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1 **Causes of the long-term variability of southwestern**
2 **South America precipitation in the IPSL-CM6A-LR**
3 **model**

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7 **Abstract** Southwestern South America (SWSA) has undergone frequent and per-
8 sistent droughts in recent decades with severe impacts on water resources, and
9 consequently, on socio-economic activities at a sub-continental scale. The local
10 drying trend in this region has been associated with the expansion of the sub-
11 tropical drylands over the last decades. It has been shown that SWSA precipi-
12 tation is linked to large-scale dynamics modulated by internal climate variability
13 and external forcing. This work aims at unravelling the causes of this long-term
14 trend toward dryness in the context of the emerging climate change relying on a
15 large set simulations of the state-of-the-art IPSL-CM6A-LR climate model from
16 the 6th phase of the Coupled Model Intercomparison Project. Our results iden-
17 tify the leading role of dynamical changes induced by external forcings, over the
18 local thermodynamical effects and teleconnections with internal global modes of
19 sea surface temperature. Our findings show that the simulated long-term changes
20 of SWSA precipitation are dominated by externally forced anomalous expansion
21 of the Southern Hemisphere Hadley Cell (HC) and a persistent positive Southern
22 Annular Mode (SAM) trend since the late 1970s. Long-term changes in the HC ex-

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23 tent and the SAM show strong co-linearity. They are attributable to stratospheric
24 ozone depletion in austral spring-summer and increased atmospheric greenhouse
25 gases all year round. Future ssp585 and ssp126 scenarios project a dominant role
26 of anthropogenic forcings on the HC expansion and the subsequent SWSA dry-
27 ing, exceeding the threshold of extreme drought due to internal variability as soon
28 as the 2040s, and suggest that these effects will persist until the end of the 21st
29 century.

30 **Keywords** Subtropical Andes drying trend · Hadley Cell expansion · Decadal
31 variability · External forcing · CMIP6 · Detection and attribution · Future
32 scenarios

33 1 Introduction

34 The southwestern South America (SWSA) region encompasses the Andean Cordillera
35 and adjacent territories from the Pacific coast to the continental arid lowlands in
36 Argentina and south of the dry Altiplano. This region is characterized by a marked
37 precipitation gradient from < 100 mm in the north ($25 - 28^\circ$ S) to well over 2000
38 mm in the south ($40 - 45^\circ$ S). Precipitation primarily occurs during austral winter
39 (June-August, JJA) associated with passing fronts embedded in the mid-latitude
40 Westerly flow and enhanced by the orographic effect of the Andes Mountains
41 (Montecinos and Aceituno, 2003; Garreaud et al., 2013; Viale et al., 2019). Lati-
42 tudinal variations in the southeastern Pacific anticyclone and the subtropical jet
43 stream modulate the seasonality of these fronts, which commonly form at the
44 poleward limit of the Southern Hemisphere (SH) Hadley Cell (HC) (Montecinos
45 and Aceituno, 2003; Barrett and Hameed, 2017). Closely linked to large-scale at-
46 mospheric circulation, SWSA precipitation supports the many glaciers and lakes
47 in the Andes and contributes to the flow of major streams and rivers along the
48 Cordillera. Indeed, most rivers originate in the upper Andes, where precipitation
49 is comparatively higher than in adjacent territories (Masiokas et al., 2019). With
50 some regional differences, SWSA has undergone a striking drying trend since the

51 1980s (e.g., Garreaud et al., 2013, 2017, 2020; Boisier et al., 2016, 2018), with
52 marked glacier retreats and lake-area reductions without precedent over the last
53 millennium (Garreaud et al., 2017; Pabón-Caicedo et al., 2020). A robust drying
54 trend of around -28 mm per decade during the austral winter rainy season has been
55 found in the southern part of SWSA (Boisier et al., 2018). In the northern drier
56 regions, the larger amplitude of the high-frequency variability in austral summer
57 (December-February, DJF) and fall (March-May, MAM) complicates the detection
58 of consistent trends in precipitation. However, a reduction in river streamflows sug-
59 gests that a drying tendency is also taking place in spring (September-October,
60 SON) and summer in the north sector of SWSA (Boisier et al., 2018). Although
61 they display lower drought than observed, model simulations for the historical pe-
62 riod of the 5th phase of the Coupled Model Intercomparison Project (CMIP5) can
63 capture such drying tendency in response to external anthropogenic forcings (Vera
64 and Díaz, 2015; Boisier et al., 2018). Due to the important socio-economic and eco-
65 logical impacts of changes in water resources in SWSA (CR2, 2015; Norero and
66 Bonilla, 1999; Rosegrant et al., 2000; Meza et al., 2012), it is critical to investigate
67 further and quantify the potential impacts of climate change in this region.

68 Low-frequency rainfall changes in SWSA are also associated with internal cli-
69 mate variability. Indeed, during austral winter and spring, SWSA annual moisture
70 conditions are tightly linked to the southeastern Pacific sea surface temperature
71 (SST) variability (Rutllant and Fuenzalida, 1991; Garreaud et al., 2009; Quintana
72 and Aceituno, 2012; Boisier et al., 2016) related to the El Niño Southern Oscil-
73 lation (ENSO), the leading mode of interannual climate variability in the Pacific
74 Ocean (e.g., McPhaden et al., 2006). Long-term ENSO fluctuations imprint a pan-
75 Pacific pattern of coherent SST decadal variability (Newman et al., 2016) through
76 atmospheric teleconnections (e.g., Alexander et al., 2002). This pattern is known
77 as the Interdecadal Pacific Oscillation (IPO) and is the leading mode of inter-
78 nal, decadal to multidecadal variability in the Pacific Ocean (Folland et al., 1999;
79 Meehl and Hu, 2006). It is also typically referred to as its manifestation in the

wintertime North Pacific SST, the so-called Pacific Decadal Oscillation (Mantua et al., 1997). During the positive (negative) phase, the IPO is characterized by an ENSO-like pattern of warm (cold) SST anomalies across the tropical Pacific, which extends in the subtropics over the eastern boundaries of the Pacific Ocean (Trenberth and Hurrell, 1994; Meehl et al., 2009). The IPO has experienced a trend from a positive (i.e., El Niño-like) to a negative (La Niña-like) phase over the 1980-2014 period associated with an anomalous southward shift and spin-up of the southeastern Pacific anticyclone (Jebri et al., 2020) and of the mid-latitude storm-tracks over SWSA during the rainy season (Quintana and Aceituno, 2012; Boisier et al., 2016, 2018). Thus, the concurrent IPO shift from positive to negative phase might have contributed to the current prevailing SWSA dry conditions (Masiokas et al., 2010; Quintana and Aceituno, 2012; Boisier et al., 2016). However, during DJF and MAM seasons, rainfall interannual-to-decadal fluctuations are also significantly influenced by the dominant mode of atmospheric circulation variability in the mid-latitudes of the SH, the Southern Annual Mode (SAM), also known as the Antarctic Oscillation (Gong and Wang, 1999; Thompson and Wallace, 2000; Thompson et al., 2000). During SAM positive phases, a zonally symmetric atmospheric pressure gradient is observed with negative and positive anomalies over Antarctica and the mid-latitudes, respectively, favoring an anomalous poleward shift of the circumpolar westerlies and more constrained zonal flow in mid-latitudes (Thompson and Wallace, 2000; Thompson et al., 2000), resulting in less frontal rainfall in SWSA (Garreaud et al., 2009). These symmetric features and the impact on SWSA precipitation are reversed during the negative SAM phase.

The SWSA drying trends over recent decades could also result from external forcings. For instance, an intensification of the global hydrological cycle is projected under global warming due to a more extensive water vapour loading of the atmosphere, resulting in enhanced E-P (evaporation minus precipitation) in evaporative regions, and reduced E-P in precipitative regions according to the "wet-get-wetter,

109 dry-get-drier” paradigm (Helpd and Soden, 2006; Seager et al., 2010). Several ob-
110 servational datasets, CMIP5 models and reanalyses also suggest an essential role
111 of externally forced large-scale dynamical changes. Observed southward shifts of
112 the subtropical drying regions and mid-latitudes baroclinic eddies during late aus-
113 tral spring and summer have been associated with the SAM positive trend and
114 the expansion of the HC in recent decades (e.g., Gillett and Thompson, 2003; Pre-
115 vidi and Liepert, 2007; Quintana and Aceituno, 2012; Boisier et al., 2018). The
116 HC expansion and SAM positive trends have both been attributed to increasing
117 atmospheric greenhouse gases (GHGs) concentration and stratospheric ozone de-
118 pletion (e.g., Polvani et al., 2011; Kim et al., 2017; Jebri et al., 2020). However, the
119 tendency toward more La Niña events in relation to the recent IPO trend may also
120 contribute to the HC poleward shift (Nguyen et al., 2013; Allen and Kovilakam,
121 2017) and to more frequent positive SAM phases (Carvalho et al., 2005; Fogt et al.,
122 2012). The counteracting effect of the recent ozone recovery (Eyring et al., 2010)
123 and the intensification of GHGs effect (Andreae et al., 2005) is a source of uncer-
124 tainty for the understanding of the future state of these modes (Fogt and Marshall,
125 2020).

126 To our knowledge, previous studies have not explicitly examined (i) the re-
127 spective contribution of anthropogenic forcing and internal climate variability on
128 the decadal-to-longer term variance of SWSA rainfall throughout the last century
129 and a half and (ii) the specific role of direct thermodynamical and dynamical
130 changes in SWSA related to the HC expansion and SAM trend. These are two
131 major points that are addressed in this study along with (iii) the attribution of
132 the modulation of SWSA precipitation to specific sources of external forcing over
133 the last decades (iv) and its projected changes for the 21st century. Here, we as-
134 sess the SWSA hydroclimate changes over the last century using observations,
135 reanalyses and sets of 20 to 32 member-ensemble simulations conducted with the
136 stand-alone LMDz6A-LR atmospheric component (Hourdin et al., 2020) and the
137 IPSL-CM6A-LR coupled model (Boucher et al., 2020) as part of the 6th phase

138 of the CMIP exercise (CMIP6; Eyring et al., 2016). The paper is organized as
139 follows: in the next two sections, the data and methods used are introduced, the
140 results obtained are presented in section 4 and discussed in section 5. A summary
141 and the main conclusions are provided in section 6.

142 **2 Data**

143 2.1 The atmospheric and coupled model

144 We used the CMIP6 version of the Institut Pierre- Simon Laplace (IPSL) stand-
145 alone atmosphere model, called LMDz6A-LR (Hourdin et al., 2020) and the cou-
146 pled atmosphere-ocean general circulation model (GCM), called IPSL-CM6A-LR
147 (Boucher et al., 2020). LMDz6A-LR is coupled to the ORCHIDEE (d’Orgeval
148 et al., 2008) land surface component, version 2.0. In IPSL-CM6A-LR, the LMDz6A-
149 LR is coupled to the oceanic component Nucleus for European Models of the
150 Ocean (NEMO), version 3.6, which includes other models to represent sea-ice
151 interactions (NEMO-LIM3; Rousset et al., 2015) and biogeochemistry processes
152 (NEMO-PISCES; Aumont et al., 2015). LR stands for low resolution, as the at-
153 mospheric grid resolution is 1.25° in latitude, 2.5° in longitude and 79 vertical
154 levels (Hourdin et al., 2020). Compared to the 5A-LR model version and other
155 CMIP5-class models, IPSL-CM6A-LR was significantly improved in terms of the
156 climatology, e.g., by reducing overall SST biases and improving the latitudinal
157 position of subtropical jets in the SH (Boucher et al., 2020). The IPSL-CM6A-LR
158 is also more sensitive to CO_2 forcing increase (Boucher et al., 2020) and repre-
159 sents a more robust global temperature response than the previous CMIP5 version
160 consistently with current state-of-the-art CMIP6 models (Zelinka et al., 2020).

161 2.2 Experimental protocol

162 This study is based on a set of climate simulations generated as part of CMIP6
163 (Table 1). We relied on the pre-industrial control (piControl) coupled run with

164 the external radiative forcing fixed to pre-industrial values for a measure of the
165 internal climate variability generated by the IPSL-CM6A-LR model.

166 We also used the 32-member ensemble of simulations for the historical period
167 (1850-2014), which is branched on random initial conditions from the piControl
168 run to ensure that ensemble members are largely uncorrelated during the simu-
169 lation. All 32 members of this ensemble (referred to as the historical ensemble)
170 use the historical natural and anthropogenic forcings following CMIP6 protocol
171 (Eyring et al., 2016). They include concentrations of GHGs from 1850 to 2014
172 provided by (Meinshausen et al., 2017) while the standard CMIP6 tropospheric
173 and stratospheric ozone concentration were obtained from the Chemistry-Climate
174 Model Initiative (Checa-Garcia et al., 2018). Tropospheric aerosols, from natu-
175 ral and anthropogenic sources (Hoesly et al., 2018; van Marle et al., 2017) are
176 also included along with historical stratospheric natural forcings, corresponding
177 to spectral solar irradiance-stratospheric ozone cycles (Matthes et al., 2017) and
178 the main volcanic eruptions prescribed with the aerosol optical depth (Thomason
179 et al., 2018).

180 Detection and attribution 10-member ensembles (Gillett et al., 2016) are also
181 utilized to understand the role of the different external forcing components in the
182 context of climate change. These experiments are equivalent to the historical ones
183 but include each forcing individually namely either GHGs (hist-GHG), strato-
184 spheric ozone depletion (hist-stratO3), aerosols (hist-aer) and natural (hist-nat)
185 while maintaining other forcing at their 1850 level.

186 Two 6-member ensembles for the 21st century (2015-2100) future projections
187 scenarios branched on randomly selected historical members at the year 2014 are
188 also analyzed (O'Neill et al., 2016): the ssp585 equivalent to $\sim 8.5 \text{ W m}^{-2}$ in-
189 creased radiative forcing in the year 2100 due to GHG emissions (the highest
190 future pathway across all CMIP6 scenarios) and the ssp126 mitigation scenario,
191 which assumes the lowest of plausible radiative forcing effects (2.6 W m^{-2}) in 2100
192 assuming substantial mitigation against global warming.

193 Finally, we also use AMIP historical (amip-hist) simulations, which use the
194 same external forcings since 1870 as the historical experiment but with imposed ob-
195 served monthly SST on the LMDz6A-LR atmospheric component. Further details
196 about external forcings implementation strategies in the experiments mentioned
197 above are given in Lurton et al. (2020).

198 2.3 Observational data

199 In addition to model outputs, different observational products are used: in situ
200 and gridded observations data sets and reanalyses (Table 2). We analyze in situ
201 monthly precipitation data from 1960 to 2017 from rain gauges in 129 stations
202 located along the Andes, between 70° - 73° W and 20.5° - 46.5° S provided
203 by the Chilean Center for Climate and Resilience Research, Climate Explorer
204 (<http://explorador.cr2.cl>) (Fig. 1). Note that the station coverage is relatively
205 sparse outside the 27° - 43° S region in SWSA. We also rely on gridded products
206 of monthly precipitation provided by the Global Precipitation Climatology Centre
207 (version 2018; GPCCv2018) and the University of Delaware Air Temperature and
208 Precipitation (version 5.01; UDEL5.01), each of which using different interpola-
209 tion methods. Note in Fig. 1 that the data coverage of the observations used for
210 the reconstruction of these gridded products is consistent with that of the *in situ*
211 observations, which are quite sparse outside SWSA in southern South America.

212 We use the Hadley Centre Sea Ice and Sea Surface Temperature version 1
213 (HadISST) gridded reconstruction of SST observations, which are used as bound-
214 ary conditions in the LMDz6A-LR model amip-hist CMIP6 simulations. To study
215 the atmospheric dynamics, we also analyze the meridional wind and the sur-
216 face pressure variables from several reanalyses: the NOAA-CIRES 20th Century
217 Reanalysis versions 2 (20CRv2), 2c (20CRv2c) and 3 (20CRv3), the ERA-20C,
218 ERA40 and ERA-Interim reanalyses of the European Centre for Medium-Range
219 Weather Forecasts and the National Center for Environmental Prediction (NCEP)

220 and the National Center for Atmospheric Research (NCAR) reanalysis version 1
221 (NCEP1) and 2 (NCEP2).

222 **3 Methods**

223 3.1 Trend analyses and statistical testing

224 All anomalies are calculated by removing the monthly mean seasonal cycle of the
225 entire covered period (Table 1) from the time series before calculating seasonal av-
226 erages (MAM, JJA, SON and DJF). To focus only on the variability from decadal,
227 interdecadal or longer time scales, time series are low-pass filtered (LPF) using a
228 Butterworth filter with a cut-off period of 8, 13 or 40 years, respectively (But-
229 terworth, 1930). The linear trend for each month over 1979-2014 is obtained by
230 applying least squares on unfiltered yearly time series of monthly mean precipi-
231 tation and represented in terms relative to the climatological mean ($\% \text{ decade}^{-1}$).
232 In this case, trend values are statistically tested with a Student t-test with the
233 number of degrees of freedom corresponding to the total number of years minus
234 one. To evaluate the ensemble-mean trend representativeness, we also indicate the
235 regions where at least 80% of the members display trends of the same sign than
236 the ensemble mean. The consistency between the anomalies of the climate param-
237 eters and the modes of variability was estimated using the regression coefficient
238 (α) between these variables. When the time series are spatially distributed the re-
239 sulting regression coefficients are presented as regression patterns. In this case, the
240 ensemble-mean regression coefficients are statistically tested using a random-phase
241 test, based on Ebisuzaki (1997), adapted to the regression (details in Villamayor
242 et al. (2018)). In turn, the level of uncertainty among individual members is indi-
243 cated by showing the grid points, where at least 80% of them display a regression
244 coefficient of the same sign as the ensemble mean. When considering regionally
245 averaged time series for the ensemble mean, the uncertainty of the resulting re-

gression coefficient is quantified by representing the range of values obtained with all members individually.

3.2 Climate indices

We consider five main sets of climate indices that are introduced below. Note that for ensemble simulations, the ensemble-mean index refers to the indices calculated from outputs previously averaged across all members. The 95% confidence interval among individual members, according to a Student t-test, is shown to quantify the uncertainty of the ensemble-mean indices .

3.2.1 SWSA precipitation indices

Empirical Orthogonal Function analysis of the 8-year LPF rainfall anomalies in southern South America (25° - 58° S, 65° - 80° W) is performed to isolate the leading mode of variability at decadal-to-longer time scales. The first Principal Component (PC1) for each season is standardized to get a measure of the variance. Regions with the highest variance are then used to build regional indices by averaging values over: 34° - 49° S, 71° - 77° W in MAM; 28° - 40° S, 70° - 76° W in JJA; 30° - 43° S, 70° - 77° W in SON and 39° - 50° S, 72° - 78° W in DJF.

3.2.2 Global Warming and IPO sea surface temperature indices

A global warming index (GW) is calculated by spatially averaging 40-year LPF annual-mean SST anomalies (SSTAs) over 45° S - 60° N. Residual SSTAs are then derived by regressing out the GW SSTAs pattern. To analyze the IPO influence on SWSA rainfall, an IPO index is computed using the residual SSTA following the tripole approach of Henley et al. (2015), defined as the central equatorial Pacific SST (10° S - 10° N, 170° E - 90° W) minus the average of northwest (25° - 45° N, 140° E - 145° W) and southwest (15° - 50° S, 150° E - 160° W) Pacific SST. The resulting IPO index is then LPF with a 13-year cut-off period.

271 3.2.3 Southern Annular Mode and Hadley Cell extent indices

272 Seasonal SAM indices are computed for each season as the standardized PC1 of
 273 the 8-year LPF sea level pressure (SLP) anomalies in the SH, south of 20° S. In
 274 order to describe the variability of the latitudinal position of the poleward edge of
 275 the HC in the SH, we define an index of the HC extent (HCE) for each season as
 276 the linearly interpolated latitude for which the 500-hPa meridional streamfunction
 277 (ψ_{500}) is equal to zero between 20° and 40° S.

278 3.3 Decomposition of SWSA precipitation variance

279 A decomposition of the variance of SWSA precipitation into the components ex-
 280 plained by the IPO and the GW SST indices is performed based on a multilinear
 281 regression analysis (Mohino et al., 2016). According to this, the total variance of
 282 SWSA precipitation ($var[PR]$) can be expressed in terms of the regression coeffi-
 283 cients from the multilinear fitting that correspond to each index (α_{IPO} and α_{GW} ,
 284 respectively) and a residual ($var[\varepsilon]$) as follows:

$$var[PR] = \alpha_{IPO}^2 + \alpha_{GW}^2 + var[\varepsilon] \quad (1)$$

285 The residual term stands for the variance of the residual of the multilinear fitting,
 286 namely the variance that cannot be explained by the IPO and the GW indices.
 287 These three components are expressed as percentage of the total variance of SWSA
 288 precipitation.

289 3.4 Total moisture budget decomposition

290 A decomposition of the change in the net moisture budget, expressed as precipi-
 291 tation minus evaporation (P-E), into purely thermodynamic and dynamic compo-
 292 nents is performed in this work. According to the moisture budget equation, P-E
 293 equals the divergence of the moisture flux (Brubaker et al., 1993). The moisture

294 flux is the mass-weighted and vertical integral of the product between the specific
 295 humidity (q) and the horizontal winds (\mathbf{u}). Therefore, a change in P-E ($\delta(P - E)$)
 296 implies changes in q (i.e., thermodynamic changes (δTH) and \mathbf{u} (i.e., dynamic
 297 changes (δDC)). Considering such a change as an anomaly with respect to a ref-
 298 erence period (denoted with subscript r), the thermodynamic and the dynamic
 299 components can be separated using the following approach (Seager et al., 2010):

$$\delta(P - E) = \delta TH + \delta DC + RES \quad (2)$$

300 Following Ting et al. (2018), we define δTH and δDC as follows:

$$\delta TH \approx -\frac{1}{g\rho_w} \cdot \nabla \int_0^{p_s} \mathbf{u}_r \cdot \delta q \, dp \quad (3)$$

301

$$\delta DC \approx -\frac{1}{g\rho_w} \cdot \nabla \int_0^{p_s} \delta \mathbf{u}_r \cdot q \, dp \quad (4)$$

302 where g is the gravitational acceleration, ρ_w the density of water, p pressure levels
 303 and p_s the surface pressure.

304 The residual component (RES) mostly accounts for the moisture convergence
 305 changes due to transient eddies, but also includes nonlinear terms that are typically
 306 neglected in the approach of δTH and δDC (Seager et al., 2010; Ting et al., 2018).

307 3.5 Probability Density Functions of precipitation linear trend

308 To further evaluate the role of the IPO phase shift (such as during the 1979-2014
 309 period) on SWSA precipitation trends in our model as compared to observations,
 310 we performed Probability Density Functions (PDFs) of the linear trend of SWSA
 311 precipitation for 36-year periods separately for ensemble members with a positive
 312 or negative IPO trend that is significant at the 5% level during the period (re-
 313 sulting in 25 members of each category). The two resulting PDFs are statistically
 314 compared by testing the null hypothesis that both present independent normal

315 distributions with equal means and equal but unknown variances at the 20% sig-
316 nificance level, according to a t-test. A t-test is also used to evaluate whether mean
317 trend values are significantly different from zero.

318 **4 Results**

319 4.1 Model validation

320 In this section, we present a comparison of the amip-hist and historical simulations
321 with observational products to evaluate the ability of, respectively, the atmospheric
322 component and the coupled model to simulate main aspects of precipitation in
323 SWSA, such as climatology and tendency. We also check simulated IPO phase
324 shifts and the emerging HC expansion, which are addressed in relation to the
325 recent drying trend in SWSA.

326 *4.1.1 SWSA precipitation*

327 The SWSA region rainfall annual cycle over the period common to observations
328 and simulations (1960-2014) is represented in Fig. 2 (left panel). Considering the
329 spatial coverage of observations in SWSA (Fig. 1), the model validation is re-
330 stricted to a latitudinal band between 20° - 47° S. This band is roughly 5° longi-
331 tude width (i.e., two grid points in the model) and centered in 73.75° W. Most of
332 the rain gauge stations are located in the west Chilean territories adjacent to the
333 Andes where precipitation is enhanced by the orographic blocking effect on the
334 westerly atmospheric flow (Falvey and Garreaud, 2007; Smith and Evans, 2007;
335 Viale et al., 2019). The rain gauge observed annual cycle shows a latitudinal migra-
336 tion of rainfall sustaining marked dry and wet seasons in austral summer (DJF)
337 and winter (JJA), respectively (Fig. 2a). The onset of the rainy season occurs
338 gradually during fall (MAM), with maximum rainfalls registered around 37° S in
339 June. This regional maximum is not well captured by the gridded observations,
340 most likely because of interpolation procedures applied to rain gauge data (Figs.

341 2c and 2e). The rainy season demise occurs in spring (SON), before minimum
342 values are reached in summer.

343 The amip-hist and historical simulations reproduce the observed mean annual
344 cycle, with the onset in MAM, the rainy season in JJA with a maximum in June
345 around 37° S, the demise in SON and the dry season in DJF (Figs. 2g and 2i).
346 The model seems to overestimate climatological values north of 35° S and south
347 of 40° S. This discrepancy may instead be attributable to the lack of observations
348 at these latitudes. There is also a high level of uncertainties in observations re-
349 garding the climatological amount. Annual rainfalls reach around 1460 mm in rain
350 gauge observations where it is most rainy between 35° S and 40° S, while gridded
351 products values are about two thirds of this amount (i.e., 910 mm in GPCCv2018
352 and 890 mm in UDEL5.01). Model ensemble-means for amip-hist and historical
353 simulations amount 1670 and 1450 mm respectively, with an ensemble standard
354 deviation of around 25 mm in both cases, which is close to in situ observations.
355 However, historical simulations generally give smaller values than amip-hist, espe-
356 cially during the rainy season. This difference is independent of the size of both
357 ensembles (not shown) and most likely related to the influence of the observed
358 SSTs in amip-hist runs.

359 The annual cycle of the linear trend relative to the climatological mean over
360 the last 36 years (1979-2014) is represented in Fig. 2 (right panel). A significant
361 drying around 33° - 40° S in April-May and around 29° - 36° S in July is observed
362 consistently in all observations datasets (Figs. 2b,d,f). These drying trends indicate
363 a delay in the onset of the rainy season and less rainfall during the rainy season.
364 Although weak and not significant, negative trend values in September, together
365 with the features described before, suggest that, in general, there is a tendency
366 towards shorter and less effective rainy season in SWSA over the last decades.
367 The Hövmoller diagrams from gridded observations show positive trend during
368 late spring and summer south of 40° S, suggesting that the southernmost part of

369 SWSA has become wetter (Figs. 2d and 2f). However, this result is poorly reliable
370 due to the lack of observations in this part of the Andes (Fig. 1).

371 Both amip-hist and historical simulations indicate a widespread drying rela-
372 tive to the simulated climatology along the year, with a maximum from around
373 $\sim 40^\circ$ S in DJFM to $\sim 27^\circ$ S in MJJAS (Figs. 2h and 2j), suggesting a shorten-
374 ing and weakening of the rainy season. The drying trend is in general stronger
375 in amip-hist simulations than in historical runs (independently of the ensemble
376 size, not shown). All amip-hist members include the same observed SST, while
377 in historical simulation, SST variability is largely uncorrelated among ensemble
378 members. Averaging the historical ensemble allows therefore damping the influ-
379 ence of SST on the simulated rainfall trends with respect to the role of external
380 forcing. Differences with amip-hist on the other hand, emphasize the contribution
381 of observed SST variability. The drying trend during the onset and demise of the
382 rainy season is more intense and significant in the amip-hist ensemble mean than
383 in the historical one. This suggests a dominant role of observed SST variability
384 (such as the shift toward a negative IPO) in the rainy season shortening during
385 1979-2014, probably amplified by external forcing influence.

386 To summarize, despite the constraints of comparing local observations with
387 gridded data from GCMs due to the complex orography of the region, the model
388 can reproduce the main observed rainfall climatological features and recent trends.
389 Differences between the amip-hist and the historical coupled simulations suggest
390 a combination of internal and external factors in driving the SWSA drying trend.
391 Previous generation CMIP5 models show an overall underestimation of the SWSA
392 precipitation response to external forcing (Vera and Díaz, 2015; Boisier et al.,
393 2018) that is coherent with the IPSL-CM6A-LR simulations. On the other hand,
394 a preliminary CMIP6 multi-model study places the IPSL-CM6A-LR among the
395 GCMs that best represent the atmospheric changes over southeast Pacific that
396 modulate SWSA rainfall during the last century (Rivera and Arnould, 2020), which

397 supports the use of this model to study the long-term precipitation variability in
398 this region.

399 *4.1.2 IPO phase shifts*

400 We also evaluate the model ability to simulate IPO phase shifts comparable to the
401 observed shift over the 1979-2014 period addressed by Boisier et al. (2016) in terms
402 of the amplitude of the associated SST anomalies. To this aim, Fig. 3 represents
403 IPO indices from observational data (HadISST1) over 1870-2014 and from the
404 piControl run over a representative period of equal length, as well as the linear
405 trend values obtained along both IPO indices in centered running windows of 15-40
406 years long. The trend graphics show blue and red colored plumes, corresponding
407 to negative and positive trend values, respectively. The plumes resulting from the
408 piControl IPO index are, in general, narrower than those from observations. This
409 reveals that the model underestimates the observed persistence of the IPO phases.
410 However, the model succeeds in simulating 36-year IPO phase shifts of up to -0.2
411 °C per decade comparable to the one observed over 1979-2014.

412 *4.1.3 Hadley Cell expansion*

413 To evaluate the model's ability to reproduce the HC expansion in recent decades,
414 we compare the linear trend of the seasonal HCE indices calculated with eight
415 different reanalyses and the amip-hist and historical simulations for their common
416 23-year period (1979-2001) and for a 40-year period (1971-2010), common to the
417 simulations and five reanalyses (Fig. 4). The trend values obtained over the short
418 period show large dispersion between the different members of the amip-hist and
419 historical simulations, compared with those of the longer period. This shows the
420 dependence of the tendency of the HCE index on short-term stochastic internal
421 variability. In the longer term, the trend is more robust across model members
422 suggesting an influence of external forcings. Comparing across seasons, the simu-
423 lated HCE trend is better constrained among members in JJA and most uncertain

424 in DJF, when the HC is most variant and presents wider expansion. Regarding
425 observations, the trend values obtained from reanalyses show large spread as well.
426 Despite the large uncertainty, when the 40-year long period is considered, there is
427 solid agreement among reanalyses regarding the Southern HC expansion towards
428 the pole in recent decades (Grise et al., 2019).

429 The model simulates trends that are within the range of those in the reanal-
430 yses. The trend of the simulated HCE in the ensemble mean is negative in all
431 seasons and in both amip-hist and historical simulations, being widest in DJF. In
432 MAM, the poleward expansion of the HCE is the weakest and close to zero when
433 simulated by the historical simulation, while the amip-hist simulation shows a
434 broader expansion. In contrast, during the other seasons, the simulated ensemble-
435 mean expansion is similar in both simulations, being wider in the historical over
436 the 40-year-period. These results suggest that the external forcing mostly induces
437 the HC poleward expansion in JJA, SON and particularly in DJF, but in MAM
438 this effect is weak and, presumably, the influence of SST internal variability is also
439 relevant.

440 4.2 Relative roles of forced versus internal variability on SWSA precipitation 441 decadal variability and trend

442 In this subsection, we focus on the amip-hist ensemble to unravel the contribution
443 of observed SST to SWSA precipitation low-frequency variability. The first PCs
444 of the 8-year LPF seasonal precipitation anomalies account for much of the total
445 decadal-scale rainfall variability in southern South America (25° - 58° S; 65° - 80°
446 W). The explained total variance of the PC1 is 59.1% in MAM, 75.7% in JJA,
447 67.0% in SON and 69.9% in DJF. These indices together with their respective
448 regression patterns show that most of the precipitation variability in southern
449 South America is concentrated in a dipole of opposite anomalies between the
450 middle (25° - 45° S) and high ($> 50^{\circ}$ S) latitudes in SWSA (Fig. 5). The seasonal
451 patterns in DJF and MAM also present significant anomalies in subtropical South

452 America east of the Andes showing an opposite sign to those recorded in middle
453 SWSA latitudes.

454 These regression patterns allow identifying where the most substantial low-
455 frequency variability occurs in the amip-hist simulations (boxed areas in Fig. 5).
456 These regions vary depending on the season, showing a north-to-south shift from
457 winter (JJA) to summer (DJF), respectively, and are co-located with the centers
458 of maximum rainfall linear trends (gray contours in Fig. 5). These regions with
459 the most substantial negative rainfall anomalies are associated with PC1s positive
460 long-term trend (Fig. 5, left panels), which accounts for nearly the total tendency
461 of the area-averaged precipitation anomalies for each season (Table 3). From 1970
462 to 2014 a deficit of 6.2 mm per decade and around 15.2 mm per decade is simulated
463 for the rainy (JJA) and dry summer (DJF) seasons respectively.

464 Previous studies have shown evidence for the influence of SST internal vari-
465 ability and external forcings on SWSA rainfall recent changes (e.g., Boisier et al.,
466 2016, 2018). In line with these previous works and to identify potential connections
467 between simulated low-frequency variations of SWSA precipitation and observed
468 internal modes of climate variability, we regress PC1s on anomalies of SST, SLP
469 and wind (Fig. 6) before and after removing the long-term trend signal in PC1s
470 with a 3rd-degree polynomial fit (dotted lines in Fig. 5). Such detrending will allow
471 emphasizing typical SST and teleconnection patterns related to internal variability
472 modes.

473 Without detrending, SST regression patterns show warm anomalies almost
474 globally distributed except in the tropical Pacific where an IPO-like relative cool-
475 ing dominates all year round (left panel in Fig. 6). The warm pattern is most
476 widespread in DJF, with strong anomalies over western Pacific, the Indian and
477 Atlantic Oceans, especially in the southern basin, and weakest in JJA. In turn,
478 the tropical Pacific cooling is more intense in SON and JJA and almost negligible
479 in DJF and MAM. This global SST pattern is reminiscent of an IPO negative
480 phase combined with the global warming signal induced by external forcing. In

481 DJF and MAM surface winds and SLP anomalies show a poleward shift of the
482 westerlies and a SLP maximum slightly south of 30° S associated with the SWSA
483 drying. This suggests a link with the strengthening of the SAM with a weaker
484 influence of the negative IPO pattern on the SWSA drying in austral summer and
485 fall, in agreement with Boisier et al. (2018).

486 Regression patterns obtained with the detrended PC1s highlight a sea-level
487 pressure high and significant anomalous easterlies in Southeastern Pacific in JJA
488 and SON, suggesting an atmospheric teleconnection with a negative IPO-like SSTA
489 pattern most prominent in austral winter and spring (right panel in Fig. 6). Ex-
490 tratropical SLP anomalies are less zonally symmetric than with the non-detrended
491 PC1s with an anomalous jet of westerlies passing through SWSA south of around
492 45° S, which is coherent with the anomalous SLP gradient. Therefore, this pattern
493 may suggest that, in response to a negative IPO, there is a poleward shift of the
494 storm-tracks embedded in the zonal flow (Garreaud et al., 2013), resulting in less
495 intrusion of humid air masses over middle latitudes in SWSA. Comparison of re-
496 gression patterns before and after PC1s detrending indicates therefore that precip-
497 itation trends in winter (JJA) and spring (SON) are largely influenced by internal
498 IPO related SST variability (Boisier et al., 2016). On the other hand, high-latitude
499 atmospheric circulation associated with the SST warming signal dominates SWSA
500 rainfall trends especially in summer (DJF) and fall (MAM) (e.g., Boisier et al.,
501 2018).

502 To further evaluate the respective roles of the global warming and internal SST
503 variability on SWSA precipitation, a multilinear regression analysis over 1870-
504 2014 is performed using as predictands the simulated seasonal indices of SWSA
505 precipitation (i.e., area-weighted mean over boxed areas in Fig. 5) and the IPO
506 and GW observed SST indices as predictors (Fig. 7). The results show that the
507 IPO barely explains 19% and 15% of the precipitation low-frequency total variance
508 in JJA and SON, respectively, and that its contribution is almost null in MAM
509 and DJF (Fig. 7). The GW index explains a larger proportion of rainfall variance

510 in all the seasons except in JJA. The external forcing influence is exceptionally
511 remarkable in DJF, where the GW dominates the precipitation variance above the
512 IPO, which shows very little impact. However, the explained variance by the GW
513 index only represents the indirect external forcing influence on SWSA rainfalls
514 through induced SST anomalies. It does not account for the externally forced
515 changes in atmospheric dynamics and direct thermodynamic effects in SWSA.
516 The fact that the residual component is considerably large in all seasons suggests
517 that SST changes are not a significant driver of the SWSA precipitation variability
518 at decadal-to-longer time scales.

519 A possible explanation for the low response of the simulated SWSA precipita-
520 tion to the IPO could be due to the unrealistically weak atmospheric teleconnection
521 simulated by the model. However, the SLP anomalies regressively associated with
522 the IPO in JJA over southeastern Pacific (blue box in Fig. 6b) in HadSLP2 ob-
523 servations (0.24 hPa per standard deviation) lies within one standard deviation
524 of the ensemble-mean values of the amip-hist and historical simulations (respec-
525 tively, 0.32 ± 0.14 and 0.34 ± 0.15 hPa per standard deviation). Hence, we cannot
526 attribute the relatively weak influence of the internal SST decadal variability on
527 simulated rainfall to insufficient sensitivity of the atmospheric component of the
528 model to SST anomalies.

529 The SWSA precipitation response to the IPO simulated in amip-hist, histori-
530 cal and piControl simulations (see Table 1) are shown in the bottom panel of Fig.
531 7. MAM and DJF rainfall responses are positive in the amip-hist and historical
532 ensembles, but weak and not emerging from the internal variability as represented
533 by the piControl simulation. In turn, this positive signal is significant in amip-hist
534 in JJA and SON, the strongest being in JJA. In the historical simulations, the
535 response to the IPO is also positive and strongest in JJA but not significant as
536 in all seasons, consistently with the unforced piControl run. These observation
537 suggest that the difference in the SWSA response to IPO between forced and un-
538 forced coupled model simulations is not significant. Therefore, according to these

539 results, it can be inferred that the IPO can impact on the simulated SWSA pre-
540 cipitation. Still, its influence is weak as compared to external forcing and only
541 robustly reproducible with the SST-forced amip-hist simulations in JJA and, to a
542 lesser extent, in SON. Next, we examine the relative influences of dynamical and
543 thermodynamical external forcings on the low-frequency variability and trend of
544 SWSA rainfalls.

545 4.3 Role of dynamical vs. thermodynamical changes in SWSA

546 In the previous subsection, anomalous poleward shift of the zonal circulation in
547 high-latitudes in the SH is attributed to external forcings. This suggests that ex-
548 ternal forcings act on SWSA precipitation through dynamical changes. However, it
549 has been shown that long-term dynamical changes cannot account for all the exter-
550 nally forced trend in subtropical precipitation with direct thermodynamic effects
551 playing an important role (Schmidt and Grise, 2017). Therefore, the question arises
552 as to whether the forced component of precipitation in SWSA is mostly induced
553 by dynamic or thermodynamic processes. To shed light on the leading process, we
554 express the change in the net surface moisture budget as the difference of precipita-
555 tion minus evaporation (P-E) over 2005-2014 versus 1851-1910 ($\delta(P-E)$). Besides,
556 we decompose P-E into a thermodynamic component (δTH), due to changes in the
557 specific humidity, a dynamic component (δDC) due to changes in circulation, and a
558 third component associated with transient eddies (Seager et al., 2010). The change
559 of the annual mean P-E obtained from the ensemble-mean historical simulations
560 roughly presents a hemispheric pattern that represents the "wet gets wetter and
561 dry gets drier" paradigm of Held and Soden (2006) (Fig. 8a). It is worth noting
562 that the deficit of moisture supply in SWSA stands out across other extratropical
563 continental regions of the SH. The decomposition of the change of seasonal mois-
564 ture supply (Fig. 8b), reveals that changes in the atmospheric circulation account
565 for a large share of the total change compared to the direct thermodynamic effects
566 of external forcing in all seasons. Consistent with these findings, in the next sec-

tion, we examine the potential large-scale dynamics sources associated with the induced SWSA drying.

4.4 Connections between the HCE, SAM and SWSA rainfall

An anomalous shift of the circumpolar circulation in the SH high-latitudes, like the one shown by the regression patterns of the undetrended PC1s of SWSA precipitation on circulation forcings (Fig. 6), has been related to a persistent positive SAM trend and a widening of the HC in response to external forcings (e.g., Gillett and Thompson, 2003; Amaya et al., 2018). The historical ensemble mean shows that the seasonal low-frequency indices of SWSA precipitation, HCE and SAM describe a consistent trend (left panel in Fig. 9; note that the SAM indices are reversed). This trend is stronger since ~1970s and more pronounced in DJF and MAM than in JJA and SON. The square of the correlation coefficient (R^2) reveals high co-linearity among the three indices in all seasons (right panel in Fig. 9). SWSA precipitation is almost equally correlated to HCE and SAM, showing the highest co-linearity in DJF ($R^2 = 0.87$ and $R^2 = 0.92$, respectively) and the lowest in JJA ($R^2 = 0.73$ and $R^2 = 0.74$, respectively). In turn, the HCE and the SAM also show a strong co-linearity between each other that is strongest in DJF ($R^2 = 0.95$) and weakest in JJA ($R^2 = 0.86$). This result suggests that both dynamical large-scale atmospheric modes vary in synchronously and modulate SWSA precipitation in the same direction in response to external forcings. Since both HCE and SAM are highly co-linear, we next focus only on HCE index to investigate and attribute the simulated trends.

4.5 Attribution of forced variability

Until the early 1970s, the HCE indices of all coupled simulation oscillate around the piControl climatological value and within the threshold of internal variations in all seasons (Fig. 10, left panel). Afterwards, the historical ensemble-mean indices

593 show a poleward expansion in all seasons, even exceeding the bounds of internal
594 variations by the 2000s in JJA, SON and DJF. At the same time, the hist-nat ex-
595 periment simulates no appreciable change of the HCE variability. Such tendencies
596 evidence the role of the anthropogenic forcings on the recent HC expansion.

597 To highlight the role of external forcings, the linear trend over 1970-2014 of
598 the ensemble-mean HCE indices is analyzed (Fig. 10, middle panel). The trend
599 displayed by the historical simulation is negative, denoting a poleward expansion,
600 and significantly different from zero (with 99% confidence interval) in all seasons,
601 being even robust among all members in DJF. The expansion is widest in this
602 season with a trend of -0.20 °lat per decade, then -0.04 °lat per decade in MAM,
603 -0.07 °lat per decade in JJA and -0.11 °lat per decade in SON. Large error bars
604 evidence the great influence of internal weather noise. GHG forcing alone has a
605 large effect all year round, with a signature that emerges out of the internal noise
606 by the early 2010s in MAM and the 2000s in JJA and SON, with significant
607 1970-2014 linear trends of -0.06 , -0.09 and -0.10 °lat per decade, in the respective
608 seasons. Such trend values highlight the leading role of GHGs in the HC expansion
609 during these seasons. In DJF, instead, the stratospheric ozone depletion leads the
610 HC expansion, inducing wider shift (-0.11 °lat per decade) than GHGs (-0.06 °lat
611 per decade) separately. Anthropogenic aerosols (aer) induce a significant HCE
612 equatorward shift in JJA (0.02 °lat per decade) and SON (0.06 °lat per decade)
613 from around the mid-1990s with a peak in the mid-2000 and a partial recovery
614 afterwards. Nevertheless, this contraction is constrained within the threshold of
615 internal variability. The 1970-2014 linear trend of the "sum" index, obtained by
616 adding the ensemble-mean HCE anomalies from individual forcings, is significantly
617 close to the historical one (with 95% confidence interval) in DJF, MAM and JJA.
618 Therefore, the effects of the external forcings are, overall, additive all year round,
619 except in SON due to the influence of anthropogenic aerosols that tends to offset
620 the trend of the "sum" index (-0.04 °lat per decade) respectively to historical
621 experiments (-0.11 °lat per decade).

622 To quantify the role of the HCE in the forced SWSA drying over 1970-2014,
623 we represent the total trend of SWSA precipitation from the ensemble-mean sim-
624 ulations and the HCE-coherent trend in Fig. 10 (right panel). The historical ex-
625 periment shows a precipitation reduction of 4.5, 4.4, 3.6 and 10.3 mm per decade
626 in MAM, JJA, SON and DJF, respectively, in response to all external forcings.
627 The HCE-coherent trend underestimates by 40% the total drying in MAM. Still,
628 it shows no significant difference with the total trend in the other seasons, which
629 highlights the overall leading role of the HC expansion on the drying trend induced
630 by external forcing. In MAM, in response to only GHGs the total precipitation
631 trend is significantly similar to the one induced by all forcings and is strongly asso-
632 ciated with the HC expansion. The rest of individual forcings generate no signifi-
633 cant trends. In JJA and SON, GHGs alone induce a drying trend that respectively
634 doubles and equals that caused by all forcing together and that is tightly linked
635 to HC expansion. In both seasons, ozone depletion also contributes to drying, but
636 its impact is not associated with the HCE shift. In turn, aerosols and, to a lesser
637 extent, natural forcings partially counteract the drying. Aerosols positive effect
638 on precipitation is consistent with simultaneous HC contraction in these seasons.
639 Nevertheless, the HCE index barely accounts for 23% and 18% of the total trend
640 induced by aerosols respectively in JJA and SON. In DJF, the effect of GHGs
641 and ozone depletion separately induce each around half of the trend displayed by
642 the historical simulation, while the other forcings induce no significant trend. The
643 HCE-coherent trend underestimates by 60% the drying attributed to GHGs but
644 accounts for the entire trend induced by ozone depletion. The difference between
645 the ensemble-mean trend of the "sum" index of SWSA precipitation and the one
646 from the historical simulation suggest that the effects of individual forcings are
647 barely additive. Nevertheless, this difference is not statistically significant with
648 high significance interval, due to the considerable uncertainty among individual
649 members.

650 **5 Discussion**

651 5.1 The role of the IPO

652 Our results highlight the leading role of the SAM and the HCE on the recent
653 SWSA long-term drying trend in response to anthropogenic forcings. This is in
654 line with Boisier et al. (2018)'s finding showing that GHGs and the ozone de-
655 pletion drive the SWSA drying trend over 1960-2016 in association with positive
656 SAM anomalies. Nonetheless, Boisier et al. (2016) concludes that the forced dry-
657 ing trend at multidecadal scale can be substantially modulated by internal SST
658 variability through the IPO. This previous work focuses on the 1979-2014 period
659 and concludes, based on a linear regression model, that approximately 40% of the
660 SWSA drying trend is attributable to the concurrent pronounced negative trend
661 in the IPO index (Fig. 3). This statement opposes our result on the decomposition
662 of the decadal variance of SWSA precipitation, which reveals little influence of the
663 IPO compared to external forcing (Fig. 7).

664 To shed light on this discrepancy, we focus on the IPO-coherent trend of pre-
665 cipitation over 1979-2014 using the ensemble-mean amip-hist precipitation (upper
666 panel in Fig. 11). This trend exceeds the one attributable only to the internal vari-
667 ability of observed SST (i.e., the trend of ensemble-mean amip-hist precipitation
668 minus the trend from the historical ensemble mean) in all seasons. This diagnostic
669 evinces that a linear regression model can mislead the IPO signal on the SWSA
670 precipitation with the trend associated with external forcing. On the other hand,
671 the amip-hist ensemble-mean 1979-2014 drying trend notably exceeds the forced
672 component represented by the ensemble mean of the historical coupled runs. This
673 may suggest that the linear evolution of observed SST in 1979-2014, character-
674 ized by a steep negative IPO phase-shift, has a major contribution to the drought
675 along with external forcing. Nevertheless, the amip-hist ensemble-mean precipita-
676 tion trend may also represent an amplification of the forced drying by non-linear

677 air-sea interactions and atmospheric processes rather than the signature of the
678 specific IPO phase-shift.

679 To further assess whether the IPO phase shifts can directly induce decadal
680 trends in the simulated SWSA precipitation, we perform PDFs of the precipita-
681 tion trend over 1979-2014 and across all other possible 36-year periods since 1870
682 (bottom panel in Fig. 11). We use all the amip-hist members separately and rep-
683 resent the precipitation trend only in 36-year periods in which the IPO presents a
684 phase shift and significant linear trend. 50 IPO trend values are obtained, equally
685 distributed between positive and negative shifts ranging between $\pm 0.09^\circ$ C per
686 decade centered on respective means of 0.17° C and -0.16° C per decade. Then
687 the precipitation trend values are classified according to whether the shift is from
688 a negative to a positive IPO phase (red bars) or vice versa (blue bars). Accord-
689 ing to a two-sample t-test, the null hypothesis that the red and the blue PDFs
690 show normal distributions with equal means and variances cannot be rejected,
691 considering low confidence intervals ($p > 0.92$) in all seasons. Their means, repre-
692 sented by the red and blue solid vertical lines, show weak changes compared to
693 the ensemble-mean trend over 1979-2014 (green solid lines; corresponding to an
694 IPO shift of -0.21° C per decade). But they are negative in all cases, except in
695 MAM where the blue line indicates an almost null positive trend (0.2 mm per
696 decade). Therefore, these results suggest no significant relationship between the
697 linear trend of the simulated precipitation and the sign of the IPO phase shifts
698 over 36-year periods.

699 This conclusion is supported by similar PDFs performed using the 32 historical
700 members (not shown). The resulting PDFs are not significantly different between
701 each other and to the previous ones from the amip-hist simulations, according to
702 a two-sample t-test. In addition, the resulting mean trends (red and blue dashed
703 vertical lines in Fig. 11) are weaker than the drying represented by the historical
704 ensemble mean over 1979-2014 in all seasons (green dashed lines) but negative

705 in most cases, consistently with the amip-hist simulations, or positive but almost
706 zero in JJA.

707 In contrast, the same analysis applied to trends over 20-year or shorter peri-
708 ods across the amip-hist simulations reveals a break up between the red and blue
709 PDFs (not shown). They show prevailing positive (negative) rainfall trends con-
710 current with positive (negative) IPO phase shifts. This means that the amip-hist
711 simulations can reproduce SWSA precipitation trends in response to IPO phase
712 shifts at intradecadal-to-decadal time scales, in agreement with other works based
713 on instrumental data (Masiokas et al., 2010). However, at multidecadal-to-longer
714 timescales, such as the 36-year-long negative IPO trend observed since the 1979,
715 the simulated effect of the IPO over the last century and a half vanishes and
716 becomes insignificant compared to the leading role of external forcings.

717 In summary, our results show that the IPO explains little variance of SWSA
718 precipitation multidecadal variability over the last century and a half, compared
719 to external forcings. However, during 1979-2014 the simulated SWSA drying trend
720 attributable only to internal variability of observed SST accounts for a large part of
721 the total trend (Fig. 11). Therefore, we can conclude that the IPO is not a leading
722 cause of long-term variability of SWSA precipitation, but it can contribute to
723 amplify forced drying trends.

724 It has to be considered that the methodology used in this work is conditioned
725 by the IPSL-CM6A-LR model biases in the atmospheric circulation. Therefore,
726 the IPO influence on SWSA precipitation long-term variability should be assessed
727 in the multi-model framework of CMIP6 to strengthen this conclusion.

728 5.2 The simulated HC expansion

729 Regarding the simulated seasonality in HCE, model trend values are comparable
730 with those from observations, with simulations reproducing a poleward expansion
731 over recent decades. In agreement with Hu et al. (2011, 2018) and model-based
732 studies (Staten et al., 2012; Hu et al., 2013; Grise et al., 2018), the HCE expan-

733 sion shows strong seasonality, being larger in austral summer (DJF) than in other
734 seasons. Nevertheless, it is difficult to constrain a precise widening rate of the
735 HCE expansion due to the large spread among the trends from reanalyses and
736 the ensemble simulations. Regarding the wide range of trend values resulting from
737 the simulated ensemble members, it can be inferred that some part of the large
738 uncertainty is attributable to internal variability. In turn, observed and simulated
739 trends are comparable when the internal variability is considered, suggesting that a
740 large part of the observed HCE expansion is accounted for by internal atmospheric
741 variability (Grise et al., 2018). On the other hand, there is also large discrepancy
742 in the HCE expansion rates derived from the different reanalyses. Apart from dif-
743 ferences associated with distinct assimilation methods and model biases, this large
744 discrepancy has been attributed, in part, to shortcomings regarding the conserva-
745 tion of mass in the meridional mean circulation (Davis and Davis, 2018), especially
746 in old generation reanalyses (Grise et al., 2019).

747 5.3 HCE and SAM co-linearity

748 Our results show strong co-linearity between the SAM, HCE and SWSA precipi-
749 tation, evidencing that both modes induce changes on precipitation in the same
750 direction. Furthermore, the simulated SAM strengthening and the HCE expansion
751 are attributed to the same external forcings. This suggests a connection between
752 circulation changes in tropical and high latitudes through a link of both vari-
753 ability modes under the effect of external forcing, in agreement with previous
754 works (Thompson and Wallace, 2000; Previdi and Liepert, 2007). GCMs simu-
755 late an anomalous rise of the extratropical troposphere in association with the
756 SAM strengthening (Previdi and Liepert, 2007) and the expansion of the HCE
757 (Lu et al., 2007) in future projections that consider strong external forcing effects.
758 These anomalies are related to the increase of atmospheric static stability, which
759 in turn is associated with a poleward expansion of the tropospheric baroclinicity.
760 Subsequently, there is a shift in the same direction of the hemispheric circulation

761 such as the westerly jets, which involves changes in both the SAM (Thompson
762 et al., 2000) and the HCE (Staten et al., 2019), and in the mid-latitude storm-
763 tracks modulating the SWSA precipitation. Nevertheless, the factors that control
764 the extratropical atmospheric static instability are still unknown and, by exten-
765 sion, the mechanisms that explain the link between changes in the SAM and the
766 HCE.

767 5.4 The role of anthropogenic forcing

768 Our results show that anthropogenic forcings have largely induced the recent
769 SWSA drying through dynamical changes associated with the HCE and SAM.
770 GHGs play the leading role on the HC expansion and the subsequent SWSA dry-
771 ing trend all year round except in DJF. In this season, the effect of GHGs is
772 combined with the ozone depletion, which is the dominant factor. Although not
773 shown in this paper, the leading effects of GHGs and ozone depletion are found to
774 be similar on the HCE and the SAM. We also find that the effect of anthropogenic
775 aerosols can offset the simulated HC expansion and the SWSA drying trend over
776 1970-2014. This is most likely related to the counteracting effect of aerosols on the
777 global warming induced by GHGs (Andreae et al., 2005). However, the HC con-
778 traction barely explains a fraction of the forced rainfall increase, which is largest in
779 JJA (Fig. 10) coincident with positive thermodynamics of the P-E imbalance (Fig.
780 8). This suggests that aerosols may induce precipitation changes in SWSA through
781 thermodynamical processes. Nevertheless, aerosols effects are highly uncertain and
782 poorly constrained by climate models (Andreae et al., 2005; Boucher et al., 2013;
783 Carslaw et al., 2013; Oudar et al., 2018), suggesting that further investigation
784 should be done to shed light on this concern.

785 5.5 Projected of HCE and SWSA precipitation

786 Aerosol emissions are likely to decrease in the future intensifying the GHGs effect
787 (Andreae et al., 2005) concurrently with the stratospheric ozone recovery (Eyring
788 et al., 2010). Consequently, an acceleration of HC expansion associated with aerosol
789 depletion during JJA and SON could be expected in the coming years. At the same
790 time, a slowdown of HC would occur in DJF with ozone recovery. However, no solid
791 assumptions on the HCE future changes and the subsequent impacts on SWSA
792 precipitation can be made without taking into account the evolution of GHG
793 emissions. To address this concern, we analyze the ssp585 and the ssp126 future
794 projections, which incorporate the aforementioned future evolution of the aerosols
795 and stratospheric ozone under business-as-usual and mitigation scenarios of GHG
796 emissions simultaneously (Lurton et al., 2020). The resulting ensemble-mean in-
797 dices of HCE and SWSA precipitation are represented in Fig. 12, following those
798 of the historical simulations. The ssp585 projection shows a sustained poleward
799 expansion of the HCE in response to the substantial radiative forcing increase
800 until the end of the current century. This result is consistent with previous work
801 based on different GCMs (Grise et al., 2018; Staten et al., 2018). A SWSA drying
802 follows the HC expansion in all seasons. The projected drying exceeds the thresh-
803 old of extreme drought due to internal variability by the 2040s in MAM and JJA,
804 the 2030s in SON. In DJF, the drying is shown to be extreme since the 2000s,
805 according to the historical simulation. The yearly precipitation over 2091-2100 is
806 projected to be 37% lower with respect to 1851-1910 along with an annual-mean
807 HCE poleward shift of 2.4 °lat. In turn, ssp126 experiment projects a 13% loss of
808 yearly precipitation and an annual-mean HC poleward expansion of 0.7 °lat. This
809 mitigation scenario represents a stabilization of the HC expansion and the SWSA
810 drying close to the threshold of extreme rates in all seasons rather than a recovery.

811 Under the ssp585 scenario, an intense moisture deficit is also projected by the
812 end of the 21st century in other subtropical regions, apart from SWSA (Fig. 13).
813 Coastal areas in southern Angola and Namibia, south of South Africa, western and

southeastern Australia also show negative P-E change associated with the HCE evolution. The historical simulation, in contrast, does not show such strong link in these regions (Fig. 8a). This suggests that the HC expansion could become an essential modulator of precipitation in these mostly dry regions as in SWSA in a future with high GHGs emissions. To further understand and attribute the future evolution of precipitation in these regions, the contribution of the HC expansion against thermodynamical direct effects and the relative role of external forcings with respect to internal long-term variability should be addressed more thoroughly.

6 Conclusions

The IPSL-CM6A-LR model simulates an emerging long-term drying trend in SWSA roughly since the early 1980s in response to the observed SST and external forcings, consistent with observations and other GCMs (e.g., Vera and Díaz, 2015; Boisier et al., 2018). A modulating effect of the internal SST variability on the simulated SWSA drying over 1979-2014 is detected, related to the IPO as suggested by previous work (Boisier et al., 2016). However, the simulated impact of the IPO on the overall decadal variability of SWSA precipitation across the last century and a half is found to be weak and secondary compared to the effect of external forcings.

The external forcing modulates the simulated SWSA precipitation indirectly through dynamic changes, prevailing over direct thermodynamic effects. Specifically, this work relates the simulated drought in SWSA with a concurrent strengthening of the SAM and expansion of the HC, which occur in response to external forcings. Simulated forced changes in these two variability modes act on SWSA precipitation in the same direction and respond to the same components of external forcings. Both modes react more strongly to external forcing in DJF than in other seasons principally in response to the stratospheric ozone depletion and, in second place, to the GHGs, in agreement with previous studies (e.g., Polvani et al., 2011; Kim et al., 2017; Jebri et al., 2020). In the rest of the seasons, the

842 ozone depletion effect is weaker and the GHG forcing prevails. Future projections
843 suggest that the GHG effect determines the HCE, and consequently, the long-term
844 variability in SWSA precipitation throughout the 21st century.

Table 1 List of simulations with the IPSL-CM6A-LR model analyzed in the paper, the period covered, the size of the ensemble and imposed boundary conditions: observed SST, external forcing effects of GHGs, stratospheric ozone depletion, anthropogenic aerosols and natural forcings (namely stratospheric eruptions and spectral solar irradiance). (*) The forcing values in the ssp585 and ssp126 future projections are set according to the homonymous scenario of socio-economic development, while the rest of simulations include external forcings that are consistent with observations since 1850.

name	period	ensemble size	forcings				
			SST	GHG	ozone	aerosols	natural
amip-hist	1870-2014	20	✓	✓	✓	✓	✓
historical	1850-2014	32	–	✓	✓	✓	✓
hist-GHG	1850-2014	10	–	✓	–	–	–
hist-O3	1850-2014	10	–	–	✓	–	–
hist-aer	1850-2014	10	–	–	–	✓	–
hist-nat	1850-2014	10	–	–	–	–	✓
piControl	1200 years	1	–	–	–	–	–
ssp585	2015-2100	6	–	✓*	✓*	✓*	✓*
ssp126			–	–	–	–	–

Table 2 List of observations and reanalyses used in our analyses.

variable	data base	data type	resolution	period	reference
			(°lat × °lon × vertical levels)		
precipitation	CR2	rain gauge	in situ	1960-2017	http://www.cr2.c1
	GPCPv2018	gridded	$0.5^\circ \times 0.5^\circ$	1891-2016	Schneider et al. (2018)
	UDEL5.01			1900-2017	Willmott and Matsuura (2006)
SST	HadISST1	gridded	$1^\circ \times 1^\circ$	1870-2015	Rayner et al. (2003)
SLP	HadSLP2	gridded	$5^\circ \times 5^\circ$	1850-2017	Allan and Ansell (2006)
wind and surface pressure	NOAA-CIRES 20CR-V2	reanalysis	$2^\circ \times 2^\circ \times 24$	1871-2010	Compo et al. (2011)
	NOAA-CIRES 20CR-V2c		$2^\circ \times 2^\circ \times 24$	1851-2014	
	NOAA-CIRES 20CR-V3		$1^\circ \times 1^\circ \times 64$	1836-2015	Slivinski et al. (2019)
	ERA-20C		$2^\circ \times 2^\circ \times 37$	1900-2010	Poli et al. (2016)
	ERA-40		$1.1^\circ \times 1.1^\circ \times 23$	1970-2001	Uppala et al. (2005)
	ERA-Interim		$0.75^\circ \times 0.75^\circ \times 37$	1979-2017	Dee et al. (2014)
	NCEP1		$2.5^\circ \times 2.5^\circ \times 17$	1948-2017	Kalnay et al. (1996)
NCEP2	$2.5^\circ \times 2.5^\circ \times 17$	1979-2017			

Table 3 Ensemble-mean 1970-2014 linear trend in mm decade^{-1} of the amip-hist precipitation (δ) and of the precipitation anomalies coherent with the corresponding PC1 index (δ_{PC1}) averaged over the boxed areas in Fig. 5. All values are significant with 95% confidence interval.

	MAM	JJA	SON	DJF
δ	-8.54	-6.18	-6.31	-15.15
δ_{PC1}	-8.04	-5.45	-5.17	-15.81

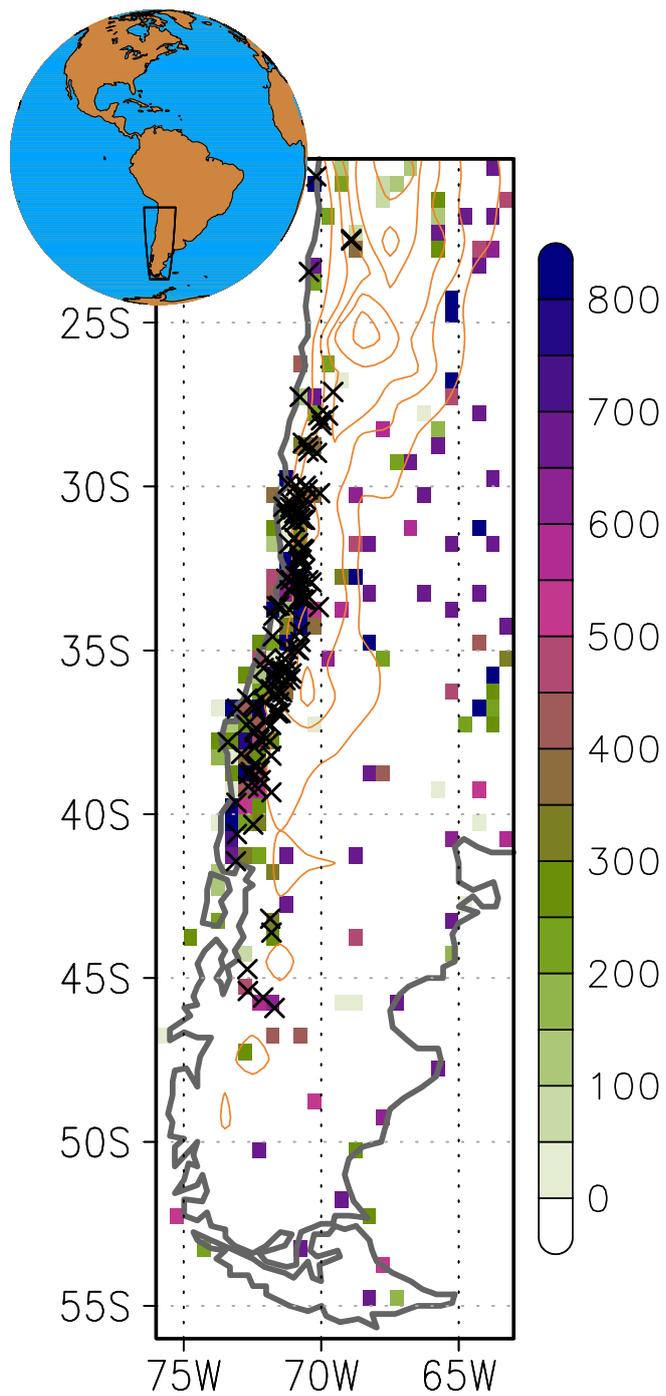


Fig. 1 Details of in situ precipitation observations in South Western South America (SWSA). Crosses: location of the 129 rain gauge stations of the observations used. Colored boxes: number of observed rainfall monthly data accumulated over 1960-2014 at 0.5° horizontal resolution in GPCCv2018. Orange contours: altitude levels in 1000-meter intervals.

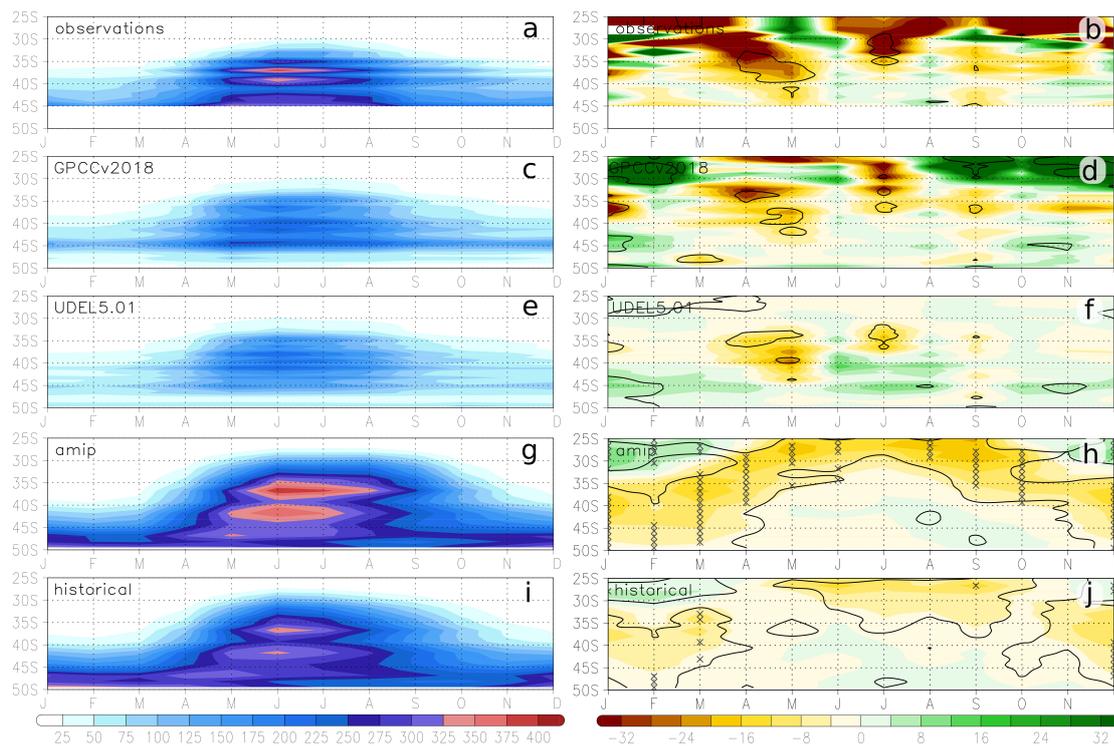


Fig. 2 Hövmoller diagrams representing the climatological annual cycle of monthly precipitation averaged over 1960-2014 (left panel; units are mm) and the linear trend over 1979-2014 (right panel; units are % decade⁻¹) averaged across 75° -70° W from (a-b) the rain gauge observations linearly interpolated to a regular grid of 1° latitude, (c-d) the GPCCv2018 and (e-f) UDEL5.01 gridded observations, (g-h) from the ensemble-mean amip-hist and (i-j) historical simulations. Crosses in (h) and (j) indicate where the trend in 80% of the simulation members show the same sign as the ensemble-mean. Contours in right panel indicate the trend significance at the 5% level according to a Student t-test with number of years minus one degrees of freedom.

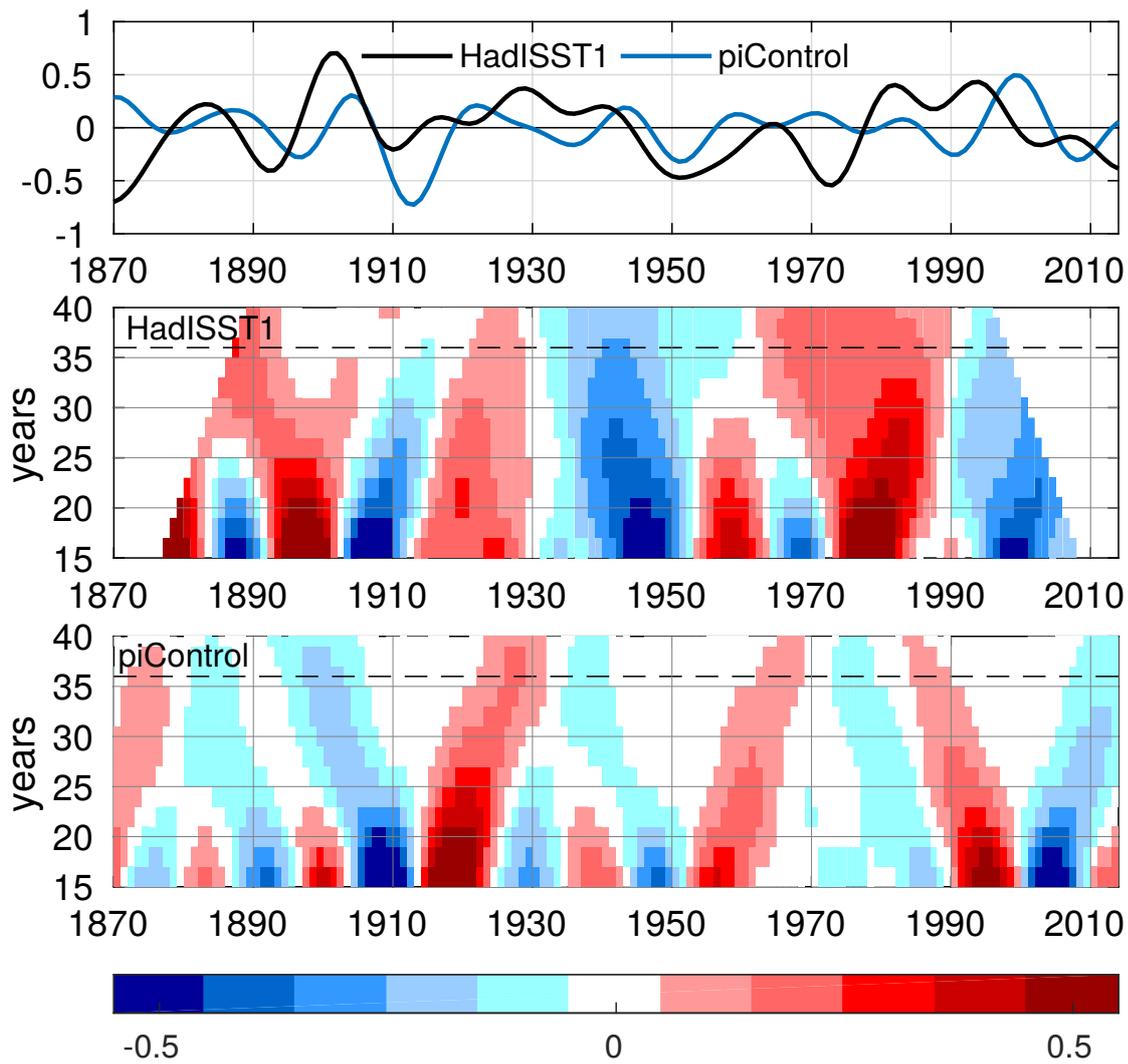


Fig. 3 Observed and simulated amplitude and persistence of the Inter decadal Pacific Oscillation Index (IPO) phase shifts. Upper panel: IPO indices from (black) HadISST1 observations over 1870-2014 and (blue) the piControl run over an equivalent 145-year period. Middle (bottom) panel: Trend values in units of $^{\circ}\text{C}$ per decade of the HadISST1 (piControl) IPO index in centered running windows of varying periods of 15-to-40 years (y axis). Dashed horizontal line indicates the 36-year running window equivalent to the 1979-2014 period.

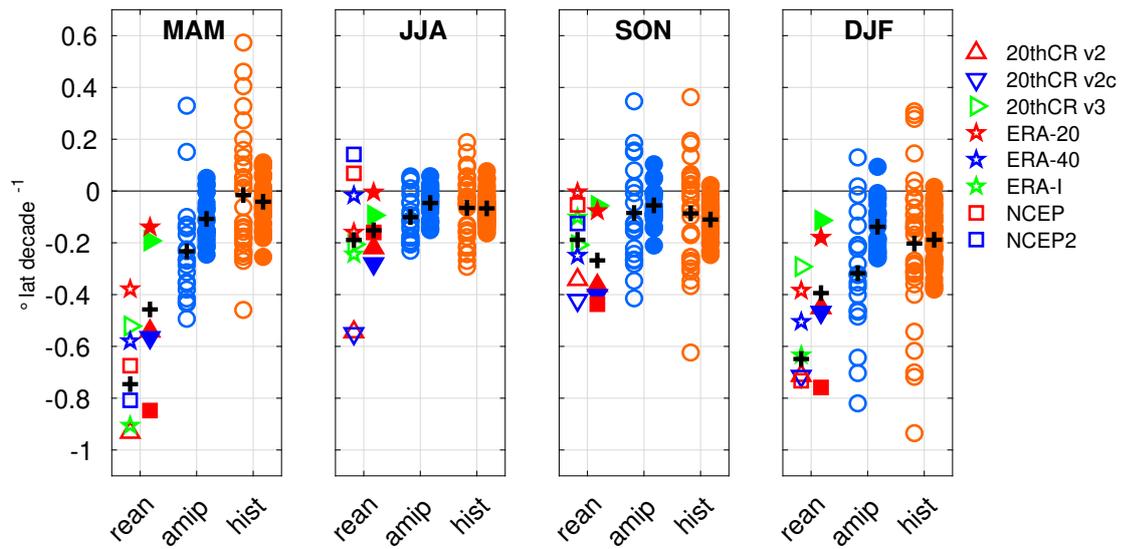


Fig. 4 Linear trend of the seasonal HCE index, expressed in latitude degrees per decade, obtained with reanalyses (*rean*; left column) and all members of the amip-hist (*amip*; blue circles) and the historical (*hist*; orange circles) simulations. Empty shapes correspond to the trend along the common period among all data used (1979-2001) and filled ones to a longer period of 40 years (1971-2010). Black crosses indicate the trend value averaged across all reanalyses or ensemble members.

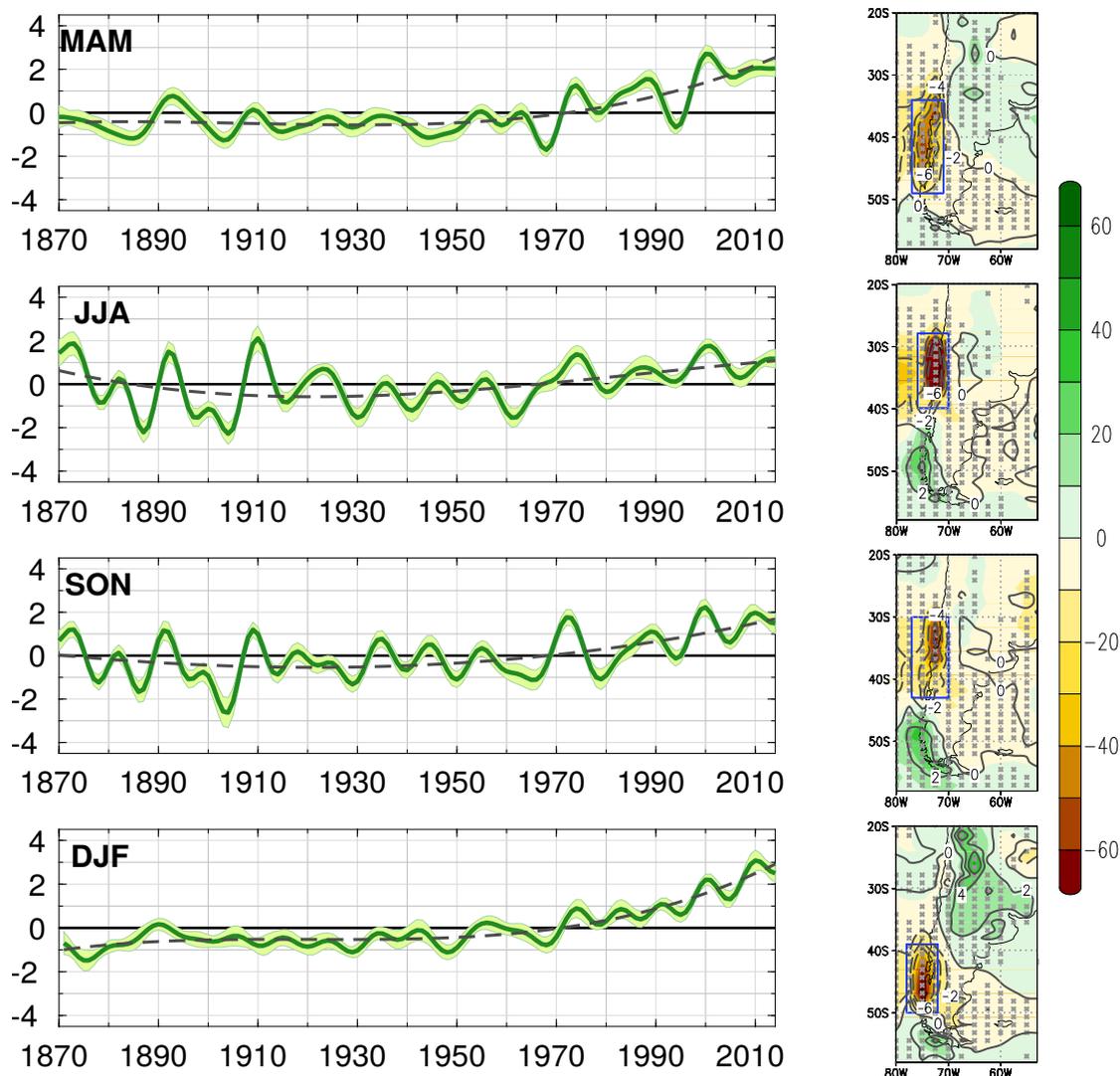


Fig. 5 Main seasonal modes of precipitation low-frequency variability. Left panel: Standardized PC1s of the 8-yr LPF seasonal precipitation anomalies over southern South America (58° - 25° S; 80° - 65° W) from the amip-hist ensemble-mean. Explained variance of the PC1s: 59.1% in MAM, 75.7% in JJA, 67.0% in SON and 69.9% in DJF. Dotted lines indicate the 3th-degree polynomial fit and green shading the spread among members at the 95% confidence level. Right panel: Regression patterns of the seasonal precipitation anomalies on the respective seasonal PC1 indices (shading; units are mm per standard deviation) and linear trend (contours; units are mm decade⁻¹) over the entire simulated period, 1870-2014. Gray dots indicate where there is at least 80% agreement among members regarding the sign of the regression coefficient.

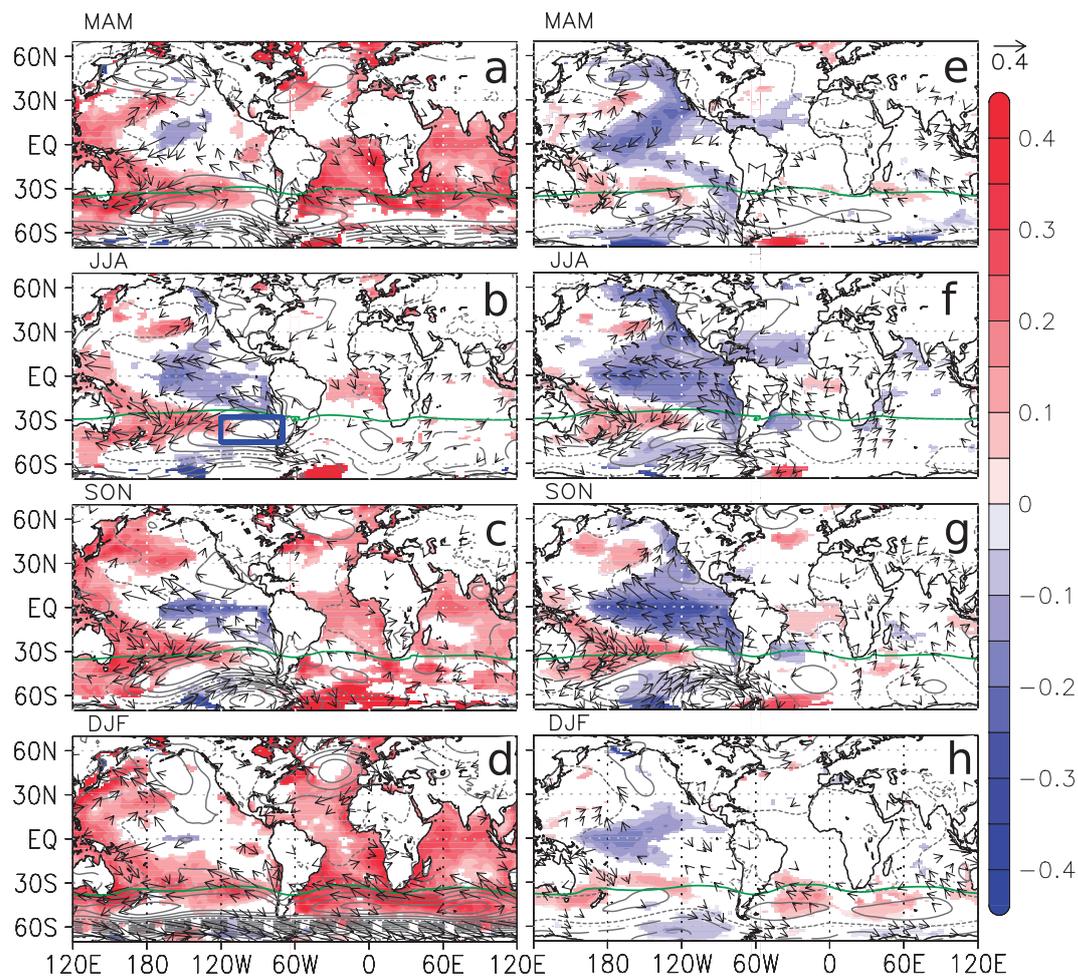


Fig. 6 Seasonal regression patterns of the non-detrended (left) and detrended (right) indices of precipitation on the seasonal anomalies of SST (shaded; units are $^{\circ}\text{C}$ per standard deviation), surface wind (vectors; m s^{-1} per standard deviation) and SLP (contours in intervals of 0.2 hPa per standard deviation). Regression values of the SST and wind anomalies with lower than 80% statistical significance, according to a random-phase test of Ebisuzaki, are masked. The green lines indicate the position of the climatological maximum of SLP. The patterns are computed using amip-hist ensemble-means. Blue box in panel b) indicate a center of maximum SLP anomalies between 70° - 120° W and 28° - 45° S.

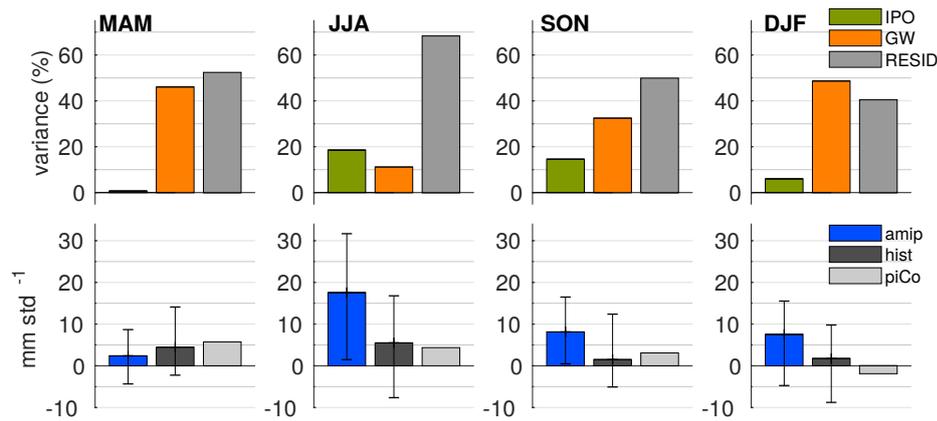


Fig. 7 Decomposition of the SWSA precipitation variance in response to the IPO and GW indices, and simulated precipitation sensitivity to the IPO. Upper panel: Bar charts of the components of the total variance (in %) of the LPF indices of amip-hist ensemble-mean seasonal SWSA precipitation explained by the IPO (green) and GW (orange) indices and the residual (gray) resulting from a multi-linear regression analysis. Bottom panel: Bar charts of the ensemble-mean regression coefficient between the SWSA precipitation seasonal anomalies and the standardized IPO index in amip-hist (blue), historical (black) and piControl (grey) simulations. Error bars indicate the total spread of the result obtained from individual members.

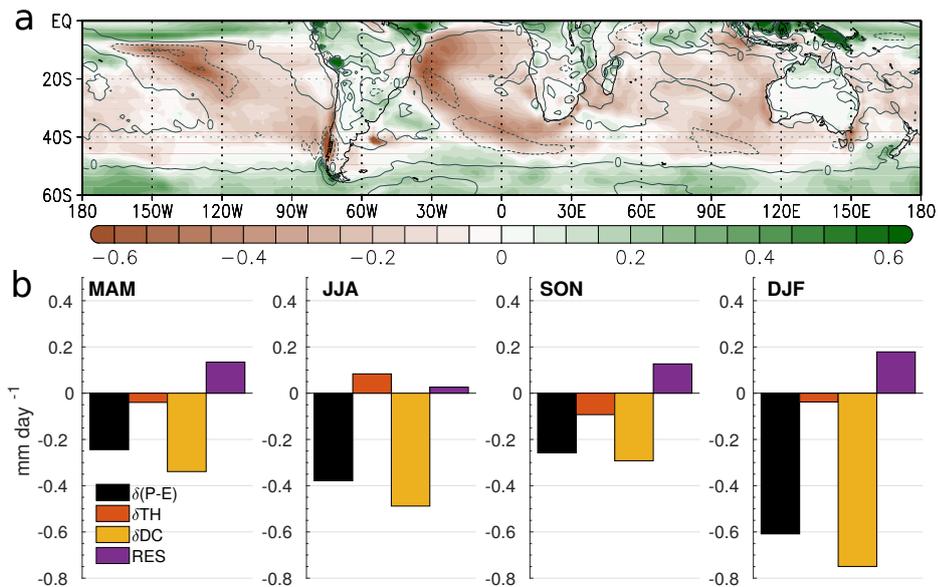


Fig. 8 Decomposition of the net moisture budget change in SWSA. (a) Colors indicate the change in the annual-mean P-E averaged over 2005-2014 with respect to 1851-1910 in units of mm day^{-1} . Contours (in intervals of 0.1 from -0.2 to 0.2 mm day^{-1} per standard deviation) indicate the regression of the annual P-E anomalies on the standardized annual HCE index over 1850-2014. (b) Bar charts of the seasonal-mean change in the net moisture budget (black), the thermodynamic (orange) and dynamic (yellow) components and the residual (purple) averaged over the corresponding seasonal SWSA areas (see boxes in Fig. 5). All plots are based on the ensemble-mean of the historical simulations.

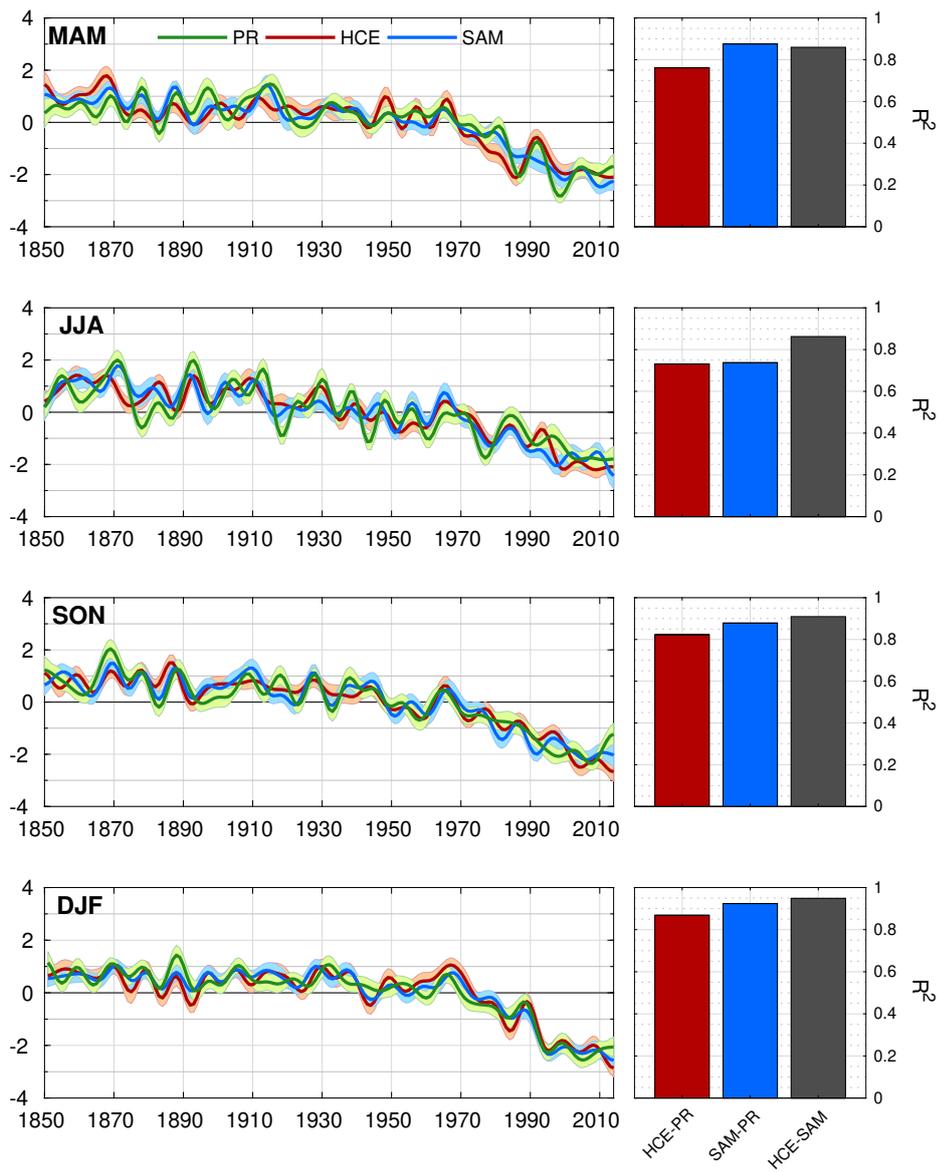


Fig. 9 Left panel: Standardized seasonal indices of (green line) SWSA precipitation (PR), (red line) HCE and (blue line) -SAM from the historical ensemble mean (units are standard deviation). Shading indicates the 95% confidence interval. Right panel: R^2 values between the ensemble-mean (red bar) PR and HCE indices, (blue bar) PR and SAM indices and (gray bar) HCE and SAM indices.

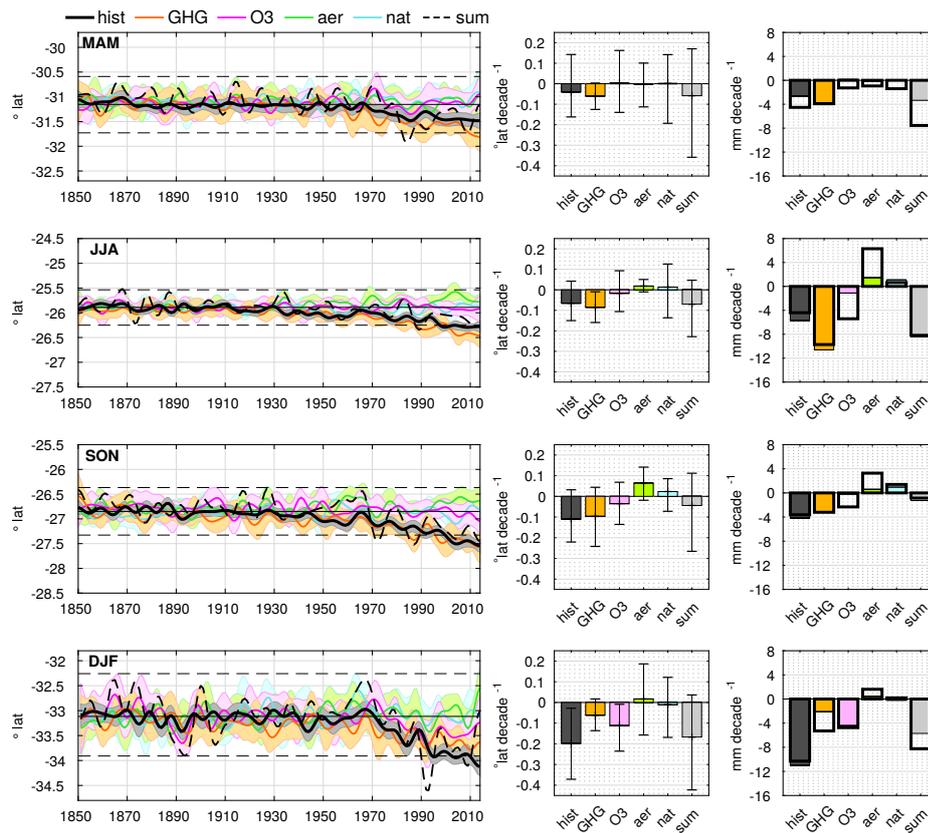


Fig. 10 Left panel: Ensemble-mean HCE indices with 95% confidence interval of the (black) historical, (orange) hist-GHG, (pink) hist-O3, (green) hist-aer, (cyan) hist-nat simulations and the (dashed black) index obtained from the sum of the individual attribution run anomalies. Horizontal lines represent the (solid) climatology and (dashed) 90% dispersion of the piControl HCE anomalies. Middle panel: (color bars) Ensemble-mean HCE index linear trend over 1970-2014 and (error bars) spread from individual members. Right panel: (outlined black bars) Linear trend of the low-frequency index of ensemble-mean SWSA precipitation over 1970-2014 and (color bars) coherent trend with the ensemble-mean HCE index ($\alpha \cdot \delta HCE$; where α is the regression coefficient calibrated over the entire simulated period and δHCE the HCE trend over 1970-2014).

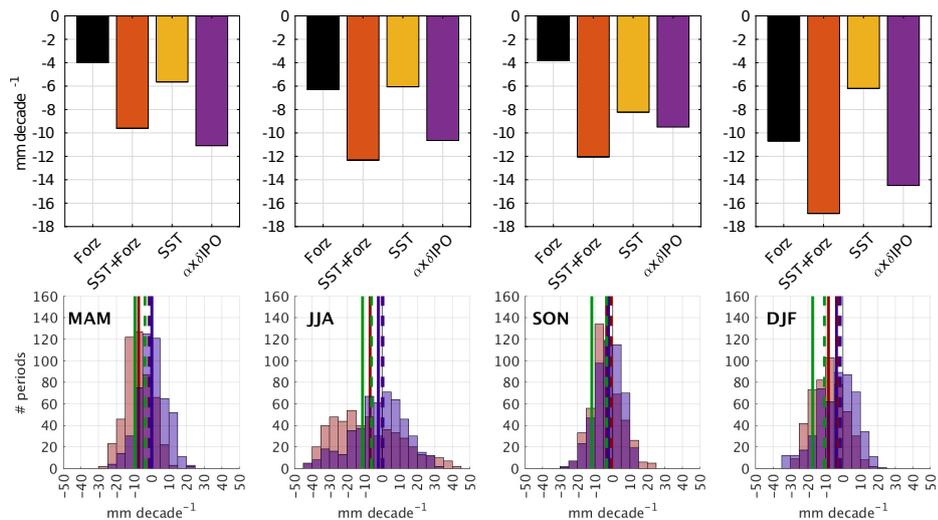


Fig. 11 Upper panel: Trend of the seasonal SWSA precipitation over 1979-2014 induced by external forcing (Forz) (i.e., computed from the historical ensemble-mean), by observed SST and external forcing (SST+Forz) (i.e., computed from the amip-hist ensemble-mean), by only observed SST (i.e., computed as the trend from the amip-hist ensemble-mean minus the one from the historical ensemble-mean) and the trend associated with the trend of the IPO index with a linear regression over 1979-2014 ($\alpha \cdot \delta IPO$). Bottom panel: PDFs of the seasonal SWSA precipitation linear trend (in mm decade^{-1}) over 36-year periods since 1870 in the amip-hist individual members in which the IPO index shows positive (red bars) and negative (blue bars) phase shifts. The red/blue vertical solid lines indicate the trend averaged over all values corresponding to positive/negative IPO shifts. Vertical green lines indicate the 1979-2014 trends in the ensemble-mean amip-hist. Vertical red, blue and green dashed lines represent equivalent trends to the respective colored solid lines but corresponding to the historical simulations.

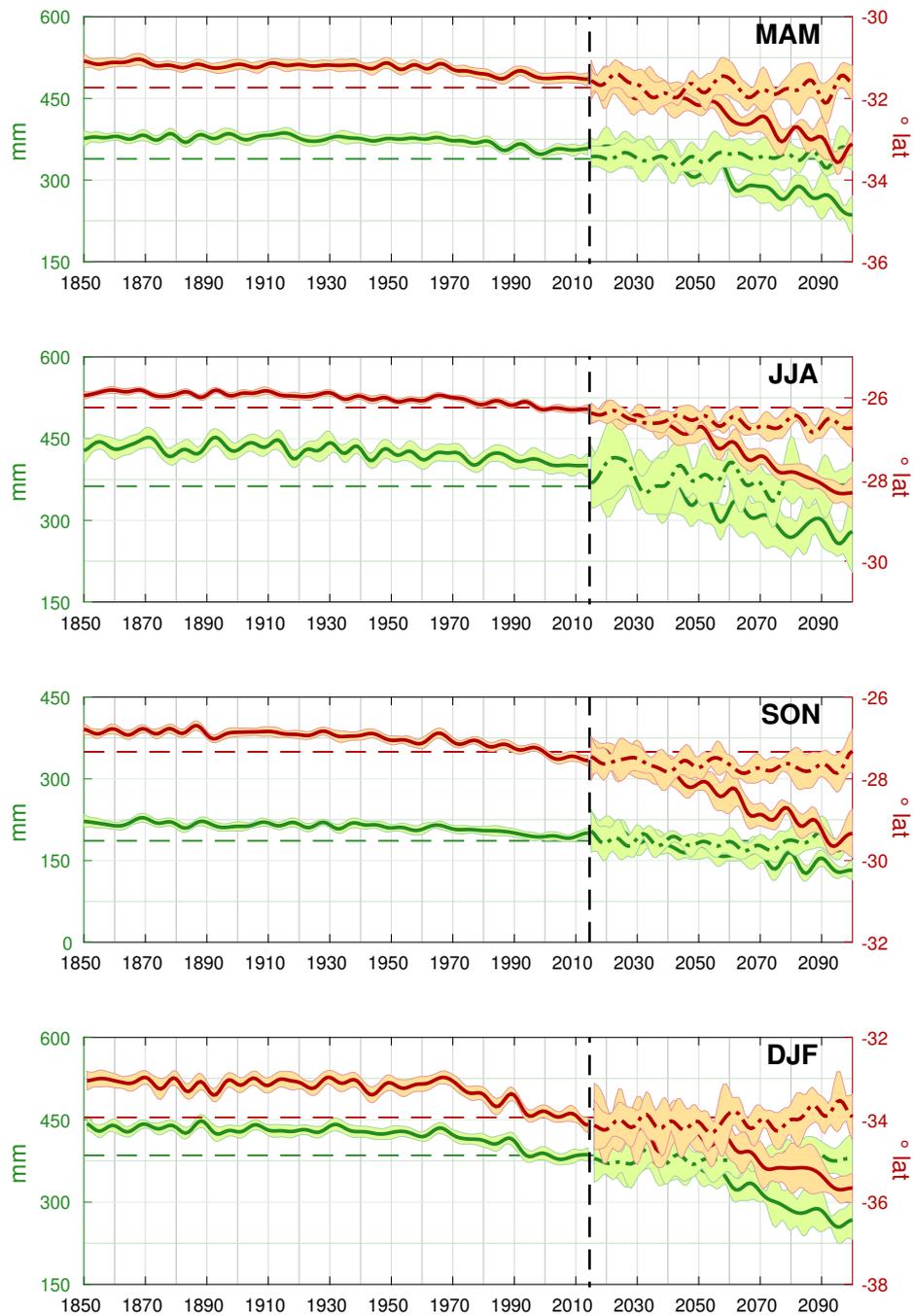


Fig. 12 Seasonal indices of (green) SWSA precipitation and (red) HCE, computed from the ensemble mean of the historical simulations in 1850-2014 and in 2015-2100 of (solid line) the ssp585 and the (dashed line) the ssp126 future projections. Shading indicates the 95% confidence interval. Horizontal dashed lines indicate the lower threshold of the dispersion of the piControl (red) HCE and (green) SWSA precipitation anomalies with 90% confidence interval.

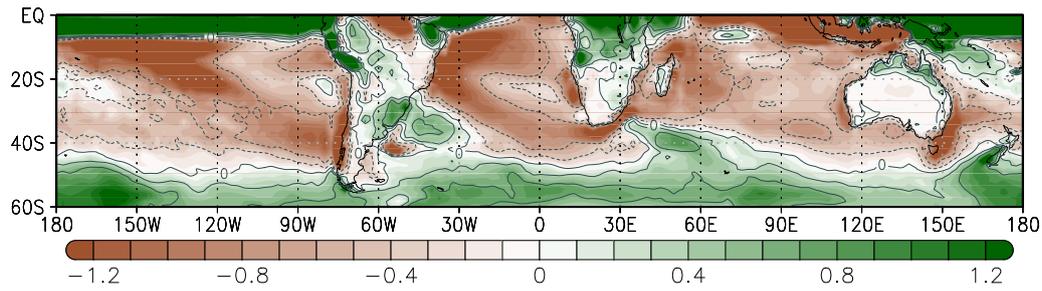


Fig. 13 Projected change of the net moisture budget in the Southern Hemisphere. Colors indicate the change in the annual-mean P-E (units are mm day^{-1}) averaged over 2091-2100 with respect to 2005-2014, based on ensemble-mean ssp585 and historical outputs, respectively. Contours (in intervals of 0.1 from -0.2 to 0.2 mm day^{-1} per standard deviation) indicate the regression of annual P-E anomalies on the standardized annual HCE index over 2015-2100.

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