

1 Authors  
2 Oliver Meseguer-Ruiz\*, Paulina I. Ponce-Philimon, Alexis Baltazar, Jose A. Guijarro, Roberto Serrano-  
3 Notivoli, Jorge Olcina Cantos, Javier Martín-Vide, Pablo Sarricolea

4  
5 Title  
6 Synoptic attributions of extreme precipitation in the Atacama Desert (Chile)

7  
8 Affiliations  
9 Oliver Meseguer-Ruiz  
10 Departamento de Ciencias Históricas y Geográficas, Universidad de Tarapacá  
11 2222, 18 de Septiembre, Arica, Chile  
12 [omeseguer@academicos.uta.cl](mailto:omeseguer@academicos.uta.cl)  
13 ORCID: 0000-0002-2222-6137

14 Paulina I. Ponce-Philimon  
15 Departamento de Ciencias Históricas y Geográficas, Universidad de Tarapacá  
16 2222, 18 de Septiembre, Arica, Chile

17 Alexis Baltazar  
18 Departamento de Ciencias Históricas y Geográficas, Universidad de Tarapacá  
19 2222, 18 de Septiembre, Arica, Chile

20 Jose A. Guijarro  
21 State Meteorological Agency (AEMET)  
22 Balearic Islands Office, Palma, Spain

23 Roberto Serrano-Notivoli  
24 Estación Experimental de Aula Dei, Consejo Superior de Investigaciones Científicas (EEAD-CSIC)  
25 Avda. Montañana 1005, Zaragoza, Spain

26 Jorge Olcina Cantos  
27 Instituto Interuniversitario de Geografía, Universidad de Alicante  
28 Edificio de Institutos Universitarios, Apartado de Correos 99, 03080 Alicante, Spain

29 Javier Martín-Vide  
30 Departamento de Geografía, Universidad de Barcelona  
31 Montalegre 6-8, 08001 Barcelona, Spain

32 Pablo Sarricolea  
33 Departamento de Geografía, Universidad de Chile  
34 Portugal 84, Torre Chica, Santiago, Chile

35  
36 \* Corresponding author

37  
38 Abstract  
39 Northern Chile is a region characterised by an extremely dry climate; however, there is a brief rainy season  
40 from December to March (austral summer), mainly above 3000 m a.s.l. It is interesting to consider where  
41 the humid air masses that generate such rain come from. For this purpose, daily precipitation data from 161  
42 meteorological stations located in this area (18°S to 19°S) were considered, and four clusters formed by k-  
43 means clustering. For each cluster, days of extreme precipitation (above 90th percentile) were selected to  
44 obtain flow strength (F), direction (D), and vorticity (Z) for each event according to the Jenkinson and  
45 Collison (JC) method. The back trajectory, for the previous 72 h, of air masses affecting the centroid of  
46 each cluster was determined by means of the Hybrid Single-Particle Lagrangian Integrated Trajectory  
47 (HYSPLIT) model. The analyses were carried out at sea level (1013 hPa) and in the middle (500 hPa) and  
48 upper (250 hPa) troposphere. Surface circulation was not a determining factor in the occurrence of extreme  
49 events, but it did influence circulation at 500 and 250 hPa. For stations located in the northern Altiplano,  
50 moisture advection from the Amazon basin is evident due to the configuration of the Bolivian high—an  
51 upper level anticyclone that develops over the Bolivian Altiplano during austral summer. For stations  
52 located in the southern part of the study area, the main source of moisture is the Pacific Ocean, and the  
53 weather is related to the arrival of frontal systems and to the configuration of cut-off low pressure systems  
54 in the mid-troposphere.

55  
56 Keywords  
57 Back trajectory, extreme precipitation, Jenkinson and Collison method, HYSPLIT model, Northern Chile

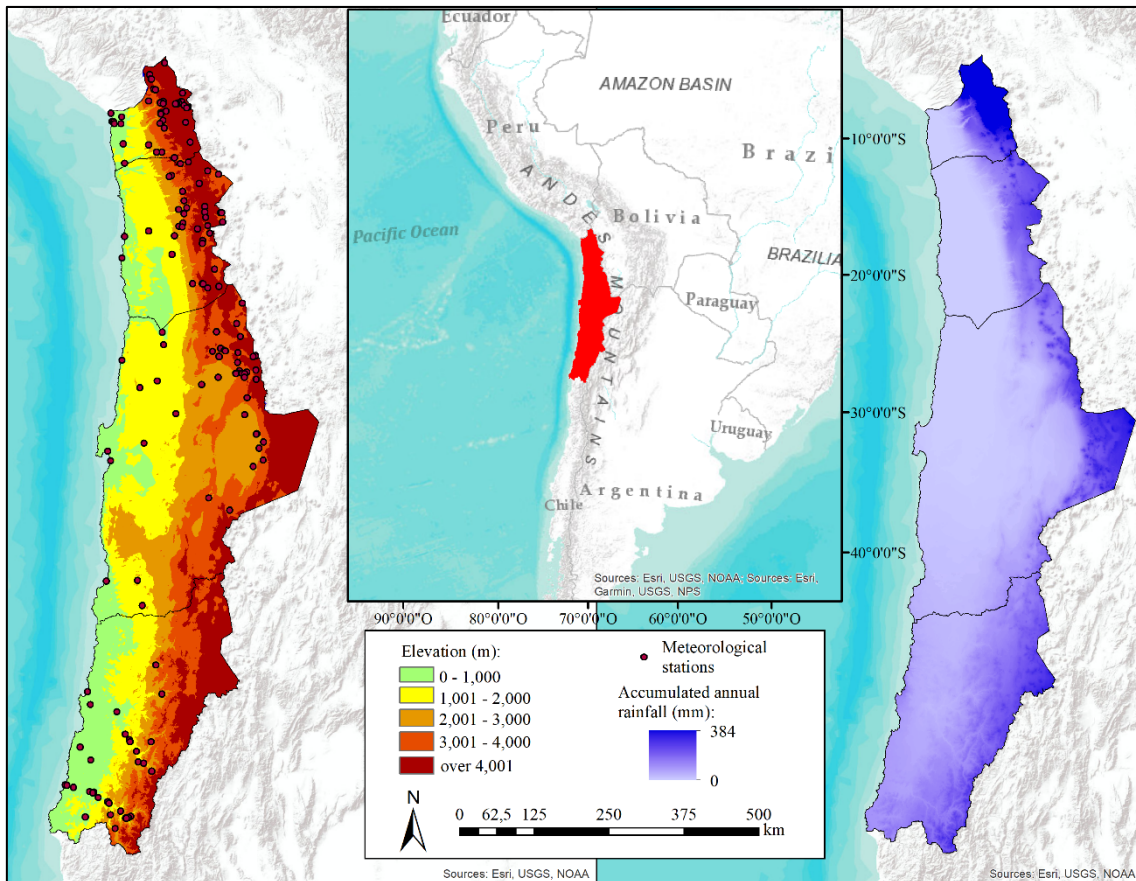
58  
59  
60

61 **1. Introduction**

62 The Atacama Desert is one of the most arid regions in the world, covering a vast area including parts of  
63 Chile, Argentina, Bolivia, and Peru. In addition to severe aridity, this region is characterised by complex  
64 orography. This is the case in Northern Chile (17°S to 29°S), a region with an extreme elevation gradient  
65 between the Pacific Ocean on the west and the ranges of the Altiplano and Andes mountains (Fig. 1,  
66 Sarricolea et al. 2017a). Arid regions, such as the one studied here, are more sensitive to climate change  
67 (Donat et al. 2016; Shukla et al. 2019). Coastal deserts such as the Atacama are also affected by cold sea  
68 currents (in this case, the Humboldt Current) contribute to the scarcity and irregularity of rainfall (Bombi  
69 2018; Rohde et al. 2019). In addition, rainfall projections in arid regions contain a high degree of uncertainty  
70 at regional and local scales (Williams 2017; Baez-Villanueva et al. 2018). This lack of consistency is  
71 characteristic of extremely high dry areas located at low latitudes (Kirtman et al. 2013), such as that found  
72 in northern Chile (Castino et al. 2017).

73 For example, rainfall exhibits irregularity at different time scales, from daily to multidecadal (Falvey and  
74 Garreaud 2005; Valdés-Pineda et al. 2018; Kock et al. 2020), that correspond to different mechanisms and  
75 the influence of the Andes mountains, which divide northern Chile into a coastal area in the west—under  
76 the influence of cold dry winds from the Pacific Ocean (Houston and Hartley 2003)—and a continental  
77 area in the east—influenced by warm wet easterly winds blowing from the Amazon basin (Garreaud 2009).  
78 The Andes mountains have a significant influence on moisture transport as well (Insel et al. 2010).

79



80

81 Figure 1. Map showing the topography of the study area and locations of the 161 meteorological stations  
82 employed (left), and the spatial distribution of mean annual precipitation (right) between 1950 and 2000  
83 (Plissock et al. 2014).

84

85 The El Niño-Southern Oscillation is one of the mechanisms modulating seasonal and interannual variability  
86 in precipitation in northern Chile (Garreaud and Aceituno 2001; Valdés-Pineda et al. 2015). The frequency  
87 of El Niño and La Niña events may change under various global warming scenarios as will its influence on  
88 the irregularity of precipitation (Marjani et al. 2019). On an intraseasonal scale, precipitation responds to  
89 the configuration of the Bolivian high, an upper-level anticyclone (250 hPa, Meseguer-Ruiz et al. 2019a),  
90 which develops during the austral summer over the Bolivian Altiplano, a high plateau region of the Central  
91 Andes. The proximity of the Bolivian high to the upper troposphere triggers the South American monsoon  
92 (Sarricolea and Romero 2015).

93 Changes in atmospheric circulation patterns modify precipitation patterns as well, contributing to  
94 irregularity in precipitation in this area. Some conclusive changes have been identified. A northward shift  
95 of the inter-tropical convergence zone (ITCZ) has been linked to changes in rainfall patterns in the Amazon  
96 basin (Schneider et al. 2014; Zappalá et al. 2018; Green et al. 2019). On a monthly scale, extreme events  
97 are linked to convection in the western Amazon basin during the wet season (Segura et al. 2019).

98 The wet season in northern Chile occurs in austral summer (between December and March). An increase  
99 in precipitation during this season is explained by an increase in vertical air instability and convection  
100 resulting from increased solar radiation and enhanced by topography (Junquas et al. 2016, 2018). At upper  
101 levels (250 to 200 hPa), the Bolivian High, related to the South American monsoon, introduces a humid  
102 easterly air mass from the Amazon Basin, which exhibits decadal and interdecadal variability (Segura et al.  
103 2016, 2020; Vasconcellos et al. 2020).

104 In the south of the study area, precipitation is also influenced by westerly flows linked to cut-off low  
105 pressure configurations in the mid-troposphere (500 hPa, Garreaud 2009). Different mechanisms occur at  
106 different tropospheric levels, and changes in tropospheric circulation modify the trajectories of the air  
107 masses, originating precipitation.

108 There remain many questions regarding the mechanisms operating at different tropospheric levels, the  
109 origin of the air masses that produce rainfall events (especially extreme rainfall events), and the origin of  
110 moisture. This is important because water is a very scarce resource in northern Chile (Sarricolea et al.  
111 2017b). Various factors, including mining urbanisation, have increased the demand for water based on the  
112 development strategy proposed by the country for the period 2017–2022 (Chilean Government 2015).  
113 However, this development strategy ignores potential future water supply problems, partly because  
114 precipitation projections can be misleading, exhibiting consistency at large spatial scales, but displaying  
115 significant uncertainty at local and regional scales (Power et al. 2012).

116 Factors causing extreme precipitation depend on physical parameters such as moisture availability, vertical  
117 stability, and event duration (Barlow et al. 2019). These factors in turn depend on mechanisms operating at  
118 different tropospheric levels and the origin of the air masses involved in such atmospheric processes (Chen  
119 and Zhai 2016). Meseguer-Ruiz et al. (2020) showed that daily precipitation anomalies in northern Chile  
120 were caused by changes in sea level pressure (slp) and geopotential height anomalies, but no research has  
121 been conducted regarding extreme precipitation.

122 The goals of this study were: 1) determine the origin of the air masses involved in daily extreme rainfall  
123 events (greater than 90th percentile of non-zero daily precipitation) in northern Chile, and consider the  
124 strength, direction, and vorticity of the flow at three different tropospheric levels (slp, 500 hPa and 250  
125 hPa) and 2) identify dominant moisture sources and pathways for each coherent region defined by back  
126 trajectories for the summer season according to the HYSPLIT model.

127 The present study states as follows. Section 2 describes the dataset quality control, the cluster analysis  
128 carried out, the calculation of the strength, direction and vorticity of the flow, and the HYSPLIT set-up.  
129 Section 3 presents the main results of the study and their comparisons with other studies. Their discussion  
130 was drawn in section 4, along with the conclusions.

## 131 132 **2. Data and methods**

133 To accomplish our objectives, we first validated data quality. Next, we performed a cluster analysis to  
134 identify areas of homogeneity within the study area. Once this step was completed, we selected the closest  
135 meteorological station to each cluster centroid and determined the values of the independent variables  
136 required for JC classification: direction (D), strength (F), and vorticity (W) (Meseguer-Ruiz et al. 2019b).  
137 Back trajectories were obtained for each day of extreme precipitation (in the 90% percentile) and for each  
138 reference station using the HYSPLIT model (Stein et al. 2015).

### 139 140 **2.1. Data quality**

141 Daily rainfall records from 161 stations across the study area for the period 1 January 1966 to 31 December  
142 2015 comprised the dataset. Quality control was performed using Climatol version 3.0 package in R  
143 software (Guijarro 2016), which uses normal ratio values (data divided by the mean of its series) from the  
144 closest precipitation data to build reference series for all stations. Differences between the observed and  
145 reference series were used to test data quality by outlier detection, and their homogeneity was assessed  
146 using the Standard Normal Homogeneity Test (SNHT) (Alexandersson 1986). At the same time, undoing  
147 the normalisation of the reference series provides estimations to fill in missing data in the series. The  
148 monthly aggregates of the series allowed to detect significant shifts in the mean. For arid climates, daily  
149 series show a high degree of variability, so detection of variations in the mean is very difficult (Vincent  
150 et al. 2002). These data were previously used and results published in a related study (Meseguer-Ruiz et al.  
151 2019a).

152

153 2.2. Cluster analysis

154 Cluster analysis using the k-means method of the data from 161 meteorological stations differentiated four  
155 distinct groups as a function of latitude, longitude, and accumulated rainfall, and the station closest to the  
156 centroid of each cluster was selected, one from each group. The decision to use four groups was based on  
157 the conclusions obtained in previous studies, and according to the altitude, latitude, terrain slope and  
158 orientation, distance to the Amazon Basin and concentration of the daily precipitation (Baltazar Fernandez  
159 and Meseguer-Ruiz 2019, Meseguer-Ruiz et al. 2019a).

160 To determine whether these groups depend on geographical location or on the amount of daily accumulated  
161 rainfall, the cluster analysis was reanalysed 1000 times by randomly rearranging accumulated rainfall  
162 values corresponding to each location while maintaining latitude and longitude. The resulting cluster  
163 assignments were compared with the original by summarising the agreement between groups by station and  
164 by cluster.

165  
166 2.3. D, F and Z calculation

167 The JC method of classification (Jenkinson and Collison 1977) is an objective classification of weather  
168 types and one of the most popular methods used to link atmospheric circulation and precipitation. It is an  
169 automated method that determines the type of atmospheric circulation from atmospheric pressure reduced  
170 to sea level in a day and can be adapted to different tropospheric levels (Meseguer-Ruiz et al. 2019b; Miró  
171 et al. 2020). Due to its simplicity, this method has been applied in many regions worldwide (Casado and  
172 Pastor 2016; Putniković et al. 2016; Fealey and Mills 2018; Putniković and Tošić 2018). The JC  
173 classification system derives from the values of and relationships between flow strength (F), flow direction  
174 (D), and vorticity (Z). Because of its simplicity, the system is also well-suited to evaluate the representation  
175 of weather types over Europe in a suite of climate model simulations (Otero et al. 2018). The behaviour of  
176 these variables has been linked to variability in precipitation intensity, focusing on the circulation vorticity  
177 in high elevation areas, such as the Tibetan Plateau (Li et al. 2019).

178 The variables which to be determined are the zonal component of the geostrophic wind (W), the meridional  
179 component of the geostrophic wind (S), the wind direction (D) in azimuthal degrees, the wind speed (F) in  
180 m/s, the zonal component of vorticity ( $Z_w$ ), the meridional component of vorticity ( $Z_s$ ), and the total  
181 vorticity (Z). These values were calculated at three tropospheric levels: slp, 500, and 250 geopotential  
182 heights, according to a 9-point grid (Fig. 2). The atmospheric pressure and geopotential height data were  
183 obtained for the 1966–2015 NCEP/NCAR Reanalysis Project (Kalnay et al. 1996). The analytical  
184 expressions adjusted for northern Chile are as follows:

185  
186 
$$W = 0.25[(P1 + 2P2 + P3) - (P7 + 2P8 + P9)] \quad (1)$$

187  
188 
$$S = 1,082[0.25(P1 + 2P4 + P7) - 0.25(P3 + 2P6 + P9)] \quad (2)$$

189  
190 
$$D = \arctan \frac{W}{S} \quad (3)$$

191  
192 
$$F = \sqrt{\frac{W^2}{S^2}} \quad (4)$$

193  
194 
$$Z_w = 1.273[(P1 + 2P2 + P3) - (P4 + 2P5 + P6)] - 0.829[(P4 + 2P5 + P6) - (P7 + 2P8 + P9)] \quad (5)$$

195  
196  
197  
198 
$$Z_s = 0.830[0.25(P1 + 2P4 + P7) - 0.25(P2 + 2P5 + P8) - 0.25(P2 + 2P5 + P8) + \quad (6)$$

199 
$$0.25(P3 + 2P6 + P9)]$$
  
200  
201 
$$Z = Z_w + Z_s \quad (7)$$
  
202



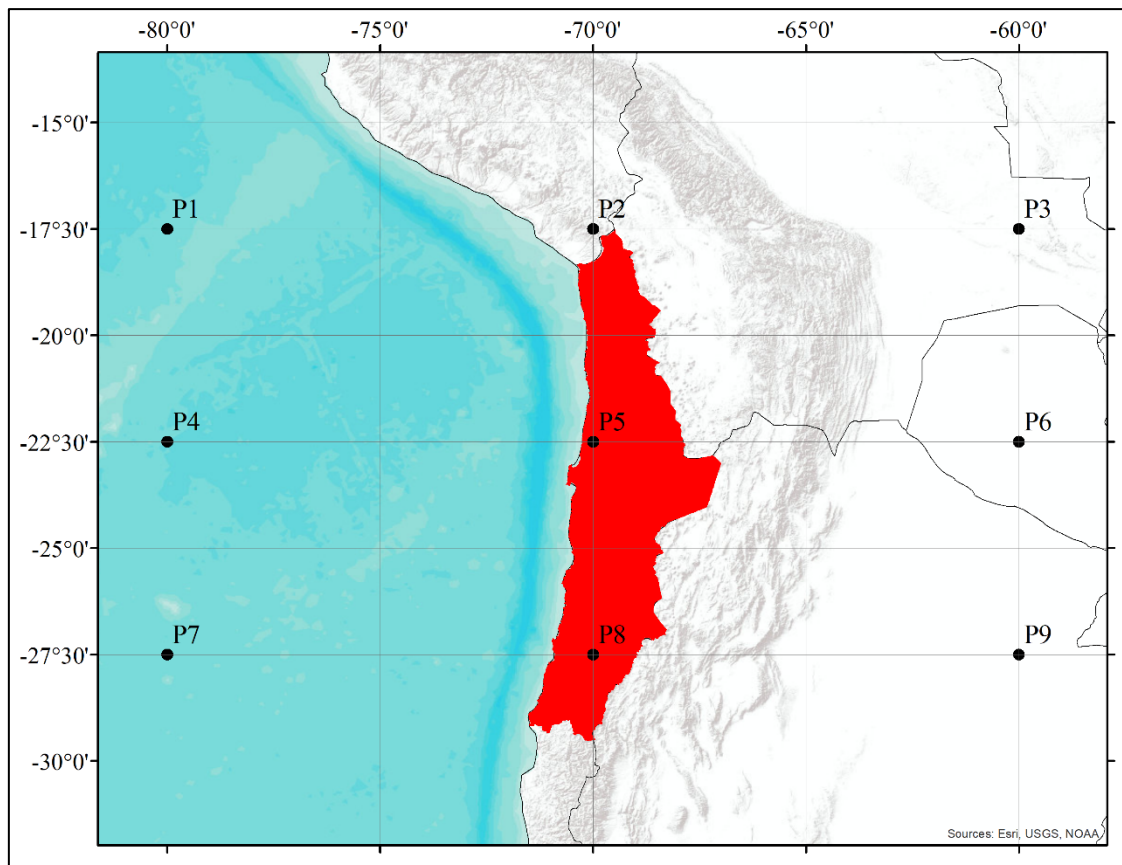


Figure 2. Location of the points used to determine the F, D and Z values.

First, we applied this methodology to obtain the behaviour of the independent variables (D, F, and Z) at three different tropospheric levels: SLP, 500 hPa, and 250 hPa. To calculate D, F, and Z at SLP, it is directly given by the above mentioned equations. For 500 hPa and 250 hPa levels, geopotential elevations were used. In this case, F and Z cannot be expressed in the same units or with the same range of magnitude as for the surface analysis because their calculation was based on geopotential height, but they do reference the above-ground flow behaviour, as shown in a study by Meseguer-Ruiz et al. (2019b). The selection of the JC classification method responds to its reliability determining the frequency behaviour of F, D and Z, and because it can be applied at different tropospheric levels and latitudes (Meseguer-Ruiz et al. 2019b).

#### 2.4. Back trajectories

The HYSPLIT model from NOAA (Draxler and Hess 1998) allows computation of simple air particle trajectories as well as simulations of transport, dispersion, and deposition (Stein et al. 2015). The HYSPLIT model is one of the most widely used in atmospheric sciences to determine the origin of air masses and define the source-receptor linkage (Gustafsson et al. 2010; Fleming et al. 2012). The use of gridded climate data allows the HYSPLIT model to define a pathway for an air parcel arriving at a specific location at a given time on a particular date (Tan et al. 2018).

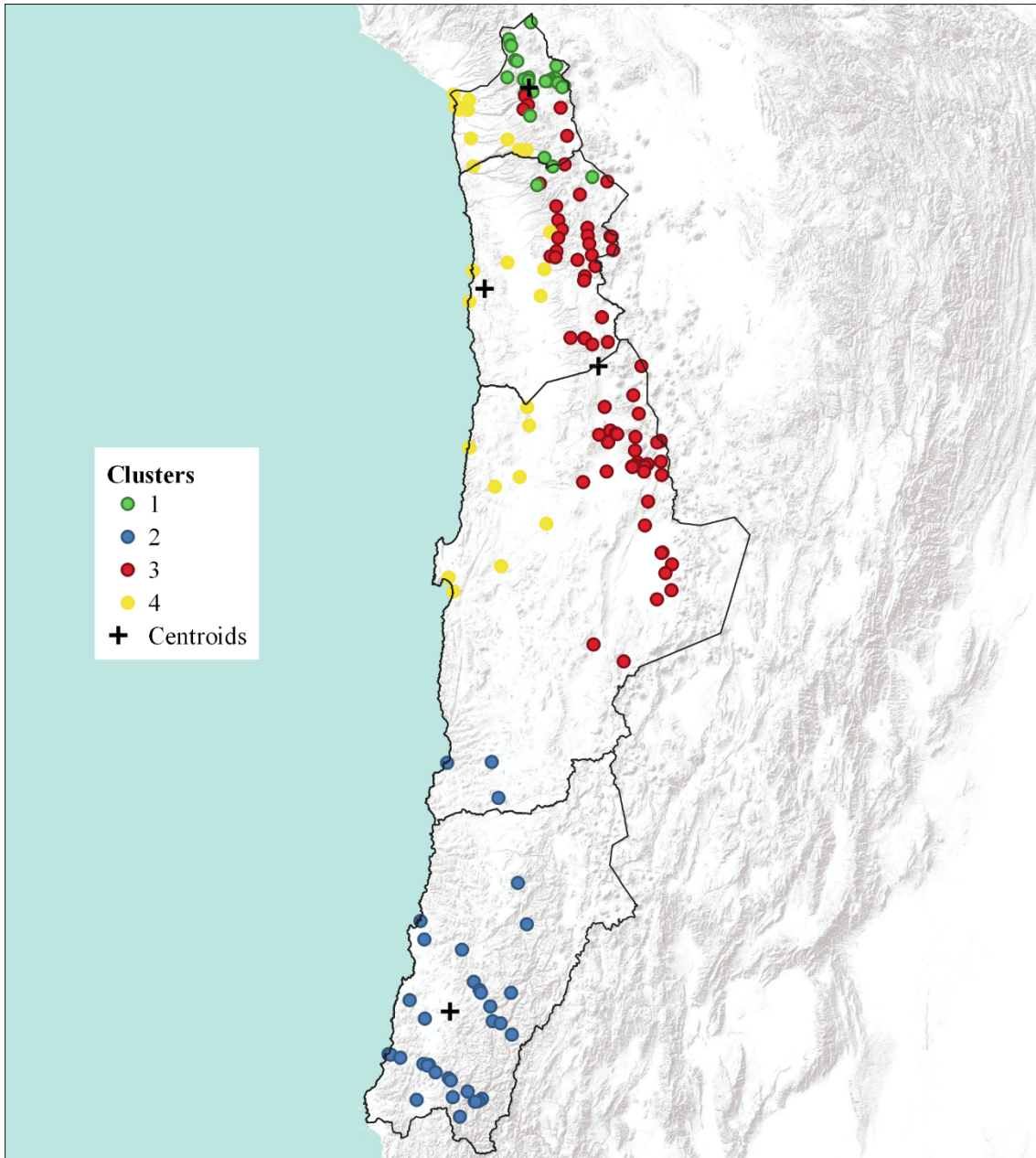
We used NOAA's HYSPLIT model to determine the back trajectories. Meteorological data were obtained from the NCEP/NCAR Reanalysis database with 2.5 degrees resolution. Due to the orographic complexity of the study area, the set-up of the model was based on mean sea level: 0 m at SLP, 5500 m at 500 hPa (to identify possible cut-off lows), and 12,000 m at 250 hPa (to identify the Bolivian High). With these settings, we obtained the back trajectories for 72 h, departing at 18:00 UTC-4 for each of the rainiest (90th percentile) days for the reference series of each cluster for the 1966–2015 period, following the methodologies applied by Gustafsson et al. (2010) and Bracken et al. (2015), but considering a 72 h pathway.

### 3. Results

#### 3.1. Grouping based on the daily rainfall

A cluster analysis that accounted for the latitude and longitude of the stations and the mean annual accumulated rainfall was performed. This procedure used a multivariate analysis with the aim of differentiating the various observatories into groups with a high degree of external heterogeneity and internal homogeneity. The final centres of the three variables considered in each cluster are listed in Table

236 1. The differences between the factors considered in each cluster were significant ( $> 99\%$ ), which  
 237 demonstrates external heterogeneity and internal homogeneity. The groupings are shown in Figure 3.  
 238



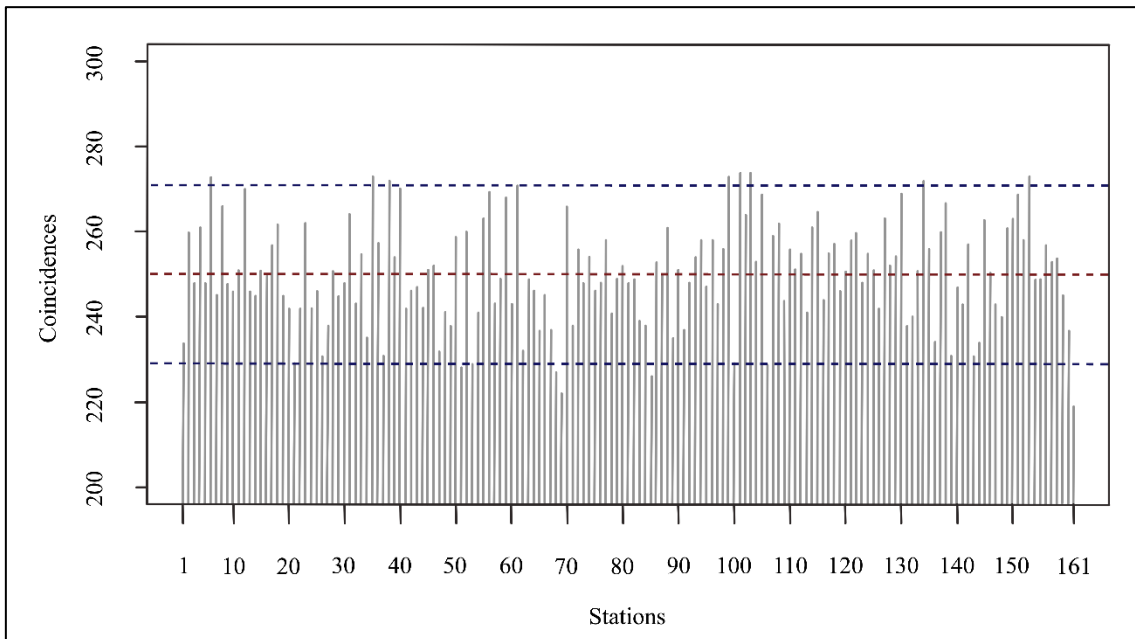
239  
 240 Figure 3. Stations grouping together after the cluster analysis  
 241

242 Table 1. Reference stations for each cluster (closest to the cluster's centroid).

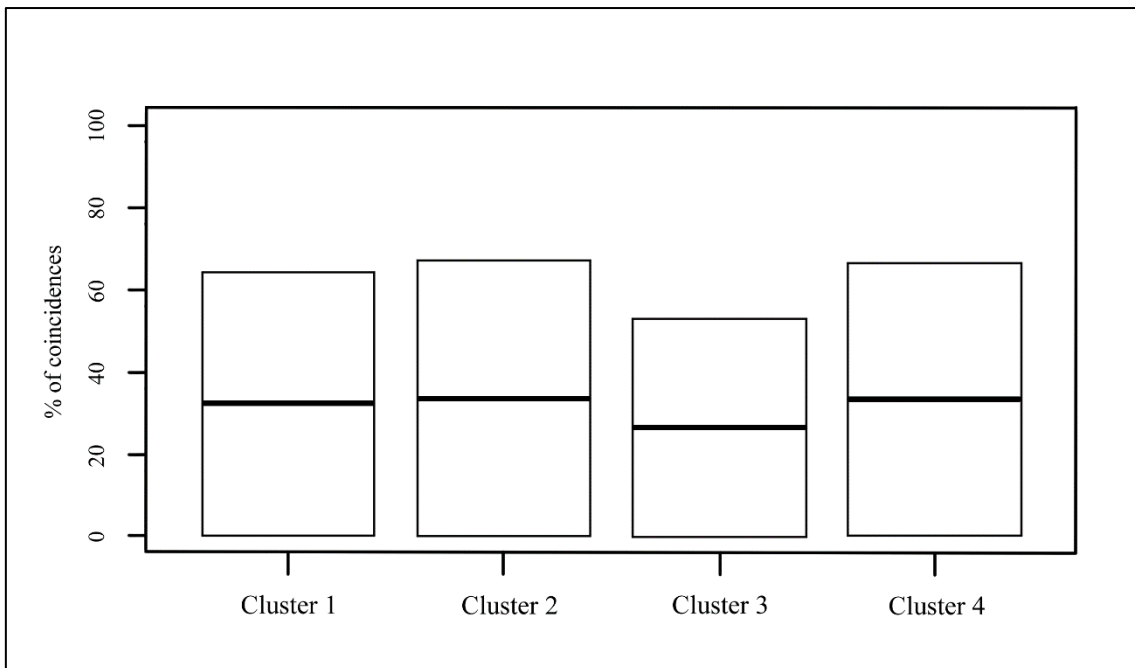
Cluster	Station	Altitude (m a.s.l.)	Latitude (° S)	Longitude (° W)
1	Portezuelo Chapiquia	4,375	18.32	69.47
2	Los Loros	940	27.83	70.11
3	Ujina	4,300	20.97	68.63
4	Iquique	50	20.21	70.13

243  
 244 The next analysis was performed to confirm that geographical constraints were not the sole cause of the  
 245 clusters obtained (Fig. 3), but clustering also depended on precipitation. After 1000 iterations to compute  
 246 simulated clusters with randomly rearranged precipitation values, comparison with the original groups  
 247 showed that the clusters were not the same in any simulation. The mean coincidence of each cluster group  
 248 between the original and the iterated cluster analysis by station (Fig. 4) was less than 35%. When we

249 considered the percentage of stations that coincided with the original clusters (Fig. 5), the average of all of  
 250 them is near zero. The iterated cluster analysis showed that the grouping was not solely geographic.  
 251 Longitude and latitude could have influenced cluster 1, but not exclusively. Instead, clustering mainly  
 252 depended on daily rainfall values. The observatories at Portezuelo Chapiquia, Los Loros, Ujina, and Iquique  
 253 were selected as representative of clusters 1, 2, 3 and 4, respectively.  
 254



255  
 256 Figure 4. Number of cluster coincidences by stations between observed and simulated in the cluster analysis  
 257 (grey bars) after 1,000 iterations. Red dashed line shows the mean number of coincidences and upper and  
 258 lower dashed blue lines show the 95th and 5th percentile, respectively.  
 259

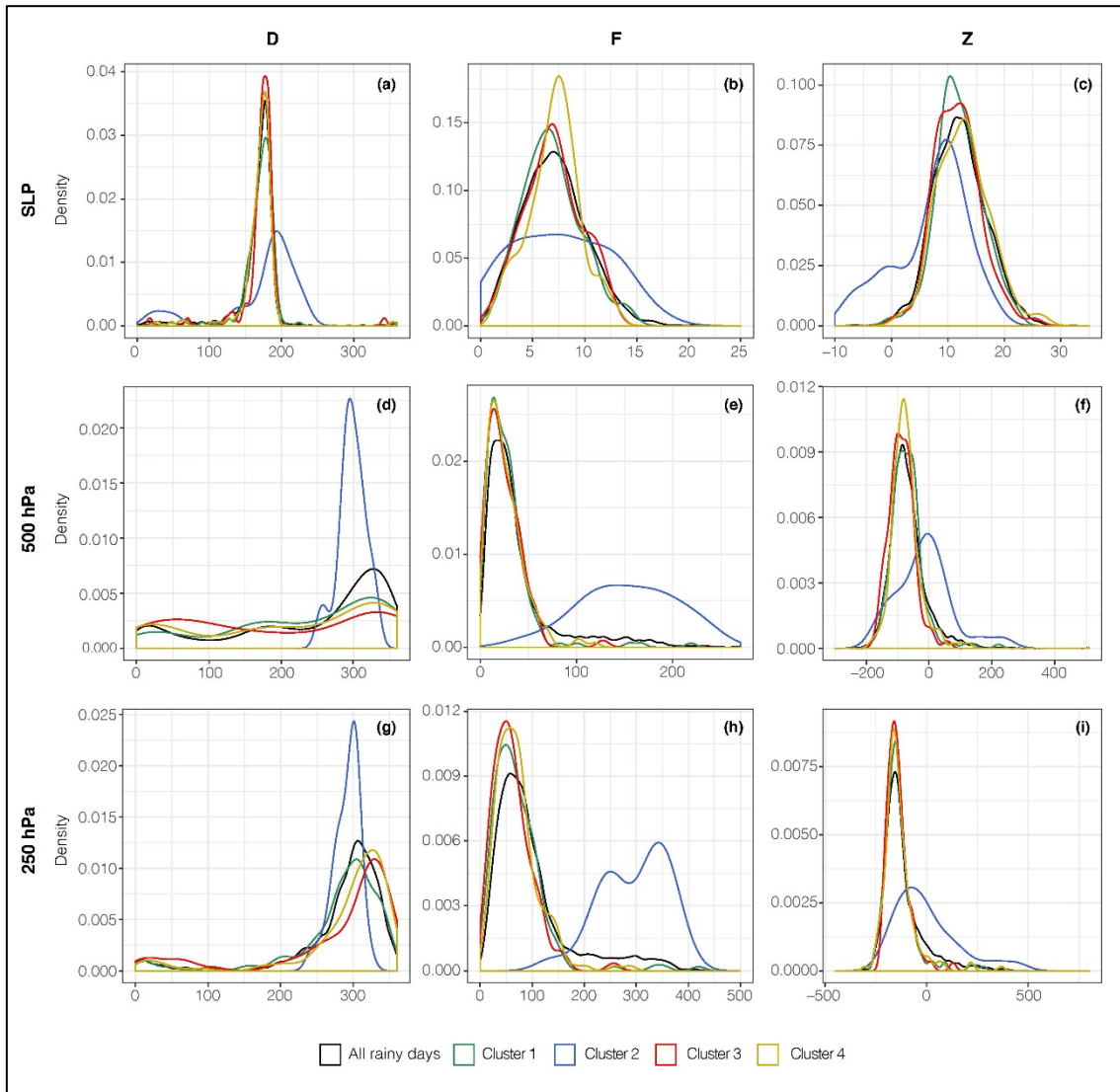


260  
 261 Figure 5. Boxplot of the percentage of coincident stations by clusters in the same analysis  
 262

### 263 3.2. D, F and Z frequency behaviour

264 The frequency of D, F, and Z was expressed as the kernel density for all the rainy days and for the extreme  
 265 precipitation events for each cluster at each of three tropospheric levels (Fig. 6).  
 266 At slp, the most frequent D for rainy days in the period 1966–2015 had a southerly component. Flow F  
 267 ranged from 6 to 8 m/s, and Z had frequent positive values that varied between 8 and 11 (black lines).  
 268 Considering extreme precipitation events, cluster 1 showed a clear predominance of a southerly component

269 for D between 180° and 195°, and F values were approximately 6 m/s. Z values (always positive) were most  
 270 frequent at 10 (green lines). Cluster 2 shows a high frequency at 200°, not centred at 180°, but with some  
 271 cases at 25° and 130°, and a lower frequency for rainy days. F values exhibited a wide distribution from  
 272 low flow strength (0 to 3 m/s) to high flow intensity (>15 m/s). The highest frequency for Z values was 10,  
 273 but the important frequencies were negative values (from -10 to 0). Clusters 3 and 4 showed the same  
 274 dominant southerly component, and flow strengths were most frequent for 7 and 7.5 m/s, respectively. The  
 275 Z values for both clusters were almost always positive (with very few negative values), and more frequent  
 276 for values around 13 in both cases.  
 277



278  
 279 Figure 6. Variation of D, F and Z at slp, 500 hPa and 250 hPa for all the rainy days (black line) of the study  
 280 period (1966-2015) and for the days of extreme precipitation in each cluster (coloured lines)  
 281

282 At 500 hPa, the most frequent flow direction was between 290° and 315°, F values were less than 50 m/s  
 283 and Z values were slightly negative for rainy days. Meanwhile, cluster 1 showed a higher frequency with  
 284 respect to D, ranging from 345° to 000°, as well as for clusters 3 and 4, but with some flows coming from  
 285 all directions; while cluster 2 showed a clear predominance for a frequency centred around 300°. Regarding  
 286 F, all the rainy days and clusters 1, 3, and 4 exhibited high frequencies for low F values. Cluster 2 showed  
 287 a different behaviour for extreme precipitation events and all events with high F values. Regarding vorticity,  
 288 for all the rainy days and in all the clusters, Z exhibited a negative behaviour. Cluster 2 showed the most  
 289 frequent Z for null values but reached high positive values as well.

290 At 250 hPa, the reference period exhibited a clear component for D centred at 300°, F values were frequent  
 291 in the range of 75 to 100 m/s, and Z values were frequently in the negative range. Cluster 1 showed a  
 292 predominance for 300°, as well as cluster 3 and cluster 4, both of which centred around 315°. Cluster 2 also  
 293 centred at 300°, and showed more affinity for this direction than the other clusters. Regarding F, all the

294 rainy days showed light strength, and the frequency increased for clusters 1, 3, and 4, all centred at 50°.  
295 Cluster 2 showed very high frequency for F, ranging between 200 and 400 m/s. The vorticity presented  
296 negative values in all cases, but rainy days and clusters 1, 3, and 4 showed some slightly positive cases,  
297 while cluster 2 showed fewer negative cases, but only a few positive and very positive ones.

298

### 299 3.3. Back trajectories analysis

300 At the surface level and for cluster 1 (Fig. 7a), the south-south-west component trajectories predominated,  
301 with those coming from the north-east appearing as the second most predominant component. Thus, a  
302 greater influence of flows from the Pacific than from the Amazon basin is highlighted. For cluster 2 (Fig.  
303 7b), in addition to a fewer number of days of intense precipitation, it is also important to note that most of  
304 these days indicate a pathway from the north, and very few days with trajectories from other quadrants.  
305 Cluster 3 (Fig. 7c) shows a dominance of flows from the south-east, with a small number coming from the  
306 north-east from the Amazon basin and the Bolivian Altiplano. The same is true for cluster 4 (Fig. 7d), with  
307 a majority of trajectories marking a southerly component that eventually turns eastward. In this case, the  
308 trajectories coming from the Bolivian Altiplano are testimonial to their frequency. Summing up, in clusters  
309 1, 3, and 4, the rainiest days have mainly southerly component trajectories at SLP; the opposite direction is  
310 true for cluster 2.

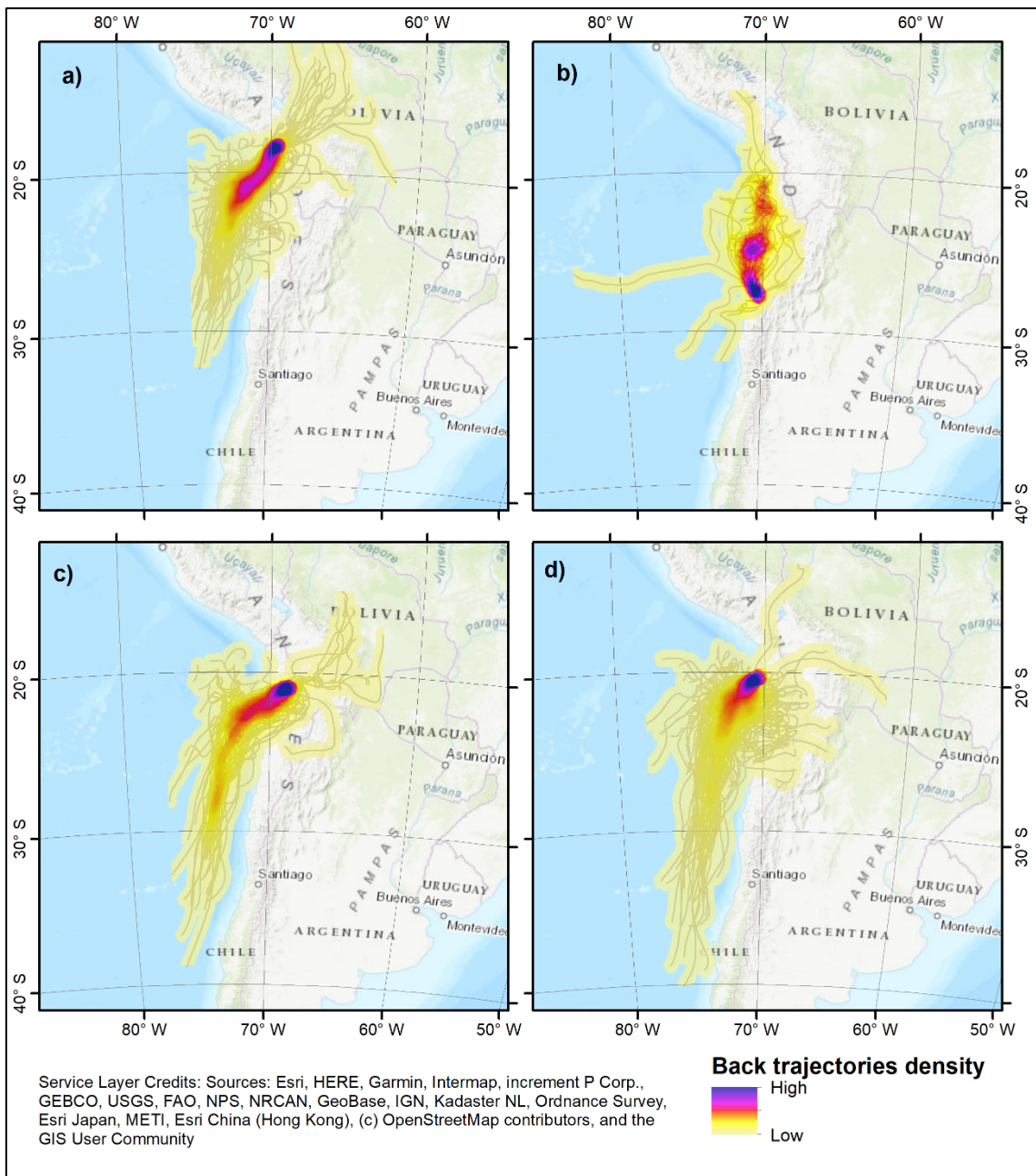
311 At 500 hPa, cluster 1 (Fig. 8a) shows a large number of trajectories coming from the Bolivian Altiplano,  
312 although they also come from the Pacific. At the same time, it is intuitive how the latter are affected by an  
313 anticyclonic turn and would describe a trajectory that intercepts the continental area. It is interesting to note  
314 that the two longest trajectories coming from the Pacific show a cyclonic turn.

315 Cluster 2 (Fig. 8b) shows how in the middle troposphere, intense precipitation occurs with predominately  
316 westerly flows, but with some flows coming from the north, although to a lesser extent. This cluster shows  
317 clear differences from the others when analysing the length of the described trajectories: the air mass that  
318 arrives at 18:00 UTC-4 on a day with extreme precipitation at the reference station in this cluster; 72 h  
319 earlier and far away (perhaps thousands of kilometres), over the Pacific Ocean. Cluster 3 (Fig. 8c) shows  
320 almost all the trajectories originating in the Bolivian Altiplano 72 h prior to extreme precipitation events in  
321 the Atacama Desert. The same is true for Cluster 4 (Fig. 8d), where a concentration of trajectories over the  
322 same area referred to for the previous cluster is evident, and in addition to a significant number of  
323 trajectories originated in the Amazon basin.

324 For the high troposphere (250 hPa) at cluster 1 (Fig. 9a), the back trajectories predominate over the Bolivian  
325 Altiplano, but this time they completely transcend the limits of that area and affect a large portion of the  
326 Amazon basin in northern Argentina, eastern Bolivia, western Paraguay, and Brazil. Trajectories  
327 originating in the Pacific Ocean and southern Peru also appear. For cluster 2 (Fig. 9b), all the back  
328 trajectories originated from the Pacific Ocean and have a westerly component. The case for clusters 3 (Fig.  
329 9c) and 4 (Fig. 9d) is similar to that of cluster 1, with the trajectories predominately over Bolivia, but in  
330 both cases, transcending this boundary and affecting the Amazon basin and Pacific Ocean regions.

331

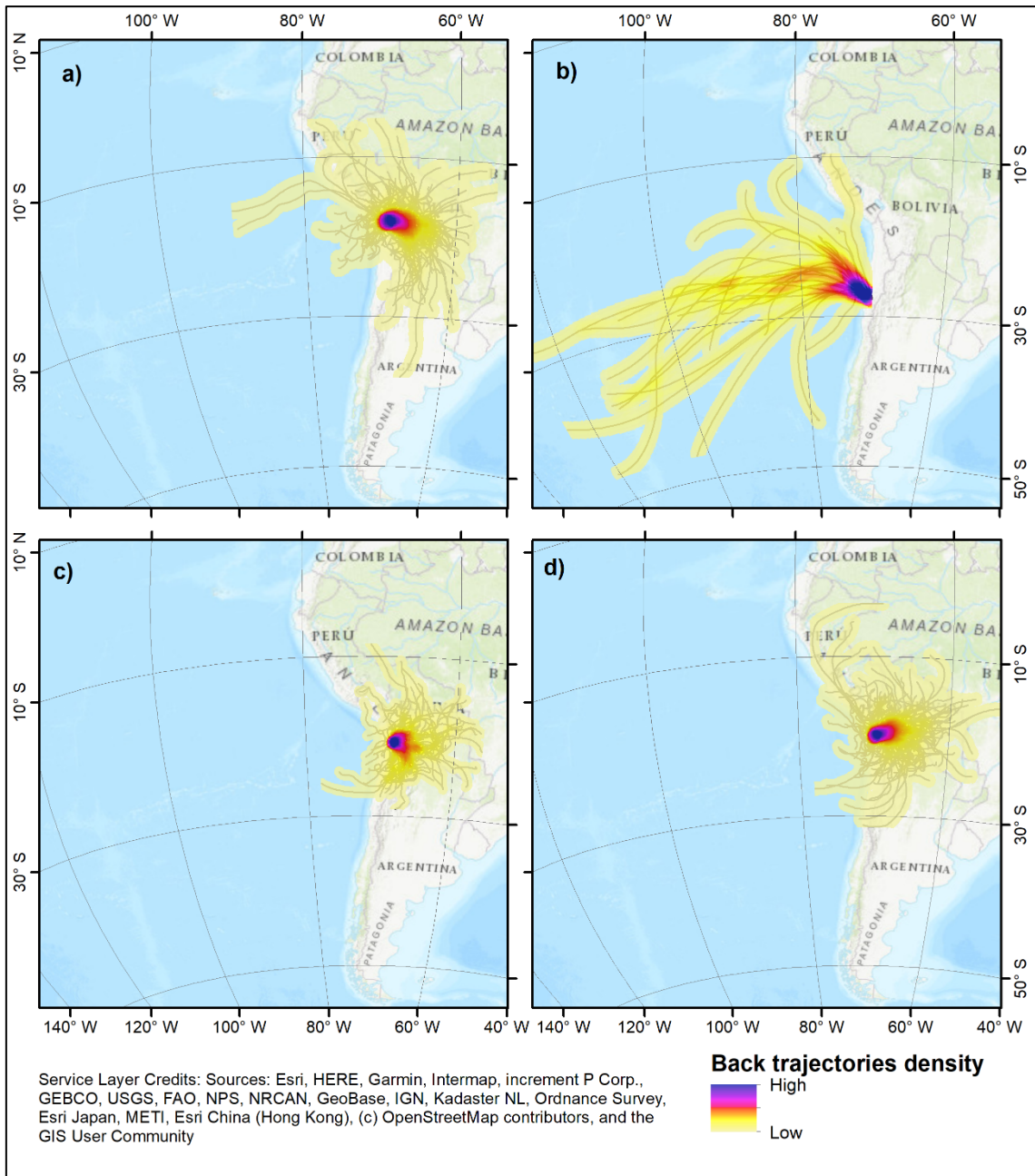




332  
 333  
 334  
 335

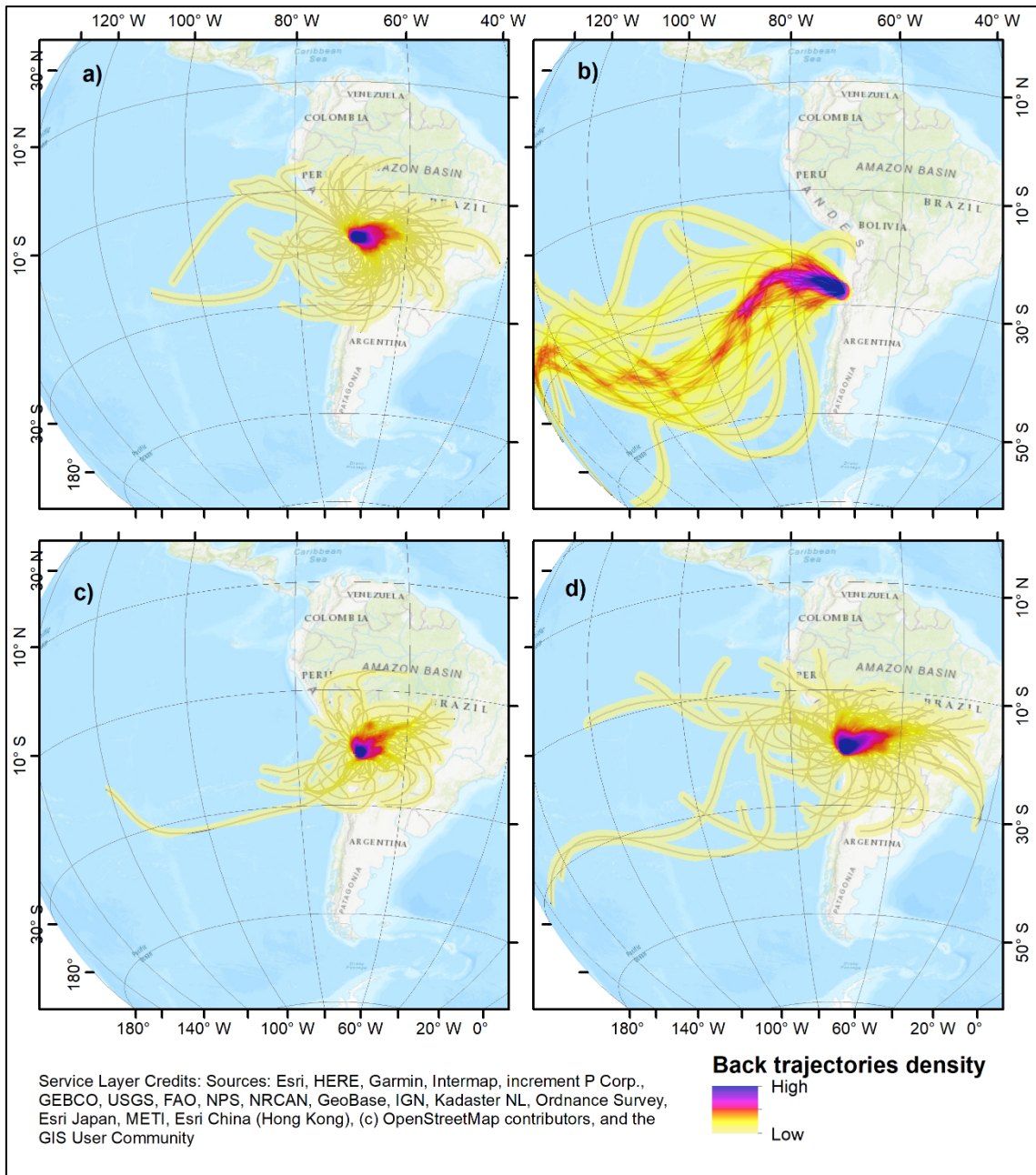
Figure 7. Back trajectories for each cluster for days with torrential rainfall at slp for a) cluster 1, b) cluster 2, c) cluster 3, and d) cluster 4





336  
 337  
 338  
 339

Figure 8. Back trajectories for each cluster for days with torrential rainfall at 500 hPa for a) cluster 1, b) cluster 2, c) cluster 3, and d) cluster 4



340 Figure 9. Back trajectories for each cluster for days with torrential rainfall at 250 hPa for a) cluster 1, b)  
 341 cluster 2, c) cluster 3, and d) cluster 4  
 342  
 343

#### 344 4. Discussion and conclusions

345 Regarding the behaviour of D, F, and Z, it is evident that the southern region of the study area (cluster 2)  
 346 behaved differently from the other clusters. This could be related to the different processes taking place at  
 347 this latitude, but also by the less frequent occurrence of extreme precipitation events registered for this  
 348 region.

349 Vorticity exhibits a quite different behaviour compared to the mean values, so it appears as an important  
 350 factor that modulates intense precipitation at different tropospheric levels (de Vries et al. 2018; Khansalari  
 351 et al. 2020). All the clusters show higher vorticity values in cases of intense precipitation, and agreeing with  
 352 the results presented by Li et al. (2019): more intense vorticity and flow intensity, associated water vapour  
 353 transport, strong upper-level jet stream, and intense South Pacific high pressure, could contribute to intense  
 354 precipitation, which in turn could be related to different flow directions. This is evident at 500 hPa, where  
 355 for all clusters, except cluster 2, where torrential events came from all directions. This explains that the  
 356 flow direction in the 24 h prior to an event is not that determinant on vorticity, but it is evident that the  
 357 Pacific Ocean component plays a key role.

358 The analysis of the back trajectories at the different tropospheric levels identified behavioural similarities  
359 in clusters 1, 3, and 4, with cluster 2 presenting the greatest degree of difference with respect to the others,  
360 mainly due to the northerly component in D that occurs at the surface and the westerly component that  
361 occurs in the middle and upper troposphere. This agrees with the results presented in Baltazar Fernández  
362 and Meseguer-Ruiz (2019), where the daily precipitation concentration was much more reduced in this  
363 southern coastal region, and likely linked to the more frontal origin of precipitation. Meanwhile, for the rest  
364 of the study area, precipitation showed a convective origin. At these latitudes (25°S), it is possible that cold  
365 fronts coming from the Pacific Ocean arrive and cause frontal precipitation to occur.

366 At the surface level, the fact that torrential rainfall caused by flows of oceanic origin is related to the possible  
367 configuration of surface low pressures off the coast of northern Chile that encourage flow of southerly and  
368 then westerly components that provide moisture to the continent (Sarricolea and Romero 2015). Colliding  
369 with the abrupt coastal relief and steep sections of the Andes Mountains, these easterly flows are forced to  
370 ascend rapidly causing precipitation (Junquas et al. 2016, 2018). This phenomenon responds to processes  
371 of a synoptic scale, coinciding with what was indicated in previous studies (Junquas et al. 2016, 2018;  
372 Barlow et al. 2019; Meseguer-Ruiz et al. 2020), or local processes derived from thermal differences  
373 between the continent and the ocean.

374 In the middle and upper troposphere, the trajectories show an evident origin in the Bolivian Altiplano,  
375 which highlights the influence that relief plays in the configuration of the so-called Bolivian High (Segura  
376 et al. 2016, 2018) and the northward extension of the low level jet stream (Segura et al. 2019), introducing  
377 humid air flow from the Amazon basin that interacts directly with the South American monsoon (Sarricolea  
378 and Romero 2015; Meseguer-Ruiz et al. 2019a). First, this shows the importance of this summertime pattern  
379 in the region (Marengo et al. 2012).

380 Nevertheless, changes in this system in recent years and those expected in the future (Jones and Carvalho  
381 2013) suggest a lesser influence on precipitation as noted in Segura et al. (2020). The results described in  
382 Segura et al. (2019) suggest that in the northern part of the study area, extremely rainy seasons are linked  
383 to sea surface temperature anomalies in the central equatorial Pacific, but none of the pathways analysed  
384 here allowed us to confirm this statement.

385 The extreme precipitation events can also be linked to the Madden-Julian Oscillation (MJO) which  
386 propagates anomalies in convection and circulation, with a key role played by the orography in the study  
387 area (Grimm 2019) and agree with the results presented here. In a warmer climate, uncertainties arise  
388 concerning this point because MJO-related precipitation can increase intensity and affect wind in these  
389 regions (Maloney et al. 2019). This phenomenon could be related to changes in the pathways and in the D,  
390 F, and Z at different tropospheric levels that could help to improve predictions of extreme weather in this  
391 region. It is well known that synoptic circulation is better represented by simulation than precipitation. The  
392 mechanisms of tropical-extratropical interactions leading to the occurrence of extreme precipitation are  
393 explained by a vorticity intrusion forcing the incursion of tropical water vapour, in agreement with the  
394 findings of de Vries et al. (2018) for another arid subtropical region.

395 Torrential rainfall in cluster 2 could be associated with the configuration of cut-off low pressure systems in  
396 front of the continent that would have their evidence in the middle troposphere and would provide flows of  
397 oceanic origin, coinciding with what was proposed by Garreaud (2009) and demonstrated in a study by  
398 Meseguer-Ruiz et al. (2020). These results show that extreme precipitation over northern Chile mainly  
399 results from the effects of synoptic scale in the north, and large-scale atmospheric circulation in the southern  
400 coastal region. This also agrees with the results presented by Segura et al. (2020), who stated that easterly  
401 circulation in upper levels no longer correlates with precipitation, and suggests that this region may have  
402 experienced a change in the dominant mode of atmospheric circulation.

403 Different regions across northern Chile show distinct patterns of pathways in timing for 24 h of extreme  
404 precipitation, but it is difficult to make a statement about which processes explain the gain of specific  
405 humidity from the present pathway analysis due to differences in criteria when considering the origin time  
406 of the different pathways as compared to other similar studies (Gustafsson et al. 2010, Bracken et al. 2015),  
407 in which times varied from 15 to 32 h. Thus, consideration of the lag time from the initial conditions that  
408 eventually produced the extreme precipitation event is highly relevant to determine the pathway followed  
409 by the relevant air masses.

#### 410 Acknowledgments

411 The authors want to thank the FONDECYT Project 11160059 of the Chilean Government, the Climatology  
412 Group (2017SGR1362, Catalan Government) and the CLICES Project (CGL2017-83866-C3-2-R) for the  
413 institutional support. R.S.N. is funded by the Spanish Ministry of Science and Innovation (grant no. FJCI-  
414 2017-31595).

#### 415 416 417 References

418 Alexandersson H (1986) A homogeneity test applied to precipitation data. *Int J Climatol* 6:661-675.  
419 <https://doi.org/10.1002/joc.3370060607>

420 Baez-Villanueva OM, Zambrano-Bigiarini M, Ribbe L, Nauditt A, Giraldo-Osorio JD, Thinh NX. (2018)  
421 Temporal and spatial evaluation of satellite rainfall estimates over different regions in Latin-America.  
422 *Atmos Res* 213:34-50. <https://doi.org/10.1016/j.atmosres.2018.05.011>

423 Baltazar Fernández A, Meseguer-Ruiz O (2019) Regionalization of the Concentration Index in northern  
424 Chile and its relation with the orographic component (1966-2015). *Investigaciones Geográficas*, 57:32-48.  
425 <https://doi.org/10.5354/0719-5370.2019.53440>

426 Barlow M, Gutowski Jr WJ, Gyakum JR, Katz RW, Lim YK, Schumacher RS, Wehner MF, Agel L,  
427 Bosilovich M, Collow A, Gershunov A, Grotjahn R, Leung R, Milrad S, Min SK. (2019) North American  
428 extreme precipitation events and related large-scale meteorological patterns: a review of statistical methods,  
429 dynamics, modeling, and trends. *Clim Dyn* 53:6835-6875. <https://doi.org/10.1007/s00382-019-04958-z>

430 Bombi P. (2018) Potential impacts of climate change on *Welwitschia mirabilis* populations in the Namib  
431 Desert, southern Africa. *J Arid Land* 10:663-672. <https://doi.org/10.1007/s40333-018-0067-1>

432 Bracken C, Rajagopalan B, Alexander M, Gangopadhyay S. (2015) Spatial variability of seasonal extreme  
433 precipitation in the western United States. *J Geophys Res Atmos* 120:4522-4533.  
434 <https://doi.org/10.1002/2015JD023205>

435 Casado MJ, Pastor MA. (2016) Circulation types and winter precipitation in Spain. *Int J Climatol* 36:2727-  
436 2742. <https://doi.org/10.1002/joc.3860>

437 Castino F, Bookhagen B, Strecker MR. (2017) Rainfall variability and trends of the past six decades (1950–  
438 2014) in the subtropical NW Argentine Andes. *Clim Dyn* 48: 1049-1067. <https://doi.org/10.1007/s00382-016-3127-2>

439 Chen Y, Zhai P. (2016) Mechanisms for concurrent low-latitude circulation anomalies responsible for  
440 persistent extreme precipitation in the Yangtze River Valley. *Clim Dyn* 47:989-1006.  
441 <https://doi.org/10.1007/s00382-015-2885-6>

442 Chilean Government (2015) Plan de Adaptación al Cambio Climático. Departamento de Cambio Climático  
443 del Ministerio del Medio Ambiente, Santiago.

444 De Vries AJ, Ouwersloot HG, Feldstein SB, Riemer M, El Kenawy AM, McCabe MF, Lelieveld J. (2018)  
445 Identification of Tropical-Extratropical Interactions and Extreme Precipitation Events in the Middle East  
446 Based On Potential Vorticity and Moisture Transport. *J Geophys Res Atmos* 123:861-881.  
447 <https://doi.org/10.1002/2017JD027587>

448 Donat M, Lowry A, Alexander L, O'Gorman P, Maher N. (2016) More extreme precipitation in the world's  
449 dry and wet regions. *Nat Clim Change* 6:508-513. <https://doi.org/10.1038/NCLIMATE2941>

450 Draxler RR, Hess GD. (1998) An overview of the HYSPLIT\_4 modeling system for trajectories, dispersion,  
451 and deposition. *Aust Meteor Mag* 47:295-308. *Geophys Res Atmos* 123:861-881.  
452 <https://doi.org/10.1002/2017JD027587>

453 Falvey M, Garreaud R. (2005) Moisture variability over the South American Altiplano during the South  
454 American low level jet experiment (SALLJEX) observing season. *J Geophys Res* 110:D22105.  
455 <https://doi.org/10.1029/2005JD00615>

456 Fealey R, Mills G. (2018) Deriving Lamb weather types suited to regional climate studies: A case study on  
457 the synoptic origins of precipitation over Ireland. *Int J Climatol* 38:3439-3448.  
458 <https://doi.org/10.1002/joc.5495>

459 Fleming ZL, Monks PS, Manning AJ. (2012) Review: Untangling the influence of air-mass history in  
460 interpreting observed atmospheric composition. *Atmos Res* 104-105:1-39.  
461 <https://doi.org/10.1016/j.atmosres.2011.09.009>

462 Garreaud RD. (2009) The Andes climate and weather. *Adv Geosci* 22:3–11. <https://doi.org/10.5194/adgeo-22-3-2009>

463 Garreaud R, Aceituno P. (2001) Interannual rainfall variability over the South American Altiplano. *J Clim*  
464 14:2779–2789. [https://doi.org/10.1175/1520-0442\(2001\)014<2779:IRVOTS>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<2779:IRVOTS>2.0.CO;2)

465 Green B, Marshall J, Campin JM. (2019) The 'sticky' ITCZ: ocean-moderated ITCZ shifts. *Clim Dyn* 53:1-  
466 19. <https://doi.org/10.1007/s00382-019-04623-5>

467 Grimm AM. (2019) Madden-Julian Oscillation impacts on South American summer monsoon season:  
468 precipitation anomalies, extreme events, teleconnections, and role in the MJO cycle. *Clim Dyn* 53:907-932.  
469 <https://doi.org/10.1007/s00382-019-04622-6>

470 Guijarro JA (2016) Package "climatol" Climate Tools (Series Homogenization and Derived Products).  
471 <https://CRAN.R-project.org/package=climatol>

472 Gustafsson M, Rayner D, Chen D. (2010) Extreme rainfall events in southern Sweden: where does the  
473 moisture come from? *Tellus A* 62(5):605-616. <https://doi.org/10.1111/j.1600-0870.2010.00456.x>

476 Houston J, Hartley AJ. (2003) The central Andean west-slope rainshadow and its potential contribution to  
477 the origin of hyper-aridity in the Atacama Desert. *Int J Climatol* 23:1453–1464.  
478 <https://doi.org/10.1002/joc.938>

479 Insel N, Poulsen CJ, Ehlers TA. (2010) Influence of the Andes Mountains on South American moisture  
480 transport, convection, and precipitation, *Clim Dyn* 35:1477-1492. [https://doi.org/10.1007/s00382-009-](https://doi.org/10.1007/s00382-009-0637-1)  
481 [0637-1](https://doi.org/10.1007/s00382-009-0637-1)

482 Jenkinson AF, Collison P. (1977) An initial climatology of gales over the North Sea. Synoptic Climatology  
483 Branch Memorandum no 62, Bracknell. Meteorological Office, London

484 Jones C, Carvalho LMV. (2013) Climate Change in the South American Monsoon System: Present Climate  
485 and CMIP5 Projections. *J Climate* 26:6660-6678. <https://doi.org/10.1175/JCLI-D-12-00412.1>

486 Junquas C, Li L, Vera CS, Le Treut H, Takahashi K. (2016) Influence of South America orography on  
487 summertime precipitation in Southeastern South America. *Clim Dyn* 46:3941-3963.  
488 <https://doi.org/10.1007/s00382-015-2814-8>

489 Junquas C, Takahashi K, Condom T, Espinoza JC, Chavez S, Sicart JE, Lebel T. (2018) Understanding the  
490 influence of orography on the precipitation diurnal cycle and the associated atmospheric processes in the  
491 central Andes. *Clim Dyn* 50:3995-4017. <https://doi.org/10.1007/s00382-017-3858-8>

492 Kalnay E, Kanamitsu M, Kistler R, Collins W, Deaven D, Gandin L, Iredell M, Saha S, White G, Woollen  
493 J, Zhu Y, Chelliah M, Ebisuzaki W, Higgins W, Janowiak J, Mo KC, Ropelewski C, Wang J, Leetmaa A,  
494 Reynolds R, Jenne R, Joseph D (1996) The NCEP/NCAR 40-year reanalysis project. *Bull Amer Meteor*  
495 *Soc* 77:437-470. [https://doi.org/10.1175/1520-0477\(1996\)077%3C0437:TNYRP%3E2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077%3C0437:TNYRP%3E2.0.CO;2)

496 Khansalari S, Mohebalhojeh AR, Ahmadi-Givi F, Sprenger M. (2020) On the determining factors in cases  
497 of moderate to heavy precipitation in Tehran. *Theor Appl Climatol* 140:1107-1123.  
498 <https://doi.org/10.1007/s00704-020-03133-3>

499 Kirtman B, Power SB, Adedoyin JA, Boer GJ, Bojariu R, Camilloni I, Doblas-Reyes FJ, Fiore AM, Kimoto  
500 M, Meehl GA, Prather M, Sarr A, Schär C, Sutton R, van Oldenborgh GJ, Vecchi G, Wang HJ. (2013)  
501 Near-term climate change: projections and predictability. In Stocker TF, Qin D, Plattner GK, Tignor M,  
502 Allen SK, Boschung J, Nauels A, Xia Y, Bex V, Midgley PM. (Eds.) *Climate Change 2013: The Physical*  
503 *Science Basis*. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental  
504 Panel on Climate Change. Cambridge: Cambridge University Press, 953-1028

505 Kock ST, Schitteck K, Mächtle B, Maldonado A, Vos H, Lupo LC, Kulemeter JJ, Wissel H, Schäbitz F,  
506 Lücke A. (2020) Multi-centennial-scale variations of South American summer monsoon intensity in the  
507 southern central Andes (24-27°S) during the late Holocene. *Geophys Res Lett* 47:e2019GL084157.  
508 <https://doi.org/10.1029/2019GL084157>

509 Li L, Zhang R, Wen M, Duan J, Qi Y. (2019) Characteristics of the Tibetan Plateau vortices and the related  
510 large-scale circulations causing different precipitation intensity. *Theor Appl Climatol* 138:849-860.  
511 <https://doi.org/10.1007/s00704-019-02870-4>

512 Maloney ED, Adames AF, Bui HX. (2019) Madden–Julian oscillation changes under anthropogenic  
513 warming. *Nature Clim Change* 9:26-33. <https://doi.org/10.1038/s41558-018-0331-6>

514 Marengo JA, Liebmann B, Grimm AM, Misra V, Silva Dias PL, Cavalcanti IFA, Carvalho LMV, Berbery  
515 EH, Ambrizzi T, Vera CS, Saulo AC, Nogues-Paegle J, Zipser E, Seth A, Alves LM. (2012) Recent  
516 developments on the South American monsoon system. *Int J Climatol* 32:1-21.  
517 <https://doi.org/10.1002/joc.2254>

518 Marjani S, Alizadeh.Choobari O, Irannejad P. (2019) Frequency of extreme El Niño and La Niña events  
519 under global warming. *Clim Dyn*. <https://doi.org/10.1007/s00382-019-04902-1>

520 Meseguer-Ruiz O, Ponce-Philimon PI, Guijarro JA, Sarricolea P. (2019a) Spatial distribution and trends of  
521 different precipitation variability indices based on daily data in Northern Chile between 1966 and 2015. *Int*  
522 *J Climatol* 39:4595-4610. <https://doi.org/10.1002/joc.6089>

523 Meseguer-Ruiz O, Osborn TJ, Sarricolea P, Jones PD, Olcina J, Serrano-Notivoli R, Martín-Vide J. (2019b)  
524 Definition of a temporal distribution index for high temporal resolution precipitation data over Peninsular  
525 Spain and the Balearic Islands: the fractal dimension; and its synoptic implications. *Clim Dyn* 52:439-456.  
526 <https://doi.org/10.1007/s00382-018-4159-6>

527 Meseguer-Ruiz O, Cortesi N, Guijarro JA, Sarricolea P. (2020) Weather regimes linked to daily  
528 precipitation anomalies in Northern Chile. *Atmos Res* 236: 104802.  
529 <https://doi.org/10.1016/j.atmosres.2019.104802>

530 Miró JR, Pepin N, Peña JC, Martín-Vide J. (2020) Daily atmospheric circulation patterns for Catalonia  
531 (northeast Iberian Peninsula) using a modified version of Jenkinson and Collison method. *Atmos Res*  
532 231:104674. <https://doi.org/10.1016/j.atmosres.2019.104674>

533 Otero N, Sillmann J, Butler T. (2018) Assessment of an extended version of the Jenkinson–Collison  
534 classification on CMIP5 models over Europe. *Clim Dyn* 50: 1559-1579. [https://doi.org/10.1007/s00382-](https://doi.org/10.1007/s00382-017-3705-y)  
535 [017-3705-y](https://doi.org/10.1007/s00382-017-3705-y)



536 Pliscoff P, Luebert F, Hilger HH, Guisan A. (2014) Effects of alternative sets of climatic predictors on  
537 species distribution models and associated estimates of extinction risk: A test with plants in an arid  
538 environment. *Ecol Model* 288:166-177. <https://doi.org/10.1016/j.ecolmodel.2014.06.003>

539 Power SB, Delage F, Colman R, Moise A. (2012) Consensus on 21st century rainfall projections in climate  
540 models more widespread than previously thought. *J Clim* 25:3792-3809. <https://doi.org/10.1175/JCLI-D-11-00354.1>

541 Putniković S, Tošić I, Durđević V. (2016) Circulation weather types and their influence on precipitation in  
542 Serbia. *Meteorol Atmos Phys* 128:649-662. <https://doi.org/10.1007/s00703-016-0432-6>

543 Putniković S, Tošić I. (2018) Relationship between atmospheric circulation weather types and seasonal  
544 precipitation in Serbia. *Meteorol Atmos Phys* 130:393-403. <https://doi.org/10.1007/s00703-017-0524-y>

545 Rohde RF, Hoffman MT, Durbach I, Venter Z, Jack S. (2019) Vegetation and climate change in the Pro-  
546 Namib and Namib Desert based on repeat photography: Insights into climate trends. *J Arid Environ*  
547 165:119-131. <https://doi.org/10.1016/j.jaridenv.2019.01.007>

548 Sarricolea P, Romero H. (2015) Variabilidad y cambios climáticos observados y esperados en el Altiplano  
549 del norte de Chile. *Rev Geogr Norte Gd* 62:169-183. <https://doi.org/10.4067/S0718-34022015000300010>

550 Sarricolea P, Herrera-Ossandon MJ, Meseguer-Ruiz O. (2017a) Climatic regionalisation of continental  
551 Chile. *J Maps* 13(2):66–73. <https://doi.org/10.1080/17445647.2016.1259592>

552 Sarricolea P, Meseguer-Ruiz O, Romero Aravena H. (2017b) Tendencias de la precipitación en el Norte  
553 Grande de Chile y su relación con las proyecciones de cambio climático. *Diálogo Andino, Revista de*  
554 *Historia, Geografía y Cultura Andina* 54:41–50. <https://doi.org/10.4067/S0719-26812017000300041>

555 Schneider T, Bischoff T, Haug GH. (2014) Migrations and dynamics of the intertropical convergence zone.  
556 *Nature* 513:45-53. <https://doi.org/10.1038/nature13636>

557 Segura H, Espinoza JC, Junquas C, Takahashi K. (2016) Evidencing decadal and interdecadal  
558 hydroclimatic variability over the Central Andes. *Env Res Let* 11:094016. <https://doi.org/10.1088/1748-9326/11/9/094016>

559 Segura H, Junquas C, Espinoza JC, Vuille M, Jauregui YR, Rabatel A, Condom T, Lebel T. (2019) New  
560 insights into the rainfall variability in the tropical Andes on seasonal and interannual time scales. *Clim Dyn*  
561 53:405-426. <https://doi.org/10.1007/s00382-018-4590-8>

562 Segura H, Espinoza JC, Junquas C, Lebel T, Vuille M, Garreaud R. (2020) Recent changes in the  
563 precipitation-driving processes over the southern tropical Andes/western Amazon. 54:2613-2631.  
564 <https://doi.org/10.1007/s00382-020-05132-6>

565 Shukla PR, Skea J, Slade R, van Diemen R, Haughey E, Malley J, Pathak M, Portugal Pereira J. (2019)  
566 Technical Summary. In Shukla PR, Skea J, Calvo Buendía E, Masson-Delmotte V, Pörtner HO, Roberts  
567 DC, Zhai P, Slade R, Connors S, van Diemen R, Ferrat M, Haughey E, Luz S, Neogi S, Pathak M, Petzold  
568 J, Portugal Pereira J, Vyas P, Huntley E, Kissick K, Belkacemi M, Malley J. (eds.) *Climate Change and*  
569 *Land: an IPCC special report on climate change, desertification, land degradation, sustainable land*  
570 *management, food security, and greenhouse gas uxes in terrestrial ecosystems*. In press.

571 Stein AF, Draxler RR, Rolph GD, Stunder BJB, Cohen MD (2015) NOAA's HYSPLIT Atmospheric  
572 Transport and Dispersion Modeling System. *B Am Meteorol Soc* 96:2059-2077.  
573 <https://doi.org/10.1175/BAMS-D-14-00110.1>

574 Tan X, Gan TY, Chen YD. (2018) Moisture sources and pathways associated with the spatial variability of  
575 seasonal extreme precipitation over Canada. *Clim Dyn* 50:629-640. <https://doi.org/10.1007/s00382-017-3630-0>

576 Valdés-Pineda R, Valdés JB, Diaz HF, Pizarro-Tapia R. (2015) Analysis of spatio-temporal changes in  
577 annual and seasonal precipitation variability in South America-Chile and related ocean-atmosphere  
578 circulation patterns. *Int J Climatol* 36:2979–3001. <https://doi.org/10.1002/joc.4532>

579 Valdés-Pineda R, Cañón J, Valdés JB. (2018) Multi-decadal 40-to 60-year cycles of precipitation variability  
580 in Chile (South America) and their relationship to the AMO and PDO signals. *J Hydrol* 556:1153–1170.  
581 <https://doi.org/10.1016/j.jhydrol.2017.01.031>

582 Vasconcellos FC, Deng Y, Zhang H, Martins G. (2020) Austral summer precipitation biases over tropical  
583 South America in five CMIP5 earth system models. *Int J Climatol*. <https://doi.org/10.1002/joc.6595>

584 Vincent LA, Zhang X, Bonsal BR, Hogg WD (2002) Homogenization of daily temperatures over Canada.  
585 *J Clim* 15:1322-1334. [https://doi.org/10.1175/1520-0442\(2002\)015<1322:HODTOC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2002)015<1322:HODTOC>2.0.CO;2)

586 Williams CJR. (2017) Climate change in Chile: an analysis of state-of-the-art observations, satellite-derived  
587 estimates and climate model simulations. *Journal of Earth Science & Climatic Change* 8:5.  
588 <https://doi.org/10.4172/2157-7617.1000400>

589 Zappalà DA, Barreiro M, Masoller, C. (2018) Quantifying changes in spatial patterns of surface air  
590 temperature dynamics over several decades. *Earth Syst Dynam* 9:383-391. <https://doi.org/10.5194/esd-9-383-2018>