, while LLJ weakens (Figure 7 i to l). On the other hand, positive SSTA (Table 2, Figure 7 c  
458 and d, and Figure S3 in Supplementary material) in the eastern equatorial Atlantic and SATL in March  
459 to May may favor convection in the ITCZ during this period and its maintenance to the south at the end  
460 of the rainy season, as confirmed by the negative OLR anomalies. The South Atlantic SSTA, which  
461 develops more from May to July (Figure S3 in Supplementary material), can then be the cause of the  
462 resumption of rainfall at the end of the rainy season. The temporality of oceanic anomalies therefore  
463 seems to strongly modulate the temporality of precipitation.

#### 464 **b) South-East region**

465 In the South-East region, seasonal type 3 (Figure 4b3) with 5 years is clearly related to the  
466 regional atmospheric circulation, and in particular the SACZ, which largely modulates the rainy season  
467 in this region. Indeed, in type 3 during the rainfall excess in October-November, negative OLR  
468 anomalies are observed over southeastern AB and Brazil indicating the enhanced presence of the SACZ  
469 (Figure 8 g and h). These excess rains are related to the decrease in IWV transport to southern South  
470 America in favor of the southeast.

471 Concerning Pacific SSTAs, 4 central La Niña events were present among the 5 years of this  
472 seasonal type, developing as early as August but ending early in February-April (table 3 and Figure 8 a  
473 to c). Negative ENSO phases are known to increase convection in the AB. Here, the short period of  
474 excess rainfall and enhanced convection could correspond to the short duration of the La Niña events.  
475 A study by Tedeschi *et al.* (2016) differentiating the action of the two types of La Niña showed relatively  
476 low precipitation anomalies in the southern AB in both cases. Similar to Tedeschi *et al.* (2016), the IWV  
477 anomalies of this seasonal type have characteristics akin of that of a central La Niña, with convergence  
478 over the northern and northeastern AB, anticyclonic flux anomalies east of Brazil and convergence over  
479 the AB. However, the La Niña phase ends while rainfall deficits have already occurred over the region  
480 from January to April. They correspond to positive OLR anomalies (Figure 8 j to m) over the extreme  
481 southeast of the AB, especially over southeastern Brazil and indicate a weak activity of the SACZ. These  
482 deficits are related to southeasterly flux anomalies that correspond to a decrease in IWV transport from  
483 the north to the southeastern region, where an anticyclonic-type anomaly is emerging, and a weakening  
484 of the SACZ. On the other hand, the flux anomalies supplying the ITCZ move towards the eastern SATL.  
485 This configuration continues through March and persists over the southern AB into April, explaining  
486 the early demise of the rainy season (Figure 4b3). Positive SSTA near the Brazilian coast from January  
487 to April (Figure 8 b, c and Figure S4 in Supplementary material) could help explain the occurrence of  
488 these atmospheric and precipitation anomalies while the La Niña phase is underway. It is possible that  
489 the Atlantic SSTA reflect a weakening of the eastern branch of the South Atlantic High and thus a  
490 weakening of the easterly flux, resulting in a weakened SACZ. These observations seem to be consistent  
491 with those of Doyle and Barros (2002) who show a less active SACZ in relation to strengthened IWV

492 towards South America and positive SSTA between 10 and 40°S, from the South American coast to  
493 10°E. Moreover, Taschetto and Wainer (2008) indicate that the influence of the SSTA between 20 and  
494 60°S in the Atlantic takes precedence over that of ENSO and modulates the position and intensity of the  
495 SACZ especially between December and February. Finally, positive SSTA east of SATL (table 3 and  
496 Figure 8 c), from February to April, could further explain the significant activity of the ITCZ over the  
497 Atlantic, the deflection of the flux east of SATL and the early demise of the rainy season in the Southeast  
498 region.

## 499 **Discussion and conclusion**

500 This analysis was performed using data from 205 rain gauges, based on which the spectral  
501 clustering method and geomatic tool allowed us to obtain seven homogeneous precipitation regions  
502 separated by transition areas, and then, within each region to detect clusters of years with similar rainfall  
503 regimes, which we referred to as seasonal types. The seven region revealed in this study show for  
504 example the separation between regions with or without dry seasons, between highlands and lowlands  
505 and between west and east. These regions also reflect the eastwards propagation of the MJO described  
506 by previous studies over the Amazon basin.

507 The seasonal types detected can be summarized by three kinds of rainfall anomalies i) deficit  
508 during the whole year ii) excess during the whole year iii) alternating sign of the anomaly, which is the  
509 most frequent type. We found that the lag of the onset or demise rainy season dates are not associated  
510 with specific rainfall anomaly. Furthermore, the onsets of the seasonal types are more often shifted to  
511 later or earlier (than the mean average date) than demise dates, particularly in the South and Rio Negro  
512 basin (northern Amazonia). A demise shift does not appear associated with an onset shift. When it  
513 occurs, an early (late) onset is most often associated with a late (early) demise, and therefore with a  
514 lengthening (shortening) of the rainy season.

515 IWV and OLR anomalies are generally consistent with the rainfall anomalies characterizing  
516 each seasonal types. In addition, in the North-East and South-East, some seasonal types are associated  
517 with SSTA in the tropical Pacific and Atlantic, that can help to understand a part of the interannual  
518 precipitation variability. In the North-East, as already showed in the literature, we confirmed the close  
519 relationship between precipitations anomalies and the ITCZ intensity and displacement. In this region,  
520 seasonal type corresponding to dry and shorter rainy seasons are frequently associated to El Niño events  
521 (mainly in central and some in the eastern Pacific), when the ITCZ moves less southwards and is less  
522 active. Conversely, longer rainy season is observed when La Niña events occur. However the influence  
523 of the Atlantic SSTA is noticeable, as they also are responsible for a longer rainy season, while rainfall  
524 anomalies vary depending on timing and signal of these SSTA. When the South Atlantic SSTA are  
525 positive, rainfalls remain in excess, however, when SSTA are positive in the South Atlantic and negative  
526 on the North Atlantic, rainfall anomalies are contrasted.

527 In the South-East the more relevant links between atmospheric and oceanic anomalies were  
528 observed for the seasonal types with alternating anomalies and an early demise of the rainy season.  
529 During La Niña conditions we observed at the beginning of the rainy season a more active SACZ and  
530 an excess of rainfall. During the rainy seasonal types rainfall deficits occur when the SACZ is weak  
531 because of a decrease of the IWV from the north and an anticyclonic anomaly of these flux in the region  
532 and South-East Brazil. These rainfall deficits and the weakening of the south Atlantic High and the  
533 SACZ could be related to positive SSTA near the Brazilian coast, which is consistent with the  
534 observations of previous studies. A relevant point to notice is that the temporality of the onset and  
535 demise of oceanic anomalies seems to strongly modulate the temporality of precipitation anomalies, as  
536 shown in previous studies (Tomasella *et al.* 2010; Marengo *et al.* 2012; Espinoza *et al.* 2013).

537 This article thus helps to summarize the diversity of regional rainy season types that can occur  
538 in Amazonia. Oceanic conditions can explain a part of the interannual rainfall variability, however, the  
539 superposition with intra-seasonal processes of rainfall variability remain a major issue to predict the  
540 precipitation anomalies (e.g. Mayta et al., 2019). That is why there is a need to continue to explore and  
541 understand the weight of the different factors behind these patterns of variability. First, for a better  
542 interpretation of the SSTA role on convection and I WV anomalies and on the precipitation of these  
543 seasonal types, it will be necessary to expand the analysis using complementary variables such as  
544 vertical motions in the atmosphere and divergence/convergence fields. Furthermore, we will need to  
545 determine the drivers that influence intra-seasonal variability and in particular to investigate the role of  
546 the MJO in modulating regional rainfall anomalies for each seasonal types here. This is relevant since it  
547 is known that MJO can have significantly affect seasonal rainfall in the tropics (Zhang 2005), especially  
548 in the SACZ area (Carvalho et al., 2004, Nogués-Paegle et al. 2000), and including extremes over the  
549 Amazon basin (Mayta et al., 2019) and tropical South America (Recalde-Coronel et al., 2020).

550 Our results also suggest a relevant role of local scale in modulating seasonal types. Therefore,  
551 we need to formulate some hypothesis, from land use changes (Debortoli et al., 2017; Arvor et al., 2017,  
552 Wongchuig et al., 2021., among others) to complex topography, mainly in Andean countries (Espinoza  
553 et al., 2015, Chavez and Takahashi 2017; Junquas et al., 2018). Certainly many other environmental  
554 parameters could contribute to influence the dates of the rainy season and rainfall anomalies (Fu et al.,  
555 2013; Yin et al., 2014; Arias et al., 2020) among them, previous dry season conditions are increasingly  
556 pointed to as factors directly influencing the mechanisms of transition to the wet season, and thus the  
557 shift in the demise of the rainy season (Yin et al., 2014; Marengo et al., 2022).

558 Thus, seasonal types provide an understanding of how rainfall patterns vary in the Amazon basin and  
559 are the first step in improving seasonal predictability. However, accurate prediction also depends on a  
560 good understanding of how multiscale factors modulate seasonal types which will be the subject of the  
561 next step.

562

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567

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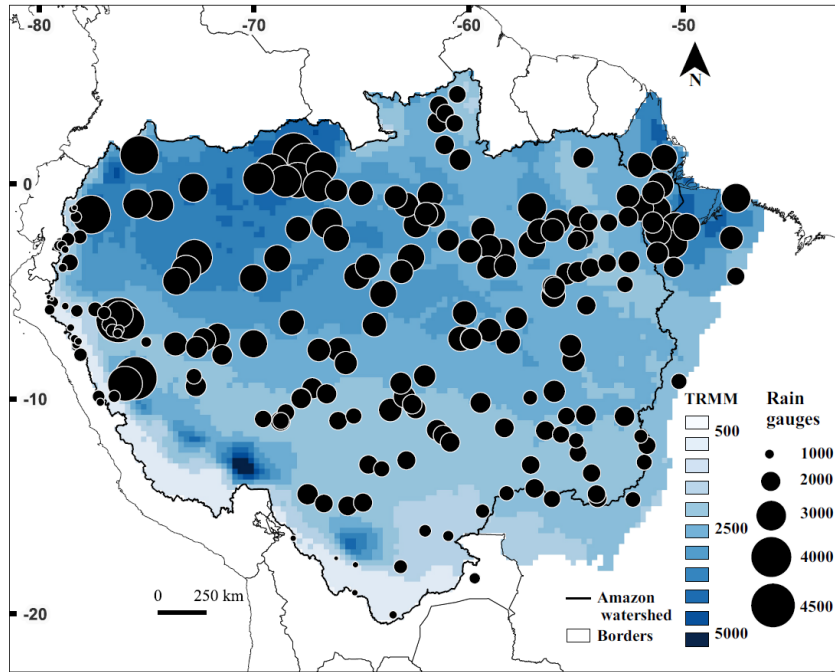


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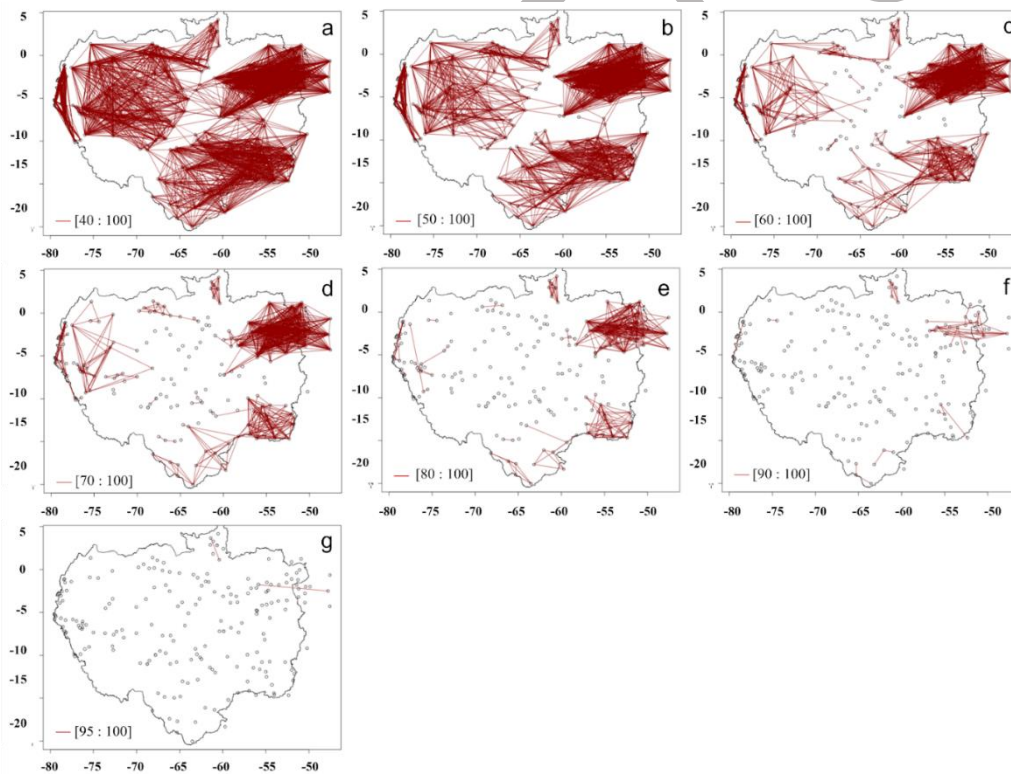
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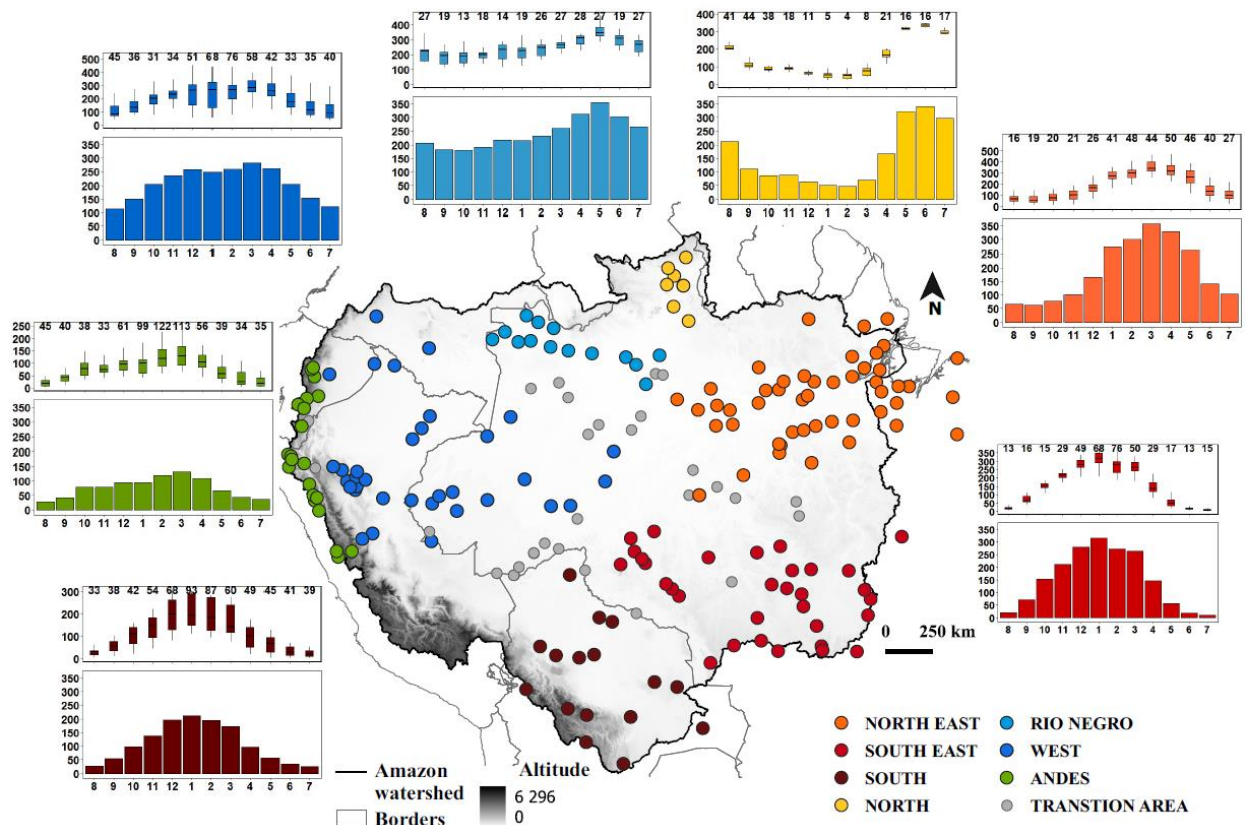
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809 *Figure 1 Annual mean rainfall in millimeters for the Amazon basin from TRMM B42 v7 (1998-2013) in blue and, from network*  
 810 *rain gauges (1981-2013) represented by the proportional circles.*

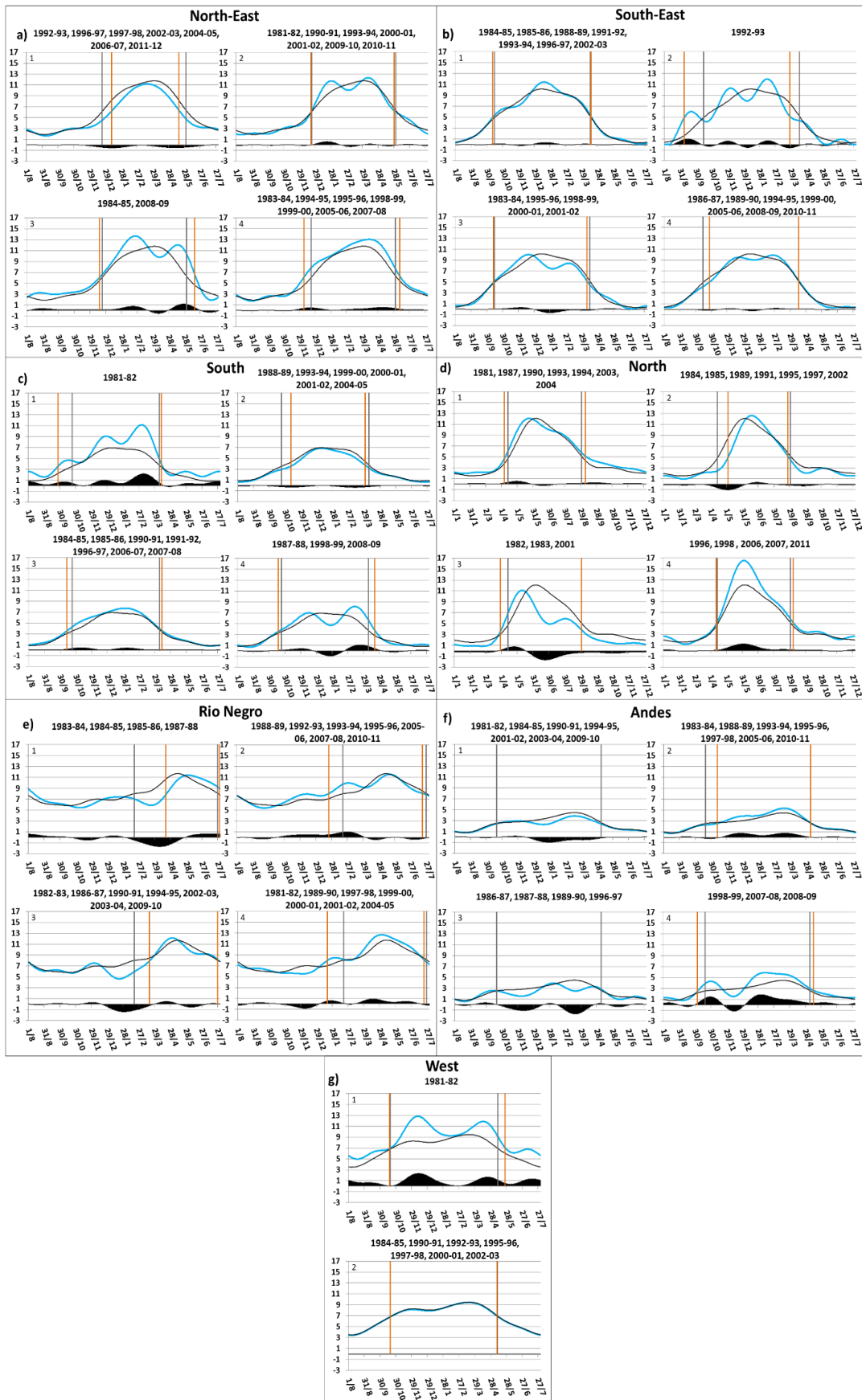


811

812 *Figure 2 Frequency of paired stations for the period 1981-2013 at least a) 40% of the time, b) 50% of the time, c) 60% of the*  
 813 *time, d)70% of the time, e) 80% of the time, f) 90% of the time, and g) 95% of the time. The gray dots represent the rain gauge*  
 814 *stations and the red lines materialize matching between two stations.*

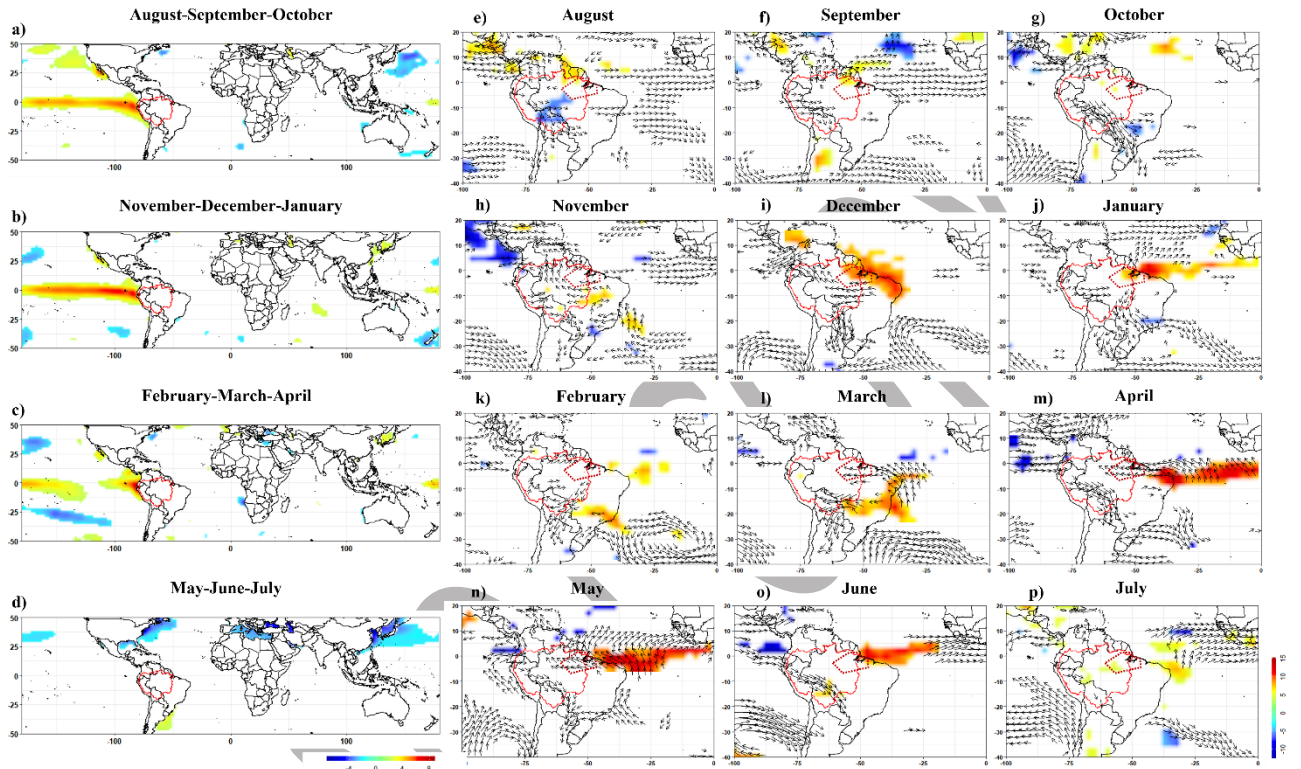


815  
 816 *Figure 3 Map of homogeneous rainfall regions, with the same average rainfall regime, and transition areas (grey dots) in the*  
 817 *Amazon basin based on spectral clustering method and results on Figure 2. The regions are formed from the stations that are*  
 818 *grouped in at least 60% of the times in the same annual cluster (1981-2013). The graphics show per region the historical*  
 819 *mean seasonal regime from August  $y-1$  to July  $y$  (axis  $x$ ) and the monthly dispersion of the station values ; the  $y$ -axis is*  
 820 *expressed in mm. The values above the boxplots express the coefficients of variation.*



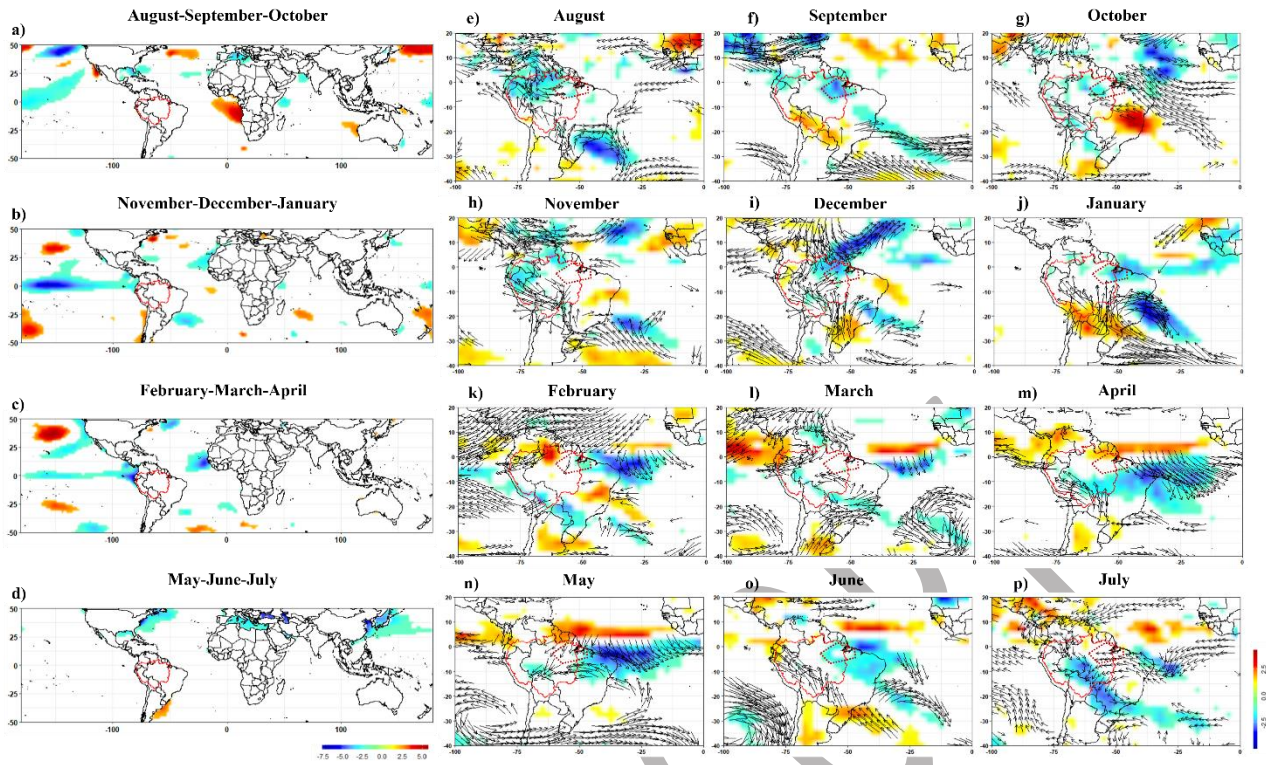
822 Figure 4. 1 to 4 seasonal types detected in the a) North East, b) South East c) South d) North e) Rio Negro f) Andes and g) West regions of the Amazon basin. Years of each seasonal type are indicated above each graph. The black line represents the average precipitation regime of the region (1981-2013), the blue line the seasonal type, and solid black the precipitation anomalies (blue line minus black line). The abscissae are expressed in days of the year and the ordinates in mm. The dates of the onset and end of the rainy season are symbolized by the grey vertical lines for the average regime of the region, orange for the seasonal type.

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830  
831 Figure 5 For seasonal type 1 of the North-East Region (Figure 4a1), seasonal composite maps of significant anomalies from August year-1 to July year-0 of SST (in °C; from a to d) in the left panels, and monthly composite maps of significant anomalies from August year-1 to July year-0 of OLR ( $W/m^2$ ) and IWV (vectors, in  $kg\ m^{-1}\ day^{-1}$ ) for South America (from e to p) in the right panels. Significant anomalies are greater than  $2\sigma$ . Anomalies are calculated with respect to the period 1981-2013. The red solid line represents the Amazon watershed and the red dotted line represents the North-East region.

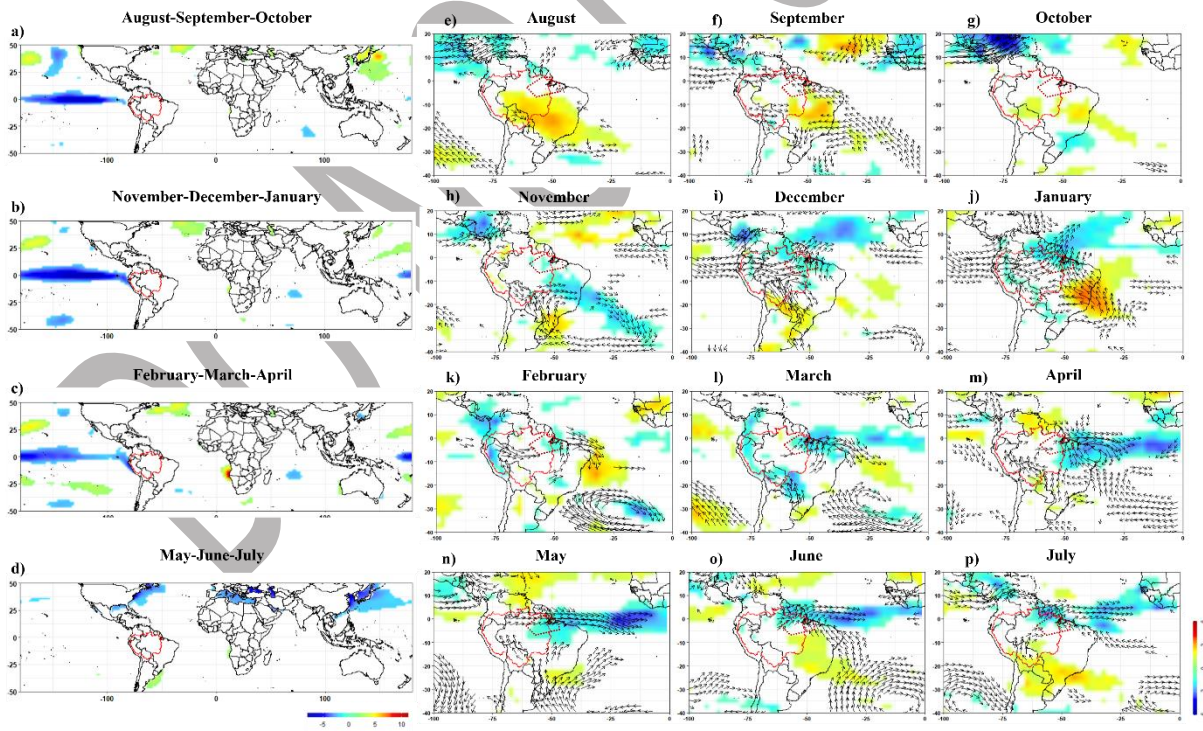




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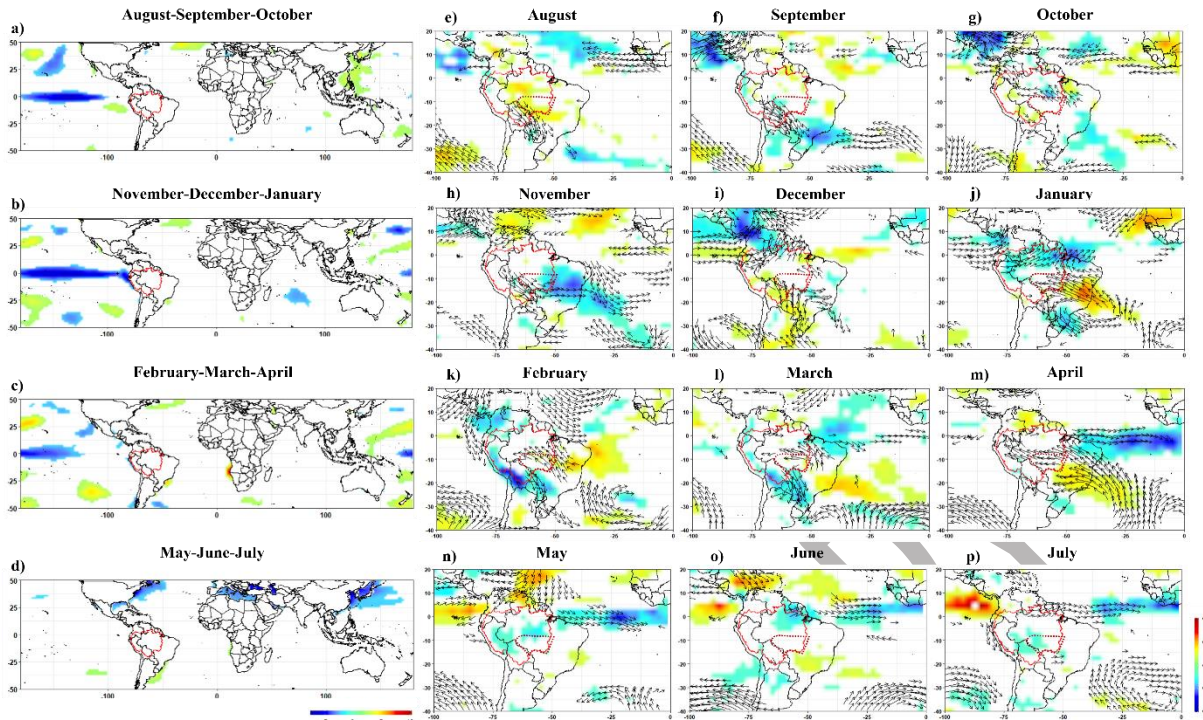
837 *Figure 6 Same legend as for Figure 5, but for type 3 (Figure 4a3) in the North-East Region.*

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*Figure 7 Same legend as for Figure 5, but for type 4 (Figure 4a4) in the North-East Region.*



841

842 Figure 8 Same legend as for Figure 5, but for type 3 (Figure 4b3) in the South-East Region except for the red dotted line that  
 843 represents the South-East region.

844

Table 1 Characteristics of seasonal type in the Amazon basin.

		Rainfall anomaly		
		Excess all season	Deficit all season	Alternating anomalies
Rainy season dates				
No rainy season dates		Andes type 2	Andes type 1	Andes type 3
Equal duration	normal onset + normal demise			South-East type 1 North-East type 2
Shifted season	early onset + early demise			South-East type 2 Rio Negro type 2
Shorter season	normal onset + early demise			South-East type 3
	late onset + normal demise			North type 2 Rio Negro type 1 Rio Negro type 3 South-East type 4
	late onset + early demise		North-East type 1 South type 2	
Longer season	normal onset + late demise	North type 4		North-East type 3
	early onset + normal demise	South type 1 Rio Negro type 4	South type 3	North type 3
	early onset + late demise	North-East type 4		North type 1 South type 4 Andes type 4

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846

847 Tableau 2 Synthesis of Pacific and Atlantic oceanic phases associated with Northeast Region seasonal types. The Central or  
 848 Eastern type of ENSO events is based on the classifications proposed by Andreoli et al. (2016), Tedeschi et al. (2016) and  
 849 Pillai et al. (2020). Event intensity is based on the CPC-NCEP NOAA Oceanic Niño Index (ONI): weak anomaly [0.5; 0.9];  
 850 moderate anomaly [1; 1.4]; strong anomaly [1.5; 1.9]; very strong anomaly  $\geq 2$ . The Atlantic phases are based on the index

851 provided by the NOAA/ESRL Physical Science Division and CIRES CU, Boulder, Colorado, from their website at  
 852 <https://psl.noaa.gov/>. The phase is considered positive or negative for an anomaly beyond 0.1°C. EN = El Niño; LN = La Niña.

Seasonal Types	Year	Pacific SST Phase	NTATL SST phase	STATL SST phase
1 : very dry and shorter rainy season	1992-93	Neutral	Negative	Negative
	1996-97	Neutral	Neutral	Negative
	1997-98	Very strong East EN	Positive	Positive
	2002-03	Moderate Central EN	Neutral	Positive
	2004-05	Weak Central EN	Positive	Positive
	2006-07	Weak East EN	Positive	Positive
	2011-12	Weak central LN	Neutral	Neutral
3 : contrasted anomalies and longer rainy season	1984-85	Weak East LN	Negative	Positive
	2008-09	Weak central LN	Neutral	Positive
4 : humid and longer rainy season	1983-84	Weak central LN	Negative	Positive
	1994-95	Moderate Central EN	Neutral	Positive
	1995-96	Weak East LN	Positive	Positive
	1998-99	Strong Central LN	Neutral	Positive
	1999-00	Strong Central LN	Negative	Positive
	2005-06	Weak LN	Positive	Positive
	2007-08	Strong LN	Neutral	Positive

853

854 *Tableau 3 Synthesis of Pacific and Atlantic oceanic phases associated with South-East Region seasonal type. The Central or*  
 855 *Eastern type of ENSO events is based on the classifications proposed by Andreoli et al. (2016), Tedeschi et al. (2016) and*  
 856 *Pillai et al. (2020). Event intensity is based on the CPC-NCEP NOAA Oceanic Niño Index (ONI): weak anomaly [0.5; 0.9];*  
 857 *moderate anomaly [1; 1.4]; strong anomaly [1.5; 1.9]; very strong anomaly  $\geq 2$ . The Atlantic phases are based on the index*  
 858 *provided by the NOAA/ESRL Physical Science Division and CIRES CU, Boulder, Colorado, from their website at*  
 859 *<https://psl.noaa.gov/>. The phase is considered positive or negative for an anomaly beyond 0.1°C.*

Seasonal Type	Year	Pacific SST Phase	NATL SST phase	SATL SST phase
3: contrasted anomalies and shorter rainy season	1983-84	Weak central LN	Negative	Negative
	1995-96	Weak East LN	Positive	Positive
	1998-99	Strong Central LN	Neutral	Positive
	2000-01	Weak LN	Negative	Negative
	2001-02	Neutral	Positive	Positive

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