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1 **Mid-Holocene to present-day evolution of the Indian monsoon**
2 **in transient global simulations**

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35 **Abstract**

36

37 The low-frequency evolution of Indian rainfall mean-state and associated interannual-to-
38 decadal variability is discussed for the last 6000 years from a multi-configuration ensemble of
39 fully coupled global transient simulations. This period is marked by a shift of Indian Summer
40 Monsoon Rainfall (ISMR) distribution towards drier conditions, including extremes, and a
41 contraction of the rainy season. The drying is larger in simulations with higher horizontal
42 resolution of the atmosphere and revised land surface hydrology. Vegetation-climate
43 interactions and the way runoff is routed to ocean modulate the timing of the monsoon onset
44 but have negligible effects on the evolution of seasonal rainfall amounts in our modeling
45 framework in which carbon cycling is always active. This drying trend is accompanied by
46 changes in ISMR interannual-to-decadal variability decreasing over north and south India but
47 increasing over central India (20°-25°N).

48

49 The ISMR interannual-to-decadal variability is decomposed into six physically consistent
50 regimes using a clustering technique to further characterize its changes and associated
51 teleconnections. From 6 to 3.8 kyr BP, the century-to-century modulations in the frequency of
52 occurrence associated to the regimes are asynchronous between the simulations. Orbitally-
53 driven trends can only be detected for two regimes over the whole 6 kyr BP to 0 kyr BP
54 period. These two regimes reflect increased influence of ENSO on both ISMR and Indian
55 Ocean Dipole as the inter-hemispheric energy gradient weakens. Severe long-term droughts
56 are also shown to be a combination of long-term drying and internally generated low-
57 frequency modulations of the interannual-to-decadal variability.

58

59 **Keywords:** Holocene, Indian Summer Monsoon, internal variability, orbital forcing, rainfall
60 mean-state and variability, transient simulations

61

62 **1. Introduction**

63

64 The Indian Summer Monsoon Rainfall (ISMR) provides about 80% of the annual rainfall
65 amounts over India from June to September (JJAS) with significant variability on a wide
66 range of timescales (Roxy and Chaithra 2018). Since ISMR variations profoundly impact on
67 livelihood, which mostly relies on rain-fed agriculture (Kesava Rao et al. 2020), improved
68 understanding of multi-scale ISMR variations is thus critical to reduce society vulnerability to
69 climate change and extremes. Two ranges of natural variability, interannual and decadal
70 timescales, are critical for adaptation planning and long-term mitigation.

71

72 At the interannual timescale, the El Niño Southern Oscillation (ENSO; Philander 1983) and
73 the Indian Ocean Dipole (IOD; Saji et al. 1999) are the two main modes of ocean-atmosphere
74 variability affecting ISMR. El Niño events are associated with an eastward shift of the Walker
75 circulation in the Indo-Pacific sector, which induces subsidence and dry conditions over India,
76 and *vice versa* during La Niñas (Walker 1924; Sikka 1980; Rasmusson and Carpenter 1983;
77 Wang et al. 2005). Positive IODs are associated with warm sea surface temperature (SST)
78 anomalies in the western tropical Indian Ocean (IO) and cold SST anomalies in the eastern
79 equatorial IO coupled to low-level easterly wind anomalies along the equatorial IO. Positive
80 IODs favor the existence of a reversed regional Hadley cell that enhances rainfall over India,
81 and *vice versa* for negative IODs (Ashok et al. 2001, 2004; Gadgil et al. 2004; Ashok and Saji
82 2007). However, positive IODs are often triggered by El Niños and negative IODs by La
83 Niñas (Behera et al. 2006). It is therefore nontrivial to disentangle the effects of both ENSO
84 and IOD on ISMR from observations alone. At the decadal time scale, ISMR is modulated by
85 the Pacific Decadal Oscillation (PDO; Mantua et al. 1997; Joshi and Kucharski 2017). A
86 negative/positive phase of the PDO promotes wet/dry ISMR anomalies. Additionally, a
87 further amplification of this link has been found when PDO and ENSO are in phase (Krishnan

88 et al. 2003; Roy et al. 2003; Krishnamurthy and Krishnamurthy 2013; Malik et al. 2017;
89 Malik and Brönnimann 2018). Finally, the Atlantic Multidecadal Oscillation (Enfield et al.
90 2001) can also influence ISMR (e.g., Malik et al. 2017; Malik and Brönnimann 2018).

91

92 The extent to which low-frequency modes of variability modulate ISMR variability at
93 interannual-to-decadal timescales is still an open question. The ISMR long-term fluctuations
94 have been related to changes in solar radiations induced by changes in Earth's orbit. For
95 example, mid- to late Holocene changes in orbital configuration weaken the inter-hemispheric
96 energy gradient as well as land-sea and moist static energy contrasts (e.g., Kutzbach 1981;
97 Joussaume et al. 1999). This leads to a decrease and southward shift of boreal summer
98 monsoons in both observations (e.g., Bartlein et al. 2011) and models (e.g., Liu et al. 2003;
99 Zhao et al. 2005; Braconnot et al. 2007a-b; Marzin and Braconnot 2009; Zhao and Harrison
100 2012; Harrison et al. 2015). Also, modeling studies have shown that the buildup of a seasonal
101 IOD-like pattern in autumn contributes to delay the monsoon withdrawal during the mid-
102 Holocene (Zhao et al. 2005; Braconnot et al. 2007).

103

104 Paleo reconstructions reveal that the mid- to late Holocene drying trend of ISMR is
105 punctuated by long periods of anomalously high/low rainfall that are not necessarily phased in
106 time across India (e.g., Kathayat et al. 2016, 2017; Kaushal et al. 2018; Kumar et al. 2019), as
107 well as mega-droughts such as the 4.2 kyr BP event (Staubwasser et al. 2003; Giesche et al.
108 2019). These studies suggest that regional changes in ISMR can often result from the highly
109 nontrivial interplay between internal variability and external forcing of the system.

110

111 Paleo-reconstruction archives are clues for characterizing and understanding past climates.

112 However, their spatial coverage and temporal resolution remain often too low or inconstant

113 over time to easily address changes in the interannual-to-decadal ISMR variability and
114 associated teleconnections during the Holocene. Transient global simulations run with general
115 circulation models are thus complementary tools to investigate ISMR evolution. Until
116 recently, such simulations were performed using accelerated changes in orbital forcing
117 (Lorenz et al. 2004, 2006; Varma et al. 2012) because of huge computational constraints. This
118 protocol does not allow us to examine the rhythm of changes nor interannual-to-decadal
119 variability since orbital forcing is accelerated by a factor 10 to 100 (Varma et al. 2016). Long
120 transient simulations performed with Earth System models and yearly-updated orbital forcing
121 (Liu et al. 2009; Jungclaus 2011) offer new possibilities to overcome these limitations since
122 they provide a thorough spatio-temporal continuum of climate evolution. Such simulations
123 have been used successfully to discuss the evolution of the Asian and African monsoons
124 during the Holocene (Dallmeyer et al. 2015, 2019; Shi and Yan 2019; Jalihal et al. 2019),
125 their multi-scale changes and teleconnections (Braconnot et al. 2019a-b), or mega-drought
126 during the 4.2 ka BP event (Yan and Liu 2019). They are also tailored for the investigation of
127 the complex spatio-temporal interrelationships in the ENSO-IOD-ISMIR system.

128

129 In this study, we build on the work of Braconnot et al. (2019b), who show that ISMR
130 interannual-to-decadal variability increases from mid- to late Holocene due to increased
131 ENSO variability in two 6000-yr transient simulations ran with the IPSL (Institut Pierre et
132 Simon Laplace) Earth system model. Here, we consider a multi-configuration ensemble of
133 transient simulations to refine their results. We investigate the effects of horizontal resolution,
134 land hydrology, vegetation and runoff on changes in ISMR, with a focus on its seasonality at
135 the monthly timescale and on its probability density function and extremes at the seasonal
136 timescale. We also investigate changes in the interannual-to-decadal variability of ISMR and
137 its teleconnections based on a clustering approach. Our aim is to highlight regional patterns in

138 ISMR interannual changes and to assess the temporal stability of ISMR teleconnections in the
139 transient simulations. The results of these analyses should also provide guidance to design
140 appropriate model-data comparison to analyze monsoon variability when dealing with a wide
141 range of spatio-temporal timescales.

142

143 Section 2 describes the simulations and the methodology used. Section 3 gives an overview of
144 trends in Indian rainfall seasonality, probability density function and extremes. Section 3 also
145 analyzes the rhythm and spatial consistency of changes at the local scale over the Indian
146 sector. Section 4 investigates changes in ISMR interannual-to-decadal variability and its
147 teleconnections through the regime approach with a focus on 6–3.8 kyr BP and discusses the
148 robustness of our results. Section 5 summarizes the results and gives the main conclusions.

149

150

151 **2. Model simulations and methodology**

152

153 2.1 The IPSL model

154

155 We consider two model versions of the IPSL Earth System model (Table 1). The first one,
156 called TR5A hereafter, is the IPSL-CM5A model version used for the CMIP5 ensemble of
157 past, present and future simulations (Dufresne et al. 2013; Mignot and Bony 2013) and
158 modified by Sepulchre et al. (2019). TR5A couples atmosphere, land surface, ocean and sea-
159 ice, and accounts for energy, water and carbon cycles. The second version, called TR6A
160 hereafter, includes a new 11-layer hydrological model, a prognostic 3-layer snow model and
161 the possibility to switch on a dynamical vegetation component (Braconnot et al. 2019a). In
162 both model versions, the vegetation in each grid box is represented by 13 Plant Functional

163 Types (PFTs) and carbon cycle is interactive. The leaf area index (LAI) is thus fully
164 interactive and varies depending on energy, water and carbon cycles regardless of the model
165 version. When vegetation is prescribed, the 13 PFTs are assigned to a fixed land cover map.
166 When vegetation is computed, only natural vegetation is accounted for and the distribution of
167 the 13 PFTs in each model grid cell varies with time.

168

169 The resolution of the atmosphere is $3.75^\circ \times 1.875^\circ$ for TR5A and $2.5^\circ \times 1.25^\circ$ for TR6A on
170 the horizontal and 39 levels for the two versions on the vertical. For ocean, TR5A and TR6A
171 share the same 2° ORCA configuration (Madec et al. 2017), with a tri-polar grid, mesh
172 refinement at the equator and 31 vertical levels. River and direct runoff are routed to the
173 corresponding ocean coastal grid box. Specific schemes connect snow accumulation over the
174 ice sheet and endorheic basins to the ocean (Marti et al. 2010) to ensure a perfect closure of
175 the water cycle. The corresponding algorithm now uses a full parallel version of the
176 interpolation weights between the land-atmosphere and ocean grids. The parameters of the
177 new algorithm were chosen to produce similar results as before. This new version of the land-
178 ocean runoff is called CM6 in Table 1.

179

180 2.2 Transient simulations

181

182 An ensemble of five transient simulations is performed with yearly changes in orbital
183 parameters and trace gases following the protocol proposed by Otto-Bleisner et al. (2017).
184 This ensemble accounts for differences in model resolution, the representation of land-surface
185 hydrology, snow and river runoff, as well as dynamical vegetation. Details on major
186 components of each simulation can be found in Table 1.

187

188 Two simulations, TR5AS-Vlr01 and TR6AV-Sr02 span the time range from 6 to 0 kyr BP
189 (Braconnot et al. 2019a-b), with 0 kyr BP corresponding to 1950 Common Era. The
190 remaining simulations also start at 6 kyr BP but are 4000-yr or 2250-yr long integrations
191 (Table 1).

192

193 The TR5AS-Vlr01 is run with the TR5A model version and with vegetation prescribed to the
194 IPSL-CM5A 1850 reference map (Dufresne et al. 2013). The other simulations are run with
195 the TR6A model version. The TR6AV-Sr02 simulation has dynamical vegetation. The
196 TR6AS-Sr10 and TR6AS-Sr12 simulations are similar to TR6AV-Sr02, except that
197 vegetation is prescribed to the 1850 reference map for the former and to the mid-Holocene
198 natural vegetation as computed by the model for the latter (Table 1). An error was found in
199 the first release of the interpolation weights, which might affect the results of the TR6AV-
200 Sr02, TR6AS-Sr10 and TR6AS-Sr12 simulations (Table 1). In these simulations, a fraction of
201 the river and direct runoff is redistributed globally and not at the river mouth (“inter” in Table
202 1). This doesn’t affect the water closure of the model, but the ocean regional circulation. A
203 simulation similar to TR6AS-Sr10, called TR6AS-Sr11 (Table 1), was thus run to test the
204 effect of this bug (“inter”) compared to the CM6 land-ocean river runoff.

205

206 The different model versions and model setups produce slightly different mean climate and
207 climate characteristics (Braconnot et al. 2019a). The initial state for each transient simulation
208 results from a 1,000-yr long mid-Holocene simulation (i.e., with Earth’s orbit and trace gases
209 prescribed to mid-Holocene values and kept constant throughout the simulation) performed
210 with the corresponding version of the model. The initial states are thus entirely consistent
211 with the mid-Holocene climate simulated by each model configuration, ensuring no artificial
212 model drift at the beginning of the transient simulations.

213

214 2.3 Statistical analysis of Indian rainfall and teleconnections

215

216 Complementary analyses allow us to characterize the long-term evolution of ISMR and of its
217 variability. We first use a rainfall index area-averaged over India (land points within 5° - 25° N
218 and 77° - 88° E) to assess changes in (i) the seasonality of the Indian rainfall mean-state and its
219 variability at the monthly timescale, (ii) the probability density function of seasonal ISMR
220 amounts during boreal summer (i.e., JJAS) and (iii) extreme ISMR years (i.e., flood and
221 drought years). This domain embeds regions strongly affected by boreal summer Indian
222 monsoon. Analyses are performed considering 100-yr adjacent windows from 6 kyr BP
223 onwards. Rainfall mean-state is defined as the average within each 100-yr window, while
224 rainfall variability is defined either as the standard-deviation within each window or as
225 anomalies with respect to the mean-state estimated in the window. The 100-yr length of the
226 windows is retained to focus on interannual-to-decadal variability. The century-to-century
227 evolution of these descriptors will be interpreted as centennial-to-multicentennial modulation
228 of rainfall mean-state and interannual-to-decadal variability.

229

230 To account for potential spatial disparity in the ISMR evolution that are hidden in the area-
231 averaged index, we then adopt a “regime” approach based on the Agglomerative Hierarchical
232 Clustering (AHC; Gong and Richman 1995). This technique, presented in Appendix A, is
233 retained to detect recurrent anomalous rainfall patterns over India at the interannual-to-
234 decadal timescale. It is preferred to Empirical Orthogonal Function analysis because it allows
235 us to highlight regional nuances without imposing artificial north–south or west–east
236 structures induced by orthogonality relationships and to account for nonlinearities in the
237 teleconnection between (wet and dry) ISMR regimes and ENSO.

238

239 For the AHC analyses, anomalies of JJAS rainfall patterns are computed for each adjacent
240 100-yr window as the departures from the corresponding rainfall mean-state inside the
241 window. The AHC is fed by the 5 simulations simultaneously to focus on the shared signals
242 across the simulations. The AHC provides a hierarchy of clusters from N to 1, with N being
243 the total number of classified (time) patterns. The patterns are grouped based on their
244 similarity and to minimize intra-cluster variance and maximize inter-cluster variance. The
245 optimal number of ISMR regimes (i.e., clusters) is a compromise between the intra- and inter-
246 cluster variance and the physical meaning of the identified clusters. Here, we detect 6 ISMR
247 regimes using different metrics (see Appendix A for details). The 6 ISMR regimes are then
248 characterized by computing their center of gravity, that is, by averaging all anomalous
249 patterns in JJAS rainfall, diverse atmospheric fields and SST belonging to each regime. Their
250 robustness is assessed through statistics to test whether or not the centers of gravity of the
251 regimes differ significantly from the center of gravity of the whole population (Student t-test
252 at the 95% confidence level) and by discussing the similarity between them and those
253 associated to each simulation (e.g., the center of gravity of all patterns of TR5AS-V1r01
254 belonging to regime #1).

255

256 2.4 Representation of the Indian monsoon by the IPSL model

257

258 The two model versions reasonably capture the mid-Holocene wetting of the Northern
259 Hemisphere monsoon regions (Kageyama et al. 2013a-b; Braconnot et al. 2019a) and the
260 teleconnection between the Indian monsoon and ENSO at the interannual timescale during the
261 recent period (Braconnot et al. 2019b). Here, their ability in capturing both the mean-state and
262 interannual-to-decadal variability of Indian rainfall is assessed for modern conditions against

263 three observational monthly products: the All Indian Rainfall index (AIR; Parthasarathy et al.
264 1995), which is an area-weighted average of 306 rain gauges distributed across India from
265 1871 onwards, and the GPCP version 2.2 (Adler et al. 2003) and CRU-TS4.01 (Harris et al.
266 2014) data.

267

268 Regarding rainfall mean-state, the magnitude of the monsoon peak and JJAS amounts is close
269 to observations in TR6AV-Sr02, while depicts a ~ 2 mm/day dry bias in TR5AS-Vlr01 (Fig.
270 1a and Table 2). The latter dry bias is typical of low-resolution global coupled simulations
271 (Sperber et al. 2013; Goswami and Goswami 2017; Terray et al. 2018). The improvement of
272 TR6AV-Sr02 on TR5AS-Vlr01 is due to a better penetration of the monsoon flow along the
273 Himalaya foothill induced by increased horizontal resolution (Fig. 2) and to increased local
274 recycling induced by the 11-layer hydrology model (Braconnot et al. 2019a). On the other
275 hand, horizontal resolution and hydrology do not affect the onset and demise phases of the
276 Indian monsoon, which are similar between the two model's versions. The fast withdrawal
277 observed between September and October is accurately captured, while the onset observed in
278 June is delayed by 1 month in the simulations (Fig. 1a). The 1-month delay in the onset
279 results from a too slow northwestward propagation of the rain belt during early summer over
280 India (Fig. 2). This slow propagation involves a too late inversion of the meridional
281 tropospheric temperature gradient between landmass and the IO due to a persistent cold bias
282 over the Himalayas (Marzin and Braconnot 2009) and Indo-Pacific SST biases intrinsic to
283 ocean-atmosphere coupled simulations (Sperber et al. 2013; Prodhomme et al. 2015). Note,
284 however, that the northernmost extent of the monsoon in August and September is consistent
285 with observations (Fig. 2).

286

287 Regarding rainfall variability, the observations depict a plateau of 1.5 mm.day^{-1} from June to

288 September when considering the standard-deviation of the long time series in the AIR and
289 CRU data (Fig. 1b). This plateau hides large observed variability of the monsoon onset and
290 demise when considering 20-yr long time series (see GPCP and thin curves for AIR in Fig.
291 1b). This double peak is captured by the two simulations despite a clear 1-month delay of the
292 first peak linked to the aforementioned bias in the mean-state and exaggerated magnitude,
293 especially in the TR6AV-Sr02 (Fig. 1b and Table 2, second column).

294

295

296 **3. Trends in seasonality and boreal summer Indian monsoon rainfall**

297

298 3.1 Evolution of Indian rainfall seasonality

299

300 The evolution of Indian rainfall seasonality, estimated in 100-yr adjacent windows, is affected
301 by the insolation and trace gas forcing (Fig. 3). The TR5AS-Vlr01 and TR6AV-Sr02 depict
302 decreased monthly rainfall amounts throughout the seasonal cycle and reduced length of the
303 Indian monsoon season from mid- to late Holocene (Fig. 3a-b), with delayed onset and early
304 withdrawal. The delayed onset is not affected by the calendar effect induced by changes in
305 orbital forcing in early boreal summer when the vernal equinox is set to March 21 in all
306 simulations (Joussaume and Braconnot 1997; Bartlein and Shafer 2019). Daily output would
307 be required to fully assess the calendar effect on the early withdrawal since this effect is
308 maximal in late boreal summer with the calendar reference used here. However, the reduced
309 length of the Indian monsoon is consistent with Marzin and Braconnot (2009). These authors
310 analyzed daily values and suggested that the gradual warming of the tropical ocean and
311 cooling of the Northern Hemisphere late summer SST throughout the Holocene are the main
312 drivers of the reduced length of the Indian monsoon from mid- to late Holocene. Despite the

313 drying trend in rainfall mean-state (Fig. 3a-b), TR5AS-Vlr01 and TR6AV-Sr02 simulate
314 increased interannual-to-decadal variability with time, in June-July for TR5AS-Vlr01, but for
315 all monsoon months in TR6AV-Sr02 (Fig. 3c-d). The opposite trends between rainfall mean-
316 state and rainfall variability highlight the need to carefully untangle mean-state and variability
317 evolutions in climate change studies.

318

319 The broad features of rainfall mean-state seasonality of the other simulations are close to
320 TR6AV-Sr02. It is, however, worth noting that vegetation-climate feedbacks and the way
321 river runoff is routed to ocean have significant impacts on the onset and early part of the
322 Indian monsoon (Fig. 3e-h). These results are consistent with the sensitivity experiments of
323 Sooraj et al. (2019), who have demonstrated that albedo perturbations play a significant role
324 on the Indian monsoon during the first part of the rainy season (e.g. June-July), while
325 convective processes dominate ISMR afterward. The simulation with “inter” river runoff
326 (TR6AS-Sr10: Fig. 3e) is much wetter in June-July than the simulation with CM6 runoff
327 (TR6AS-Sr11: Fig. 3f). In addition, the simulation with prescribed 1850 vegetation (TR6AS-
328 Sr10: Fig. 3e) is wetter in June and drier in July than the simulations with prescribed mid-
329 Holocene vegetation (TR6AS-Sr12: Fig. 3g) or dynamical vegetation (TR6AV-Sr02: Fig. 3h).
330 Finally, the use of a fixed vegetation map in TR6AS-Sr12 induces a significant shift of ISMR
331 from the second half to the first part of the rainy season from 6 to 3.8 kyr BP (Fig. 3e), but
332 this specific time evolution is not simulated with a dynamical vegetation in TR6AV-Sr02. A
333 dedicated study is needed to understand the mechanisms by which vegetation-climate
334 feedbacks and river runoff affect the Indian monsoon onset. The interannual-to-decadal
335 variability remains very noisy from one TR6A simulation to another from 6 to 3.8 kyr BP (not
336 shown), highlighting the chaotic nature of this range of variability.

337

338 3.2 Pace of change in Indian summer monsoon rainfall

339

340 We now consider JJAS seasonal averages to assess the pace and magnitude of changes in
341 rainfall mean-state and variability during the Holocene. The JJAS season is chosen to
342 encompass differences in Indian rainfall amounts arising from changes in both the amplitude
343 and length of the monsoon with time. The Probability Density Function (PDF) of JJAS
344 rainfall amounts shifts towards lower values from mid- to late Holocene in the two 6000-yr
345 long simulations (Fig. 4a-b). While the shape of the PDF barely varies along the period in the
346 TR5AS-Vlr01 (Fig. 4a), it flattens and is more skewed to the left towards present-day
347 conditions in TR6AV-Sr02 (Fig. 4b). This reflects stronger changes in ISMR interannual-to-
348 decadal variability in the TR6AV-Sr02 than the TR5AS-Vlr01.

349

350 Both experiments simulate changes that are more significant for the 10% driest than 10%
351 wettest seasons estimated through a percentile approach (Fig. 4c). This confirms that the
352 drying concerns not only the center of the distribution, but also the drought years, which
353 become more severe towards present-day climate, especially in TR6AV-Sr02. The evolution
354 in the number of extreme ISMR seasons in each simulation (and not on each 100-yr window
355 extracted from the simulations) highlights that the drying trend in ISMR mean-state results
356 also from increased occurrence of droughts (Fig. 4d) rather than decreased occurrence of
357 floods (Fig. 4e). This result is very robust across the 5 simulations and is thus interpreted as a
358 response to orbital forcing. Note that the magnitude of changes is weaker in TR5AS-Vlr01
359 than the other simulations, especially after 4 kyr BP (Fig. 4d).

360

361 The evolution in the intensity and, to a greater extent, frequency of the driest and wettest JJAS
362 seasons depicts large inter-centennial fluctuations (Fig. 4c-e). The most striking example is

363 the evolution of the 10% driest ISMR seasons with up to 500 consecutive years depicting
364 increase or decrease in their occurrence (Fig. 4d). Associated changes exceed, by far, the
365 magnitude of the trend simulated for the last 6000 years. Such events are independent from
366 orbital forcing since they are not synchronized between the simulations and occur at any time
367 during the last 6000 years (Braconnot et al. 2019b).

368

369 3.3 Broader context of boreal summer changes

370

371 The broader context of JJAS changes in 100-yr mean-state and associated interannual-to-
372 decadal variability is now analyzed in the framework of Indo-Pacific rainfall and SST patterns
373 to provide a more comprehensive view of ISMR changes (Figs. 5-6). We focus the discussion
374 on the aspects that emerge in all simulations but show the results only for the TR6AV-Sr02
375 from 6 to 0 kyr BP.

376

377 Compared to 6-5.9 kyr BP, significant dry rainfall anomalies emerge over the Himalayas, and
378 immediately east of the western Ghats as soon as 5.9-5.5 kyr BP. They rapidly expand over
379 all India and grow in intensity afterwards (Fig. 5; contours). This drying reflects the direct
380 response of the Indian monsoon to the orbitally-driven weakening (strengthening) of
381 seasonality in the Northern (Southern) Hemisphere (Braconnot et al. 2007a). It is associated
382 to progressive SST cooling in most regions of the Northern Hemisphere and SST warming in
383 the Southern Hemisphere (Fig. 6; contours). Over the IO, anomalous SSTs reflect a
384 combination of these orbitally-driven global forcing and of changes in monsoon circulation
385 (Zhao et al. 2005). The negative rainfall anomalies trend over the IO between the Equator and
386 8°N (Fig. 5) is a key feature that emerges also in coral-based reconstructions (Abram et al.
387 2007) and is coupled to a shift towards negative IOD conditions in simulated SST patterns

388 (Fig. 6).

389

390 The pattern of changes in rainfall variability (Fig. 5; shadings) is well correlated to the pattern
391 of changes in rainfall mean-state (Fig. 5; contours). Regions experiencing a drying trend are
392 characterized by decreased variability, and *vice versa*. The only exception is central India
393 (between 20°N and 25°N) where rainfall mean-state decreases but rainfall variability starts
394 increasing after 3 kyr BP and keeps increasing until late Holocene (Fig. 5; shadings). The
395 increased rainfall variability over central India is embedded in a pattern extending from the
396 northeastern part of the Arabian Sea to the Bay of Bengal. This suggests a coherent large-
397 scale amplitude variation of the monsoon trough over land and ocean in the IO region.
398 Moisture convergence into central India and the monsoon trough becomes thus more variable
399 with time. This also indicates that the increased variability of area-averaged Indian rainfall
400 discussed in Braconnot et al. (2019b) reflects the behavior of central India.

401

402 Contrary to rainfall, the pattern of changes in SST variability (Fig. 6; shadings) does not
403 resemble that of changes in SST mean-state (Fig. 6; contours). For the mean-state, the main
404 tropical features in the simulations are the emergence of a well-defined cold tongue along the
405 Equator in the eastern Pacific and eastern Atlantic and of a negative IOD-like SST pattern in
406 the IO. Consistent with these mean-state changes and slow adjustments by ocean dynamics,
407 SST variability increases in the central-to-eastern equatorial Pacific (e.g. enhanced ENSO
408 variability) and the eastern equatorial Atlantic, but decreases in the eastern equatorial IO (off
409 Java). Interestingly, the SST variability is enhanced in the western tropical IO, despite the
410 decreased SST variability in the eastern tropical IO. This suggests that IOD modulations are
411 not the main trigger of these changes of variability over the western IO. To sum up, these
412 mid- to late Holocene changes of tropical SST are consistent with the amplification (cooling)

413 of the Pacific and Atlantic upwelling (Braconnot et al. 2012), increased ENSO variability
414 reported during the Holocene (Emile-Geay et al. 2016) and a strong control of ENSO on
415 tropical IO SST variability, especially the western IO, as seen in present-day climate (Crétat
416 et al. 2017).

417

418 **4. Changes in recurrent ISMR regimes and teleconnections**

419

420 We now go one step further by analyzing ISMR interannual-to-decadal variability no more as
421 a standard-deviation around the mean-state but through six recurrent ISMR regimes extracted
422 from a cluster analysis of Indian rainfall anomalies with respect to the mean-state estimated in
423 100-yr windows (see Appendix A for details). The ISMR regimes are mainly discussed from
424 6 to 3.8 kyr BP within the 5-simulation ensemble to account for uncertainties induced by the
425 model versions and setups. They are also briefly discussed for late Holocene and from mid- to
426 late Holocene within the two 6000-yr long simulations to assess model skill and long-term
427 trends in rainfall interannual-to-decadal variability, respectively.

428

429 4.1 Ensemble mean

430

431 Regimes #1 and #6 represent respectively ~14% and ~26% of the JJAS anomalous Indian
432 rainfall patterns classified from 6 to 3.8 kyr BP (Table 3; 1st column). They both describe a
433 weak meridional dipole in rainfall anomalies over India (Fig. 7a,f) associated with modest
434 SST anomalies in the tropics (Fig. 8a,f). Regime #1 consists in significant wet and dry rainfall
435 anomalies in central and south India, respectively (Fig. 7a). Regime #6 depicts a similar
436 dipole structure but slightly shifted northward 6 (Fig. 7f: wet/dry anomalies north/south of
437 ~15°N). Regime #1 is associated with warm tropical Pacific SST anomalies and cold North

438 subtropical SST anomalies in the three basins (Fig. 8a). This weakens the meridional
439 temperature gradient, hence limiting westerly wind anomalies over the Arabian Sea (Fig. 9a).
440 On the other hand, the increased meridional temperature gradient in regime #6 strengthens
441 westerly wind anomalies over Africa and North Africa and favors a northwestward shift of the
442 monsoon trough over India (Figs. 8f and 9f). Interestingly, the whole circulation pattern of
443 regime #6 is very similar to the one obtained by reducing the albedo over Sahara and Pakistan
444 arid regions in sensitivity coupled experiments (Sooraj et al. 2019).

445
446 The four remaining ISMR regimes are all associated with strong rainfall anomalies (Fig. 7b-
447 e), ENSO-like SST anomalies over the tropical Pacific (Fig. 8b-e) and significant modulations
448 of the monsoon circulation in the Indian sector (Fig. 9b-e). Despite weaker tropical Pacific
449 variability in mid- rather than late Holocene (Emile-Geay et al. 2016; White et al. 2018; Chen
450 et al. 2019), tropical Pacific SST variability remains the main driver of ISMR variability
451 during the mid-Holocene in our 5-simulation ensemble. Regimes #2 and #3 both represent
452 ~15% of the classified patterns from 6 to 3.8 kyr BP (Table 3; 1st column). They describe
453 significant wet anomalies over most or all India during La Niña-like conditions (Figs. 7b-c
454 and 8b-c). These wet anomalies are due to moisture convergence favored by significant
455 easterly wind anomalies over the western Pacific and eastern IO and an anticyclonic center
456 over the Bay of Bengal (Fig. 9b-c). However, regime #3 only is associated with a clear Gill-
457 type response to convection and latent heat release over India, with strengthened 850-hPa
458 southwesterlies over the Arabian Sea (Fig. 9c). On the other hand, regimes #4 and #5 (20%
459 and 10% of the classified patterns, respectively: Table 3; 1st column) describe either
460 widespread dryness or a north-south dipole over India under El Niño-like conditions (Figs.
461 7d-e and 8d-e). These abnormally dry conditions are systematically associated with 850-hPa
462 westerly wind anomalies over the eastern IO–western Pacific (north of 10°N), reduced

463 southwesterly monsoon fluxes over the Arabian Sea and a split of the monsoon trough over
464 India (Fig. 9d-e), which is typical of drought conditions during the monsoon (Terray et al.
465 2005).

466

467 Interestingly, the two ISMR regimes associated with widespread wet/dry rainfall anomalies of
468 same sign over all India (i.e., regimes #3 and #4: Fig. 7c-d) occur under well-defined ENSO-
469 like – IOD collocation (Fig. 8c-d). Wet conditions over all India (Fig. 7c) are favored when
470 these two coupled modes of variability collocate during their negative phases (i.e., La Niña-
471 like and negative IOD: Fig. 8c). The reverse prevails for regime #4 (Figs. 7d and 8d).
472 Regimes #3 and #4 reflect the strong control of ENSO-like variability on both IOD and ISMR
473 variability during mid-Holocene. This does not preclude positive/negative IODs to promote
474 wet/dry ISMR at times, as suggested by e.g. Ashok et al. (2001 and 2004) but this is clearly
475 not the dominant signal found here at the interannual-to-decadal timescale.

476

477 To verify whether or not these results are consistent with regimes obtained for modern ISMR
478 variability, the clustering has been applied to the last 100 years of the TR5AS-V1r01 and
479 TR6AV-Sr02 ensemble (i.e., 1851-1950) and to 1901-2016 observations (CRU data for
480 rainfall and HadISST data [Rayner et al. 2003] for SST). Despite the 6 ISMR regime patterns
481 simulated for the last 100 years are not strictly similar to those derived from 6 to 3.8 kyr BP,
482 the main modes affecting ISMR variability persist (Fig. 10). This demonstrates that ENSO
483 remains the main driver of interannual-to-decadal ISMR variability for the last 6000 years
484 regardless of the strength of orbital and trace gas forcing. The comparison with CRU-
485 HadISST data reveals that the simulations reasonably represent the observed ISMR regimes
486 and associated teleconnections, despite distorted ENSO patterns and exaggerated coupling
487 with IO and Atlantic SST variability. The only exception concerns the wet ISMR regime

488 associated to the strong IO basin-wide warming following strong El Niños (especially the
489 1982-83 and 1997/98) and the fast El Niño to La Niña transitions (Fig. 10), which represent
490 the “best conditions” for a strong monsoon during recent decades (Boschat et al. 2011, 2012).
491 The fact that this mode is not emerging in the simulations is probably linked to the inability of
492 the model to produce a realistic negative IO feedback on ENSO, a bias shared by many
493 current climate models, especially those with a significant cold tongue bias in the equatorial
494 Pacific like the IPSL model (Kug and Ham 2012; Li et al. 2019; Terray et al. 2020).

495

496 4.2 Model version dependency

497

498 The model-version dependency of ISMR regimes is assessed in terms of anomalous rainfall
499 patterns (Fig. 11) and total number of regime occurrence (Table 3). These analyses confirm
500 that TR5AS-Vlr01 is an outlier, with largest departure from the ensemble mean (Fig. 11a) and
501 underestimated occurrence of regimes #3 and #5 (Table 3). This suggests that ENSO and its
502 teleconnection with ISMR and IO are much less diverse in this simulation compared to the
503 others.

504

505 Overall, the ISMR regimes are very similar between the four TR6A simulations in terms of
506 mean rainfall anomalies (Fig. 11b-e) and population (Table 3). We note, however, that the
507 simulation set with fixed 1850 vegetation map (i.e., TR6AS-Sr10: Fig. 11b) deviates slightly
508 more from the ensemble mean than those set with fixed mid-Holocene vegetation (i.e.,
509 TR6AS-Sr12: Fig. 11d) or dynamical vegetation (i.e., TR6AV-Sr02: Fig. 11e). Braconnot et
510 al. (2019a) showed that bare soil and grass increase over India from mid- to late Holocene at
511 the expense of forest (see their Fig. 10). Imposing more bare soil and grass and less forest
512 over India during the mid-Holocene in TR6AS-Sr10 compared to TR6AS-Sr12 and TR6AV-

513 Sr02 may thus affect vegetation-climate feedbacks through the modulation of albedo and local
514 recycling.

515

516 4.3 Evolution of the ISMR regimes and associated teleconnections

517

518 The frequency of occurrence of the ISMR regimes, computed every adjacent 100-yr windows,
519 is marked by stochastic centennial-to-multi-centennial fluctuations (Fig. 12). Cross-
520 correlations between each ISMR regime are very noisy regardless of the metric. In particular,
521 the frequency of occurrence of regime #3 is poorly correlated with that of regime #4 ($r \sim |0.1|$
522 in most simulations). This indicates that centuries with changes in the occurrence of wet
523 rainfall anomalies associated with La Niña-like – IOD- collocation are not necessarily
524 accompanied by changes in the occurrence of dry anomalies associated to El Niño-like –
525 IOD+ collocation. These results are consistent with the trends for ISMR wet and dry years
526 displayed in Fig. 4d-e.

527

528 On the other, none of the ISMR regimes depict a clear orbitally-driven trend from 6 to 3.8 kyr
529 BP shared by the five simulations (Fig. 12). This contrasts with boreal summer SST
530 variability in the Niño3.4, western tropical IO and eastern equatorial IO, which increases for
531 the two former regions and decreases for the latter region in most simulations (Fig. 13a-c).
532 The variability of the IOD, however, does not depict significant change (Fig. 13d) from 6 to
533 3.8 kyr BP. A first explanation of the absence of orbitally-driven changes in the ISMR
534 regimes from 6 to 3.8 kyr BP could involve interferences with low-frequency modes of
535 internal variability. However, this hypothesis is rejected since the AHC procedure applied to
536 2-20-yr filtered rainfall anomalies leads to similar results. Another explanation could be that
537 6–3.8 kyr BP changes in ENSO-like variability may be too weak to influence ISMR regimes

538 or, alternatively, that IOD variability may counteract ENSO's effects on ISMR, as suggested
539 for the historical period (Ashok and Saji 2007).

540

541 When considering the last 6000 years, the ISMR regimes strongly connected to ENSO-like –
542 IOD collocation become significantly more frequent from mid- to late Holocene in TR6AV-
543 Sr02 (Fig. 14). This trend concerns both wet and dry regimes associated to ENSO – IOD
544 collocation in their negative and positive phases, respectively (Fig. 14), while only the dry
545 regime associated to El Niño-like – IOD+ collocation in TR5AS-Vlr01 (not shown). We thus
546 make the hypothesis that the 6–3.8 kyr BP period is too short to detect changes in rainfall
547 interannual-to-decadal variability, but that other variables could be better precursors of its
548 long-term changes.

549

550 We thus examined changes in climate anomalies associated to the two ISMR regimes linked
551 to ENSO – IOD collocation between the first and second half of the 6–3.8 kyr BP period (Fig.
552 15). In both cases, we identified modest but significant changes in precipitable water over
553 India linked to changes in both local evaporation and moisture advection from the western IO
554 and southeast Asia. These changes are associated with strengthened tropical SST anomalies,
555 particularly over the Pacific under El Niño conditions, suggesting increased control of ENSO
556 on the Indian monsoon thermodynamics from 6 to 3.8 kyr BP. Precipitable water is thus a
557 good precursor of changes in the interannual-to-decadal variability of the water cycle due to
558 the Clausius-Clapeyron relationship.

559

560

561 **5. Discussion and conclusion**

562

563 Results of the five transient simulations with the IPSL model show that the last 6000 years are
564 marked by a significant drying trend in ISMR, consistent with previous studies (e.g., Bartlein
565 et al. 2011; Dallmeyer et al. 2015; Braconnot et al. 2019b). This drying trend affects the
566 whole ISMR distribution, including the intensity and number of the 10% driest and 10%
567 wettest seasons and is associated to a contraction of the ISMR season (Figs. 3-4). The
568 magnitude of the drying trend is larger in simulations set with higher resolution and the new
569 surface hydrology. Both the way runoff is routed to ocean and vegetation-climate feedbacks
570 modulate the onset phase of the Indian monsoon (Fig. 3), but have modest effects at the
571 seasonal timescale, including the time evolution of the number and intensity in flood and
572 drought years (Fig. 4).

573

574 The interannual-to-decadal variability of JJAS rainfall significantly decreases over most of
575 India from mid- to late Holocene but increases over central India and adjacent oceanic regions
576 (Fig. 5). This variability is then decomposed into 6 physically consistent rainfall regimes to
577 investigate changes in the ISMR teleconnections.

578 • From 6 to 3.8 kyr BP, the regimes are very similar between the 4 TR6A simulations
579 but depict less diversity in the TR5A simulation (Table 3). In all simulations, the two most
580 energetic regimes in terms of rainfall intensity describe widespread dry/wet rainfall anomalies
581 over India when ENSO and IOD collocate in their positive/negative phase (Figs. 7-8). This
582 suggests that the ENSO forcing is dominant on both ISMR and IOD. The century-to-century
583 frequency of occurrence of these two regimes is poorly correlated (Fig. 12), indicating that the
584 relationship between ISMR and the two phases of ENSO-like – IOD is strongly skewed from
585 6 to 3.8 kyr BP at the centennial timescale. Furthermore, the frequency of occurrence of these
586 two ISMR regimes does not change from 6 to 3.8 kyr BP, contrary to the intensity of
587 precipitable water (Fig. 15) identified as a good parameter to detect emerging trends in

588 monsoon variability. Rainfall appears to be the last link in the chain and is mediated by
589 dynamic changes, which may have different and remote origins. This leads us to conclude that
590 a threshold needs to be reached prior to see significant trends in rainfall interannual-to-
591 decadal variability.

592 • From 6 to 0 kyr BP, these two ISMR regimes become more frequent from 6 to 0 kyr
593 BP (Fig. 14) as ENSO amplitude strengthens in our modeling framework (Fig. 6), consistent
594 with increased ENSO variability as inferred by paleo reconstructions (Cobb et al. 2013; Carré
595 et al. 2014; Emile-Geay et al. 2016). This suggests a reorganization of ISMR (and IOD)
596 interannual-to-decadal variability from mid- to late Holocene, with increasing influence of
597 ENSO and associated changes in the Walker circulation on both ISMR and IOD. On the other
598 hand, the slow evolution of interannual-to-decadal variability of JJAS Indian rainfall is never
599 linked to IOD only in our modeling framework, consistent with ENSO-removed experiments
600 (Crétat et al. 2017). This does not preclude the existence of IOD forcing on the Indian
601 monsoon in some specific years or periods when ENSO is at rest. It would be worth to
602 examine the ISMR–IOD relationship at different timescales and for different time slices with
603 partially coupled simulations (e.g. without ENSO) to refine these results.

604

605 We also highlight that ISMR mean-state and interannual-to-decadal variability exhibit
606 significant modulations at the multidecadal-to-multicentennial timescales that are not
607 synchronous between the simulations (Figs. 4 and 12). This is consistent with previous
608 modeling findings (e.g., Dalmeyer et al. 2015) and paleo reconstructions over India (Kathayat
609 et al. 2016; Kaushal et al. 2018). Long-term droughts such as the multidecadal-to-centennial
610 4.2 kyr BP event (Lézine et al. 2017; Yan and Liu 2019) may thus be caused by internal
611 variability since solar cycles are not imposed in our simulations. Understanding their
612 triggering mechanisms would lead to a strong improvement of ISMR predictability.

613

614 Finally, this study provides interesting tracks to model-data comparisons in the Indian sector.
615 Mean-state and variability should be examined separately since their evolution can be
616 opposite from mid- to late Holocene and their forcing mechanisms may differ. This also
617 stands for the IO, which displays a positive-to-negative IOD-like SST mean-state, but
618 increased (decreased) interannual-to-decadal variability in its western tropical (eastern
619 equatorial) part from mid- to late Holocene (Fig. 6). Our results strongly suggest that this
620 enhanced SST variability in the western IO is remotely forced by ENSO changes,
621 independently of IOD or local mean-state changes. This complexity among timescales
622 requires more investigations and may explain the disagreement among paleo-proxies for IOD
623 evolution (Kwiatkowski et al. 2015 versus Abram et al. 2007). Diagnosing changes or specific
624 events in paleoclimate records remains, however, a great challenge since there is only a small
625 probability for “big events” resulting from multiscale forcing and internal variability to be
626 synchronous between paleo-reconstructions and simulations. Strategies based on analog are
627 thus necessary to foster model-data comparisons and to utilize models for assessing the
628 mechanisms driving such compound events.

629 **Appendix A: AHC procedure**

630

631 In this study, the AHC is adopted to identify recurrent ISMR regimes within our multi-
632 configuration ensemble and analyze their temporal evolution primarily from 6 to 3.8 kyr BP
633 but also from 6 to 0 kyr BP and for the 1851-1950 period. The AHC procedure consists in
634 merging N anomalous JJAS rainfall patterns into N -to-1 clusters of size 1-to- N according to
635 their spatial similarity. The similarity between all pairs of patterns is measured using the
636 Euclidean distance. Each pattern is merged with another according to the smallest pairwise
637 Euclidean distance. The resulting merged cluster is then paired with another pattern or another
638 cluster according to the Ward algorithm (Ward 1963), which minimizes intra-cluster variance
639 and maximizes inter-cluster variance at each merging step. The procedure continues until all
640 patterns are merged in a unique cluster. For a more detailed explanation of the AHC
641 procedure we refer the reader to Cr tat et al. (2019).

642

643 The AHC is applied to a time-space matrix constructed in three successive steps. Anomalous
644 JJAS rainfall patterns are computed over India (i.e., 58 land points within 5 -25 N and 77 -
645 88 E) for each simulation and each of the 22 100-yr windows from 6 to 3.8 kyr BP. For each
646 100-yr window, the 100 anomalous rainfall patterns are defined as the departure from the
647 corresponding 100-yr rainfall mean-state pattern. The anomalous rainfall patterns from the
648 TR5AS-Vlr01 simulation are interpolated onto the horizontal resolution of the TR6A
649 simulations (see Table 1) using a bilinear algorithm. The anomalous patterns from the five
650 simulations are concatenated in a time-space matrix. A total of 10980 patterns are classified
651 from 6 to 3.8 kyr BP instead of 11000 (2200 JJAS seasons x 5 simulations) because two
652 simulations, TR6AS-Sr10 and TR6AS-Sr12, have 10 years missing due to technical issues
653 with the data storage. The same methodology is applied from 6 to 0 kyr BP and for the 1851-

654 1950 periods for the ensemble of the two 6000-yr simulations.

655

656 Figure A1 shows the dendrogram tree obtained at the end of the AHC procedure applied to
657 the 5-simulation ensemble from 6 to 3.8 kyr BP. The dendrogram summarizes all merging
658 from N to 1 clusters of size 1 to N (with $N=10980$ being the total number of anomalous
659 rainfall patterns feeding the AHC), as well as the evolution of the similarity metric for the last
660 10 merging. The evolution of the similarity metric indicates three main abrupt jumps: from 6
661 to 5 clusters, from 3 to 2 clusters and from 2 to 1 clusters. A 2-cluster cutoff would mix ISMR
662 regimes with opposite rainfall and SST anomalies (compare regimes #1 and #6 in Figs. 7-8).
663 A 3-cluster cutoff would also mix very different regimes (compare regimes #1 with regimes
664 #2-3 in Figs. 7-8) and would merge regimes related to ENSO-like conditions with regimes
665 related to ENSO-like – IOD collocation (compare regimes #2 and #3 and regimes #4 and #5
666 in Fig. 8). Here, we retain a 6-cluster cutoff to avoid such a mixing, which has no physical
667 consistency, and because we aim at discussing internal variability and orbitally-forced
668 evolution of only physically consistent regimes. This high-degree of refinement is required
669 here since changes in ISMR variability remain subtle from mid- to late Holocene (Braconnot
670 et al. 2019b).

671

672 The statistical robustness of a 6-regime cutoff is assessed by examining the spatial correlation
673 between each anomalous rainfall pattern within a regime and its center of gravity defined as
674 the ensemble mean anomalous rainfall pattern of the regime. When this spatial correlation is
675 below 0.2, the patterns are considered as outliers. The 0.2 threshold is deliberately low to
676 avoid excluding patterns which are close to the center of gravity but slightly shifted in space.
677 About 20% of outliers are detected. They are almost equally distributed within the five
678 simulations. These outliers are removed from all the analyses. Once the outliers removed, the

679 6 ISMR regimes synthesize 8963 patterns. The spatial correlation values obtained between
680 each pattern and its corresponding outlier-free center of gravity exceed +0.35 more than 75%
681 of the time and the median correlation reaches at least +0.5 whatever the regime.

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683

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691

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917 hemispheric differences in the responses to orbital forcing and ocean feedbacks. *Clim*

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920 **Table Captions**

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923 **Table 1:** Major differences in model set up between the five simulations analyzed in this
924 study. All simulations have the ORCA 2° ocean grid with 31 vertical levels. The LR and MR
925 resolutions stand for the atmospheric horizontal resolution. The LR (MR) atmosphere
926 resolution is 3.75° (2.5°) in longitude and 1.875° (1.25°) in latitude with 39 vertical levels in
927 both LR and MR. See main text for the description of the model version, the land surface, the
928 land-ocean river runoff and the experimental design for prescribing or computing vegetation.

929

930 **Table 2:** Mean-state and variability of July-to-September seasonal ISMR (land points in the
931 5°-25°N – 77°-88°E region) for the 1871-2012 AIR, 1979-2010 GPCP and 1901-2016 CRU
932 data and for the last 100 years of the two simulations (1851-1950). Variability is computed as
933 the standard-deviation of JJAS rainfall anomalies. Values in parenthesis correspond to the
934 standard-deviation of the results obtained for the 123, 97 and 81 20-yr windows of the AIR,
935 CRU and two simulations.

936

937 **Table 3:** Number of patterns (and percentage) within each of the six ISMR regimes after
938 removing the outliers. The percentage in the first column is expressed against the total
939 number of classified patterns. The percentage in the remaining columns corresponds to the
940 weight each simulation has on the total number of patterns spent in each ISMR regime (given
941 in the first column).

942

943 **Figure Captions**

944

945 **Figure 1:** (a) Mean seasonal cycle of monthly rainfall area-averaged over India (land points
946 within the 5°-25°N – 70°-88°E region) for the 1871-2012 AIR and 1901-2016 CRU and
947 1979-2010 GPCP data, and the last 100 years of the TR5AS-Vlr01 and TR6AV-Sr02
948 simulations (i.e., 1851-1950). For the AIR data and the two simulations, thin curves show
949 sampling errors defined as the mean seasonal cycle of all possible 20-yr windows along the
950 1871-2012 and 1851-1950 period, respectively. (b) Same as (a) but for Indian rainfall
951 variability defined as the standard-deviation of rainfall anomalies.

952

953 **Figure 2:** Location of the 2 mm/day isohyet from April to September for the (a) 1979-2010
954 GPCP and (b) 1901-2016 CRU data and the last 100 years (i.e., 1851-1950) of the (c)
955 TR5AS-Vlr01 and (d) TR6AV-Sr02 simulations.

956

957 **Figure 3:** Mid- to late Holocene changes in the (a-b) mean-state and (c-d) variability of
958 rainfall area-averaged over India for the TR5AS-Vlr01 and TR6AV-Sr02 simulations. Both
959 mean-state and variability are computed along 100-yr adjacent windows. (e-h) Relative
960 differences in rainfall mean-state between the four TR6 simulations and their ensemble mean
961 for each 100-yr adjacent from 6 to 3.8 kyr BP.

962

963 **Figure 4:** (a-b) PDF of JJAS rainfall area-averaged over India computed along each 100-yr
964 adjacent window for the TR5AS-Vlr01 and TR6AV-Sr02 simulations, respectively. Blue-to-
965 red colors show the evolution of the PDFs from mid- to late Holocene. (c) Evolution of the
966 10th and 90th percentile threshold values of ISMR anomalies for the five simulations. The
967 percentile threshold values for each simulation are computed along each 100-yr moving
968 window with 1-year increment between each window, giving one value for each year, and are
969 displayed as ISMR departures from the first 100-yr climatology to remove offset differences
970 between the simulations induced by different initial conditions. (d-e) Evolution in the number
971 of the 10% driest and 10% wettest ISMR seasons in each simulation defined using the 10th
972 and 90th percentile threshold values from the first 100-yr window.

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974 **Figure 5:** Snapshots of mid- to late Holocene significant changes in JJAS rainfall mean-state
975 (dashed/solid contours for |0.2|, |1.5| and |3| mm/day dry/wet differences) and standard-
976 deviation (shadings) for the TR6AV-Sr02 simulation. For each 100-yr adjacent window,

977 changes in mean-state and standard-deviation are computed as the departure from the first 100
978 years of the simulation (e.g. 6-5.9 kyr BP period) and significance is tested at the 95%
979 confidence level according to a Student t test and a Chi-square test, respectively. (a) Mean
980 and standard-deviation changes (averages on 4 100-yr adjacent windows) from 5.9 to 5.5 kyr
981 BP that reach the 95% confidence level in at least 2 100-yr windows. (b-f) Same as (a) but for
982 changes averaged on 5 100-yr consecutive windows from 5 to 4.5, 4 to 3.5, 3 to 2.5, 2 to 1.5
983 and 1 to 0.5 kyr BP time interval, respectively, that reach the 95% confidence level in at least
984 3 100-yr windows in the time interval.

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986 **Figure 6:** Same as Fig. 5 but for changes in SST mean-state (dashed/solid contours for $|0.1|$,
987 $|0.4|$ and $|1|$ °C cold/warm differences) and standard-deviation (shadings).

988

989 **Figure 7:** (a) Ensemble mean anomalous rainfall pattern associated to ISMR regime #1. Only
990 rainfall anomalies that are significant at the 95% confidence level according to a Student t test
991 are shown. (c-f) Same as (a) but for the 5 remaining ISMR regimes.

992

993 **Figure 8:** Same as Fig. 8 but for SST anomalies (°C), except significant SST anomalies are
994 contoured in black.

995

996 **Figure 9:** Same as Fig. 8 but for 850-hPa wind anomalies (vectors; m/s) and associated
997 divergence (shadings; m/s; blue = anomalous convergence; red = anomalous divergence).
998 Wind anomalies are shown in black when at least one component (i.e., zonal or meridian)
999 depict significant changes tested as in Fig. 8.

1000

1001 **Figure 10:** (a) Ensemble mean rainfall and SST anomalies associated with the six ISMR
1002 regimes identified by applying the AHC clustering to the last 100 years (i.e., 1851-1950) of
1003 the TR5AS-Vlr01 and TR6AV-Sr02 ensemble. Significant anomalies at the 95% confidence
1004 level according to a Student t test are contoured in black. (b) Same as (a) but for the CRU-
1005 HadISST data for the 1901-2016 period.

1006

1007 **Figure 11:** Model dependency in the mean anomalous rainfall pattern associated to the six
1008 ISMR regimes for (a-e) the five simulations. For each ISMR regime, model dependency is
1009 defined as the departure of the mean anomalous rainfall pattern of each simulation from the
1010 ensemble mean anomalous rainfall pattern shown in Fig. 8. Only departure significant at the

1011 95% confidence level according to a Student t test is shown.

1012

1013 **Figure 12:** (a-f) Century-to-century evolution in the frequency of occurrence of the six ISMR
1014 regimes from 6 to 3.8 kyr BP for the five simulations. The frequency of occurrence is
1015 displayed as the departure from its mean value.

1016

1017 **Figure 13:** Century-to-century evolution in boreal summer (JJAS) SST variability of the (a)
1018 Niño3.4, (b) western tropical IO (10°S-10°N; 50°-70°E), (c) eastern equatorial IO (10°S-Eq;
1019 90°-110°E) and (d) IOD indices from 6 to 3.8 kyr BP for the five simulations. Variability is
1020 defined as standard-deviation of SST anomalies for each 100-yr adjacent window. The
1021 variability is displayed as the departure from the mean value of the standard-deviations
1022 computed for each 100-yr adjacent window.

1023

1024 **Figure 14:** (a; upper panel) Century-to-century evolution in the frequency of occurrence of
1025 the ISMR regime associated to La Niña-like – IOD- collocation from 6 to 0 kyr BP for the
1026 TR6AV-Sr02 simulation. This ISMR regime is extracted by applying the AHC clustering to
1027 anomalous Indian rainfall patterns from 6 to 0 kyr BP in the two 6000-yr simulation ensemble
1028 (TR5AS-Vlr01 and TR6AV-Sr02) with a 6-regime cutoff. The frequency of occurrence is
1029 expressed as the departure from its mean value. (a; bottom panels) Number of occurrences of
1030 this ISMR regime along each 2000-yr adjacent window and associated mean anomalous SST
1031 patterns. Only SST anomalies significant at the 95% confidence level according to a Student t
1032 test are shown. (b) Same as (a) but for the ISMR regime associated to El Niño-like – IOD+
1033 collocation.

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1035 **Figure 15:** (a) 4.9-3.8 minus 6-4.9 kyr BP differences in precipitable water, SST, moisture
1036 transport and evaporation ensemble mean anomalies associated to regime #3. Significant
1037 anomalies at the 95% confidence level according to a Student two-tailed t test are contoured
1038 in black for precipitable water, SST and evaporation and shown with black vectors for
1039 moisture transport. (b) Same as (a) but for regime #4.

1040

1041 **Figure A1:** (a) Dendrogram tree obtained at the end of the AHC procedure applied to JJAS
1042 rainfall anomalies simulated over India by the 5-simulation ensemble from 6 to 3.8 kyr BP.
1043 (b) Evolution of the similarity metric for the 10 last mergings. See Appendix A for details.

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1045 **Tables**

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Simulation	Model version	Resolution	Land-surface	Dynamical vegetation	Land-ocean river runoff	Integration length
TR5AS-Vlr01	TR5A	LR	CM5	No – present	CM6	6,000 years
TR6AV-Sr02	TR6A	MR	11 layers	Yes	Inter	6,000 years
TR6AS-Sr10	TR6A	MR	11 layers	No – present	Inter	4,000 years
TR6AS-Sr11	TR6A	MR	11 layers	No – present	CM6	2,250 years
TR6AS-Sr12	TR6A	MR	11 layers	No – Mid-Holocene	Inter	2,250 years

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	Mean State (mm.day ⁻¹)	Variability (mm.day ⁻¹)
AIR	6.93 (+/- 0.15)	1.22 (+/- 0.13)
GPCP	7.36	1.13
CRU	6.73 (+/- 0.15)	1.33 (+/- 0.10)
Vlr01	4.84 (+/- 0.18)	1.05 (+/- 0.11)
Sr02	5.72 (+/- 0.15)	1.34 (+/- 0.13)

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Table 2: Mean-state and variability of July-to-September seasonal ISMR (land points in the 5°-25°N – 77°-88°E region) for the 1871-2012 AIR, 1979-2010 GPCP and 1901-2016 CRU data and for the last 100 years of the two simulations (1851-1950). Variability is computed as the standard-deviation of JJAS rainfall anomalies. Values in parenthesis correspond to the standard-deviation of the results obtained for the 123, 97 and 81 20-yr windows of the AIR, CRU and two simulations.

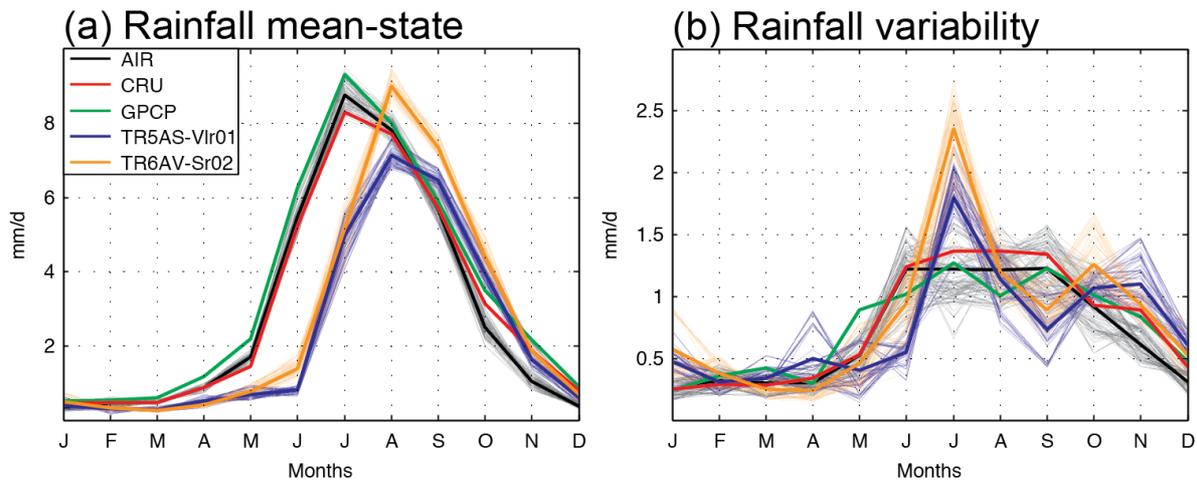
	Total	TR5AS- Vlr01	TR6AS- Sr10	TR6AS- Sr11	TR6AS- Sr12	TR6AV- Sr02
CL#1	1227 (13.7%)	333 (27.1)	199 (16.2)	221 (18)	238 (19.4)	236 (19.2)
CL#2	1344 (15%)	425 (31.6)	229 (17)	229 (17)	237 (17.6)	224 (16.7)
CL#3	1278 (14.2%)	131 (10.3)	280 (21.9)	292 (22.8)	296 (23.2)	279 (21.8)
CL#4	1821 (20.3%)	477 (26.2)	345 (18.9)	344 (18.9)	328 (18)	327 (18)
CL#5	933 (10.4%)	39 (4.2)	252 (27)	222 (23.8)	209 (22.4)	211 (22.6)
CL#6	2360 (26.3%)	288 (12.2)	489 (20.7)	518 (21.9)	521 (22.1)	544 (23.1)
Total	8963	1693	1794	1826	1829	1821

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Table 3: Number of patterns (and percentage) within each of the 6 recurrent ISMR regimes after removing the outliers. The percentage in the first column is expressed against the total number of classified patterns. The percentage in the remaining columns corresponds to the weight each simulation has on the total number of patterns spent in each ISMR regime (given in the first column).

1071 **Figures**

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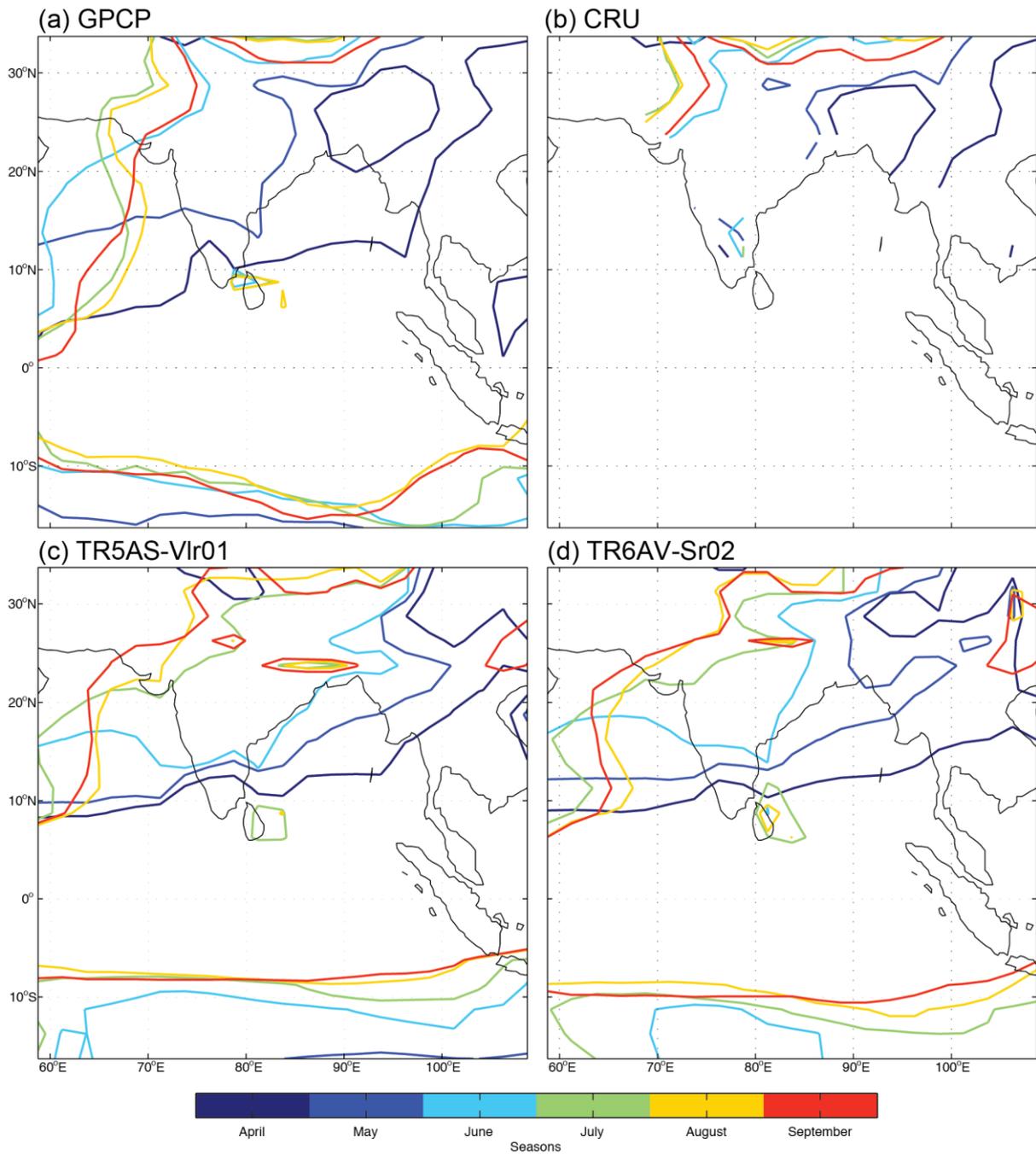


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1079 sampling errors defined as the mean seasonal cycle of all possible 20-yr windows along the
1080 1871-2012 and 1851-1950 period, respectively. (b) Same as (a) but for Indian rainfall
1081 variability defined as the standard-deviation of rainfall anomalies.

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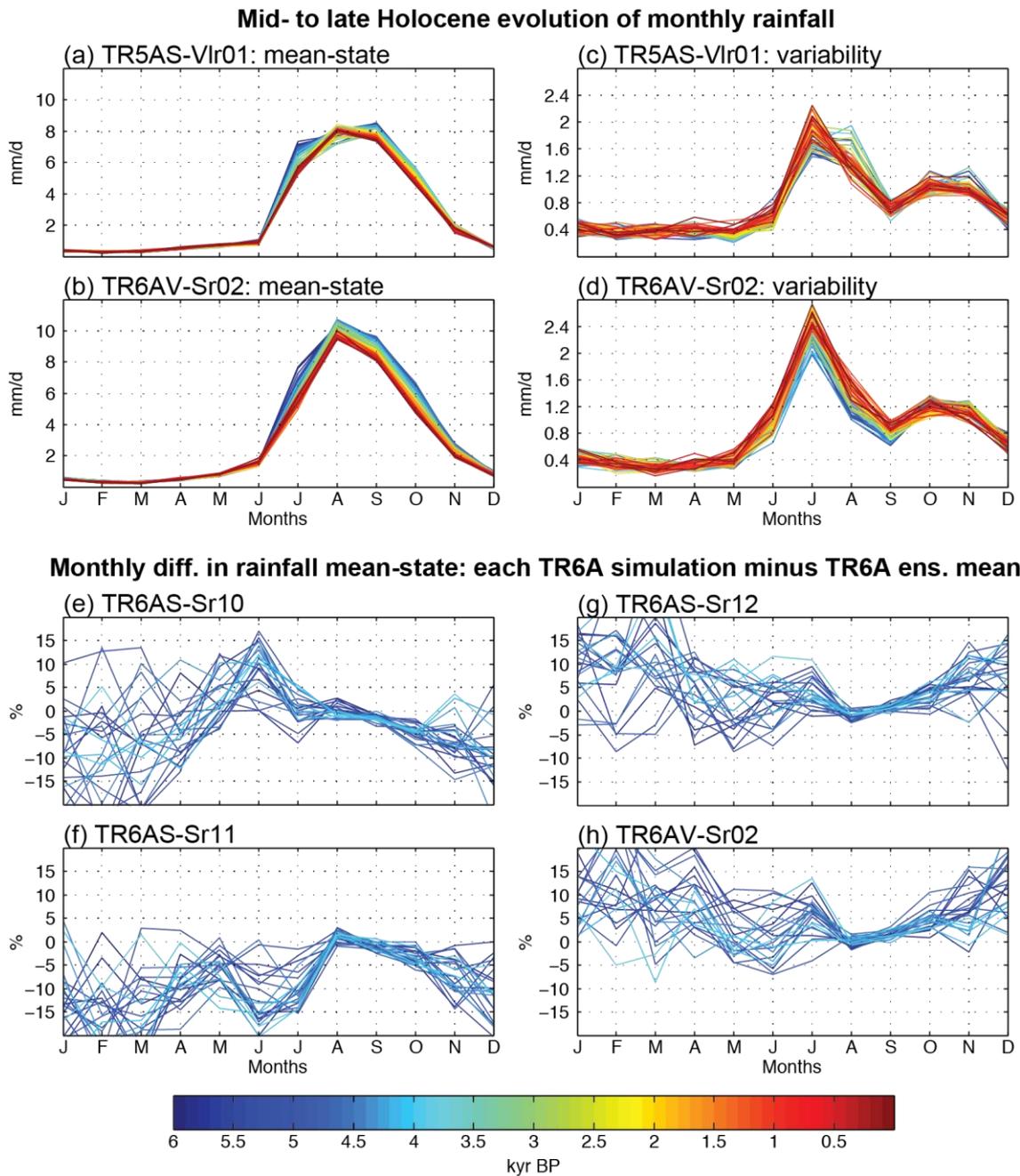


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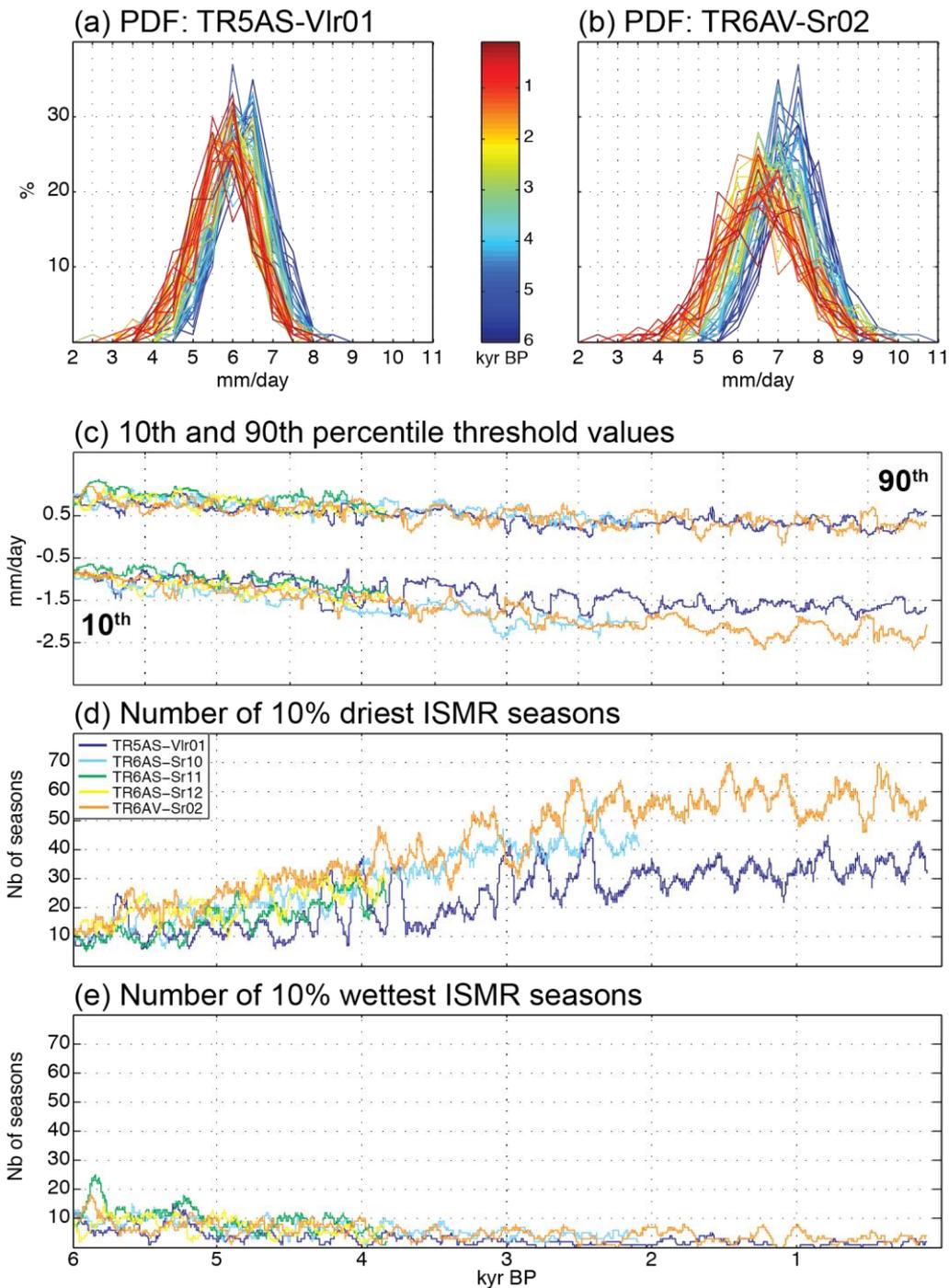
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 1094 differences in rainfall mean-state between the four TR6 simulations and their ensemble mean
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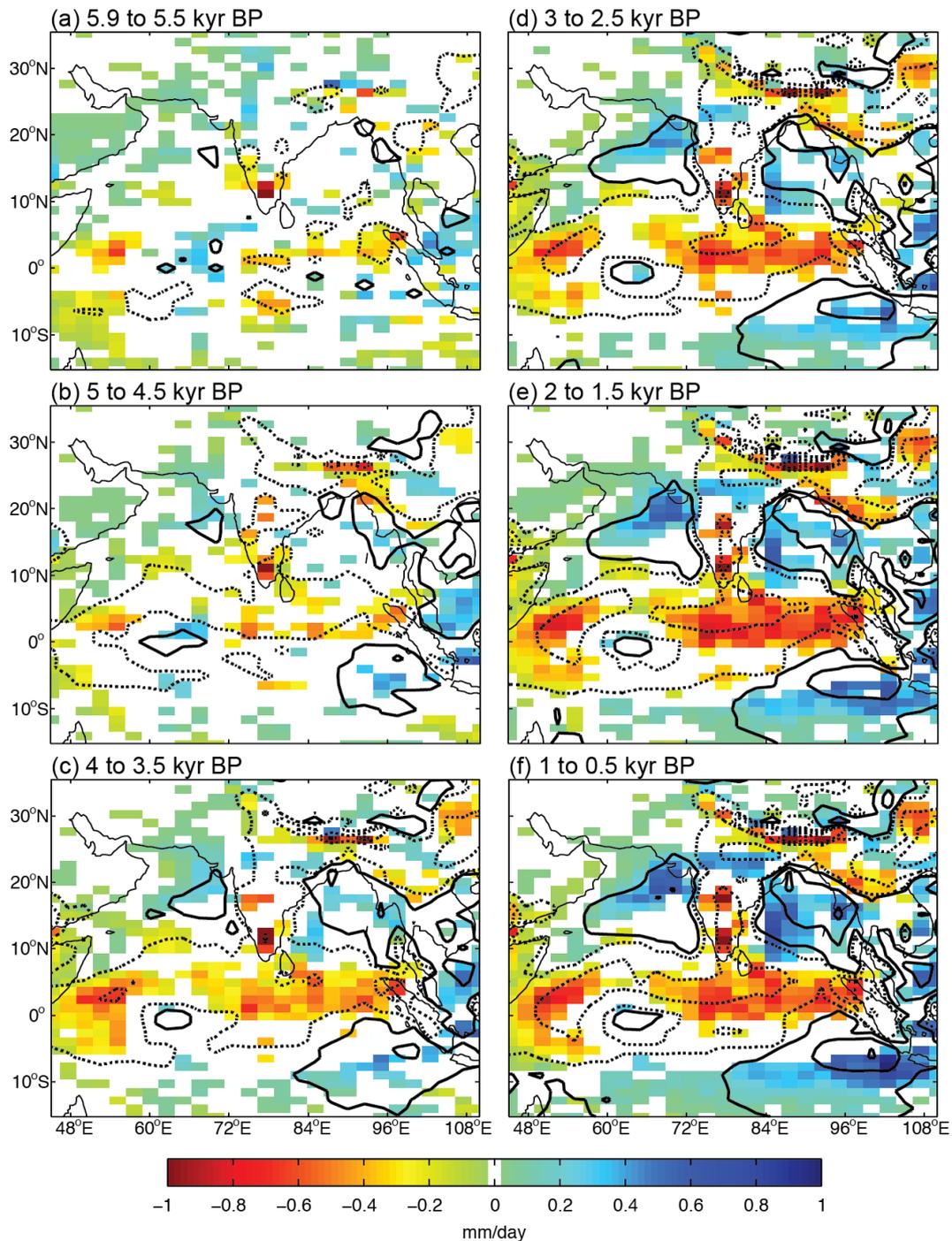
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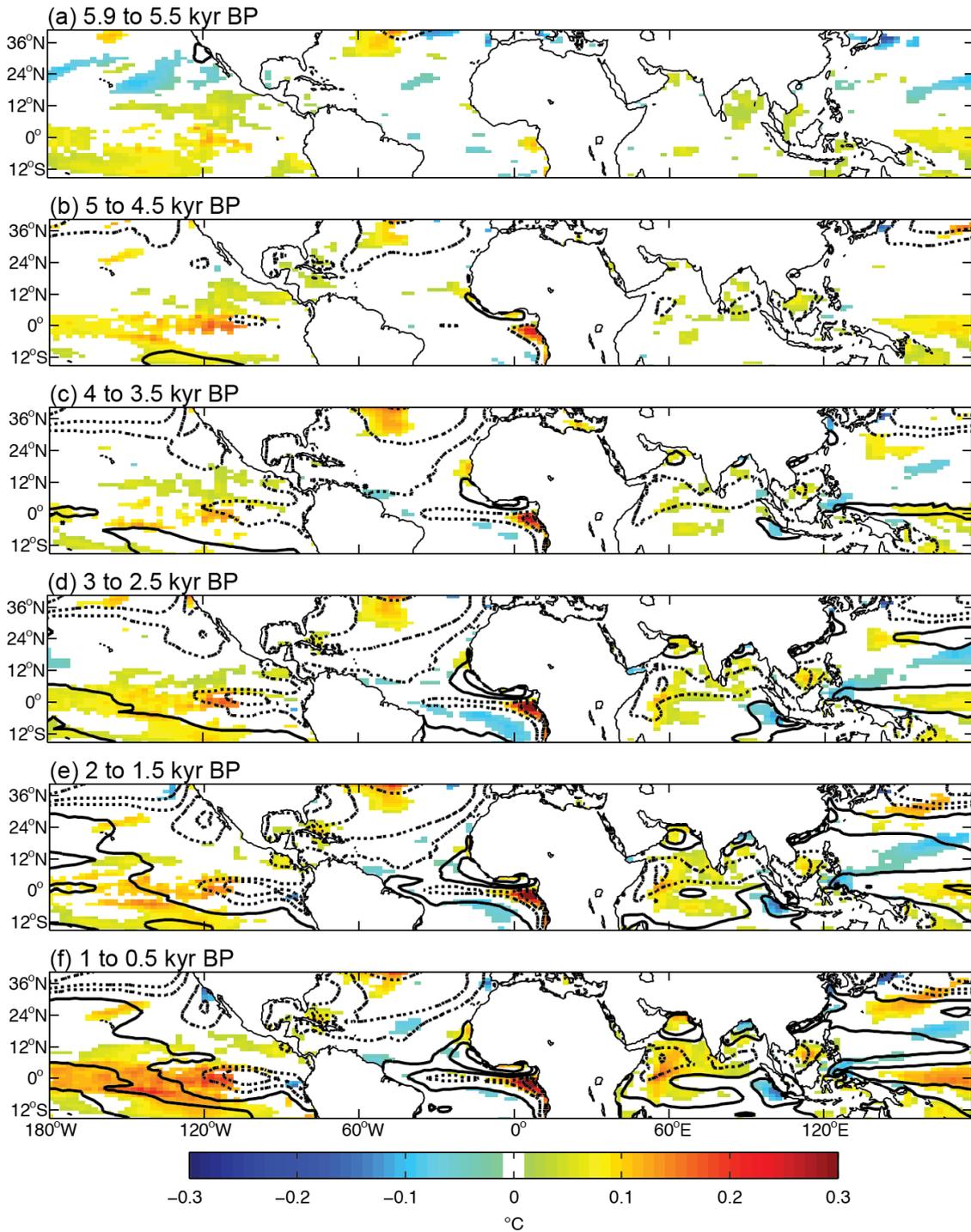
Figure 4: (a-b) PDF of JJAS rainfall area-averaged over India computed along each 100-yr adjacent window for the TR5AS-Vlr01 and TR6AV-Sr02 simulations, respectively. Blue-to-red colors show the evolution of the PDFs from mid- to late Holocene. (c) Evolution of the 10th and 90th percentile threshold values of ISMR anomalies for the five simulations. The percentile threshold values for each simulation are computed along each 100-yr moving window with 1-year increment between each window, giving one value for each year, and are displayed as ISMR departures from the first 100-yr climatology to remove offset differences between the simulations induced by different initial conditions. (d-e) Evolution in the number of the 10% driest and 10% wettest ISMR seasons in each simulation defined using the 10th and 90th percentile threshold values from the first 100-yr window.



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1110 **Figure 5:** Snapshots of mid- to late Holocene significant changes in JJAS rainfall mean-state
 1111 (dashed/solid contours for $|0.2|$, $|1.5|$ and $|3|$ mm/day dry/wet differences) and standard-
 1112 deviation (shadings) for the TR6AV-Sr02 simulation. For each 100-yr adjacent window,
 1113 changes in mean-state and standard-deviation are computed as the departure from the first 100
 1114 years of the simulation (e.g. 6-5.9 kyr BP period) and significance is tested at the 95%
 1115 confidence level according to a Student t test and a Chi-square test, respectively. (a) Mean
 1116 and standard-deviation changes (averages on 4 100-yr adjacent windows) from 5.9 to 5.5 kyr
 1117 BP that reach the 95% confidence level in at least 2 100-yr windows. (b-f) Same as (a) but for
 1118 changes averaged on 5 100-yr consecutive windows from 5 to 4.5, 4 to 3.5, 3 to 2.5, 2 to 1.5
 1119 and 1 to 0.5 kyr BP time interval, respectively, that reach the 95% confidence level in at least
 1120 3 100-yr windows in the time interval.

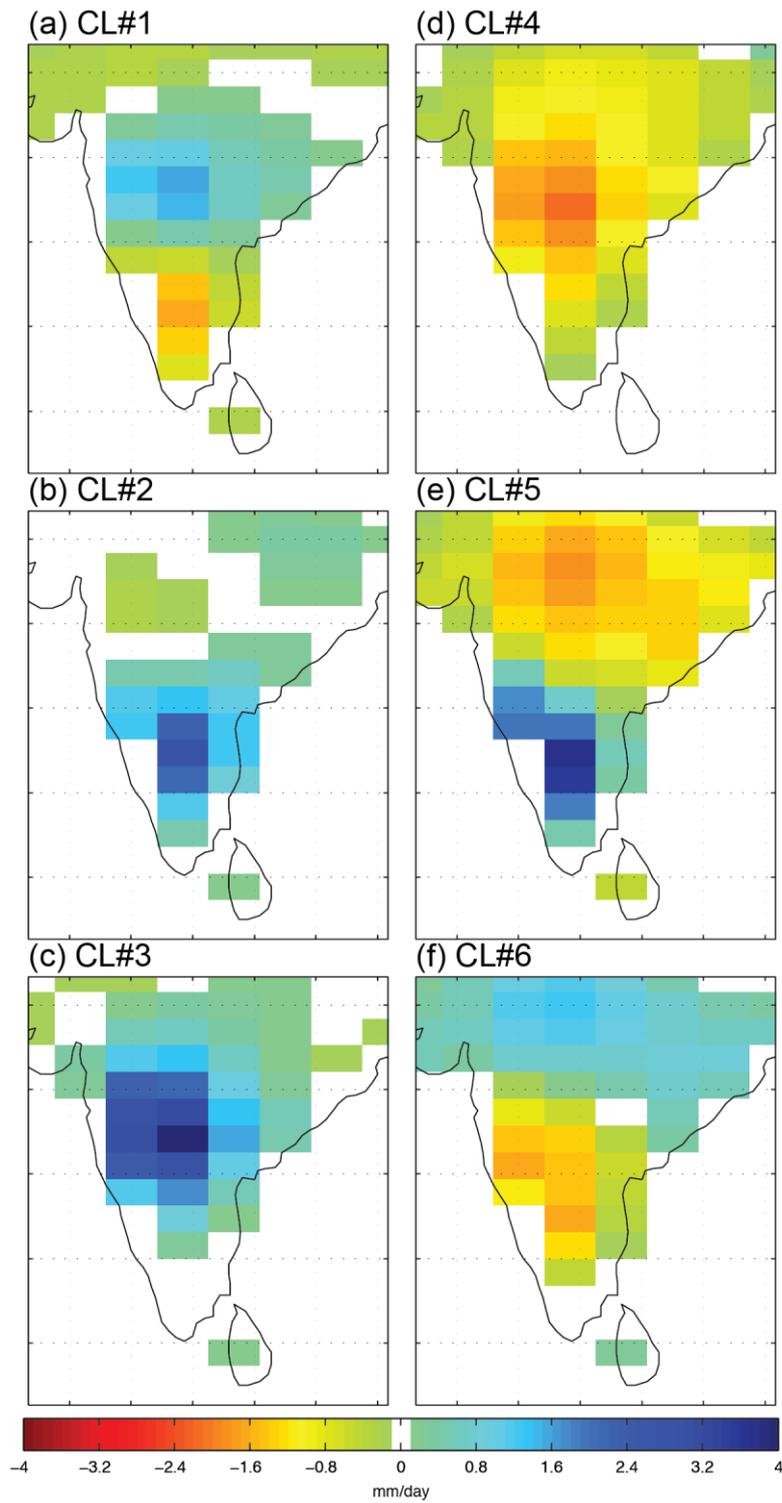


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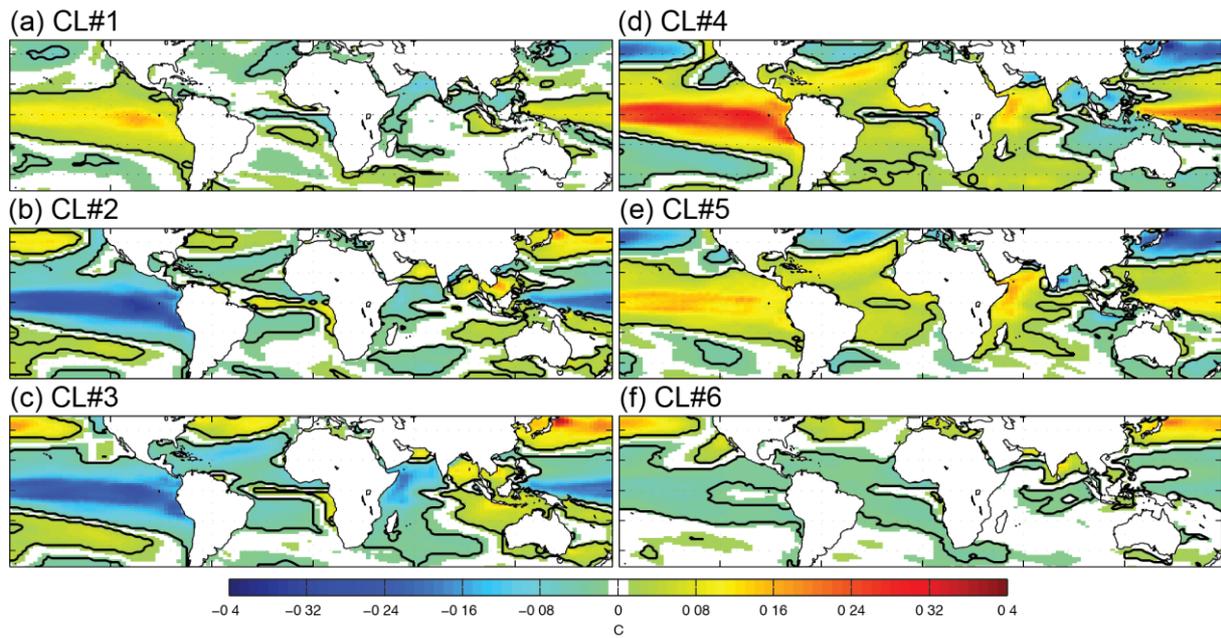
1123 **Figure 6:** Same as Fig. 5 but for changes in SST mean-state (dashed/solid contours for $|0.1|$,

1124 $|0.4|$ and $|1|$ °C cold/warm differences) and standard-deviation (shadings).



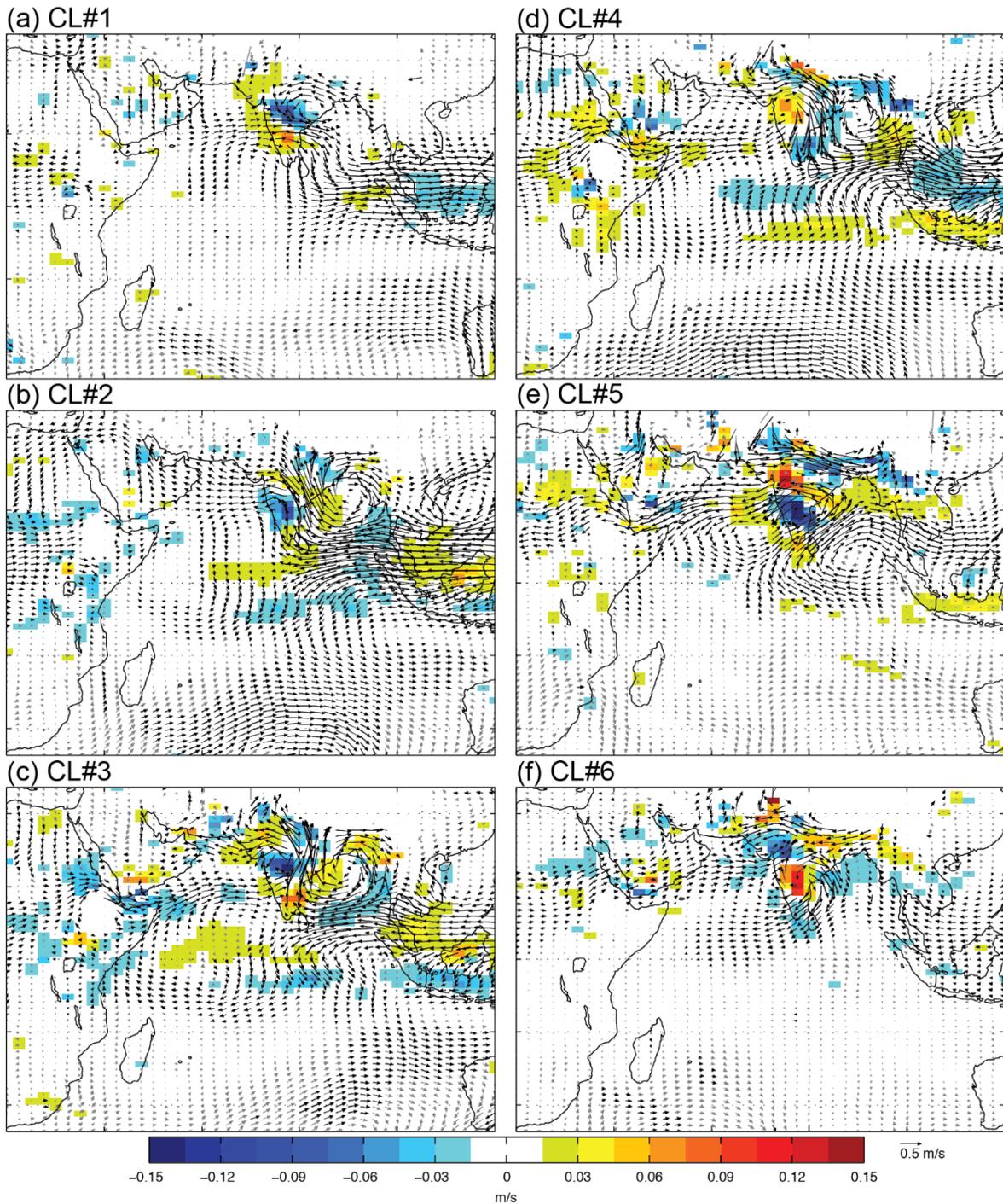
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Figure 7: (a) Ensemble mean anomalous rainfall pattern associated to ISMR regime #1. Only rainfall anomalies that are significant at the 95% confidence level according to a Student t test are shown. (c-f) Same as (a) but for the 5 remaining ISMR regimes.



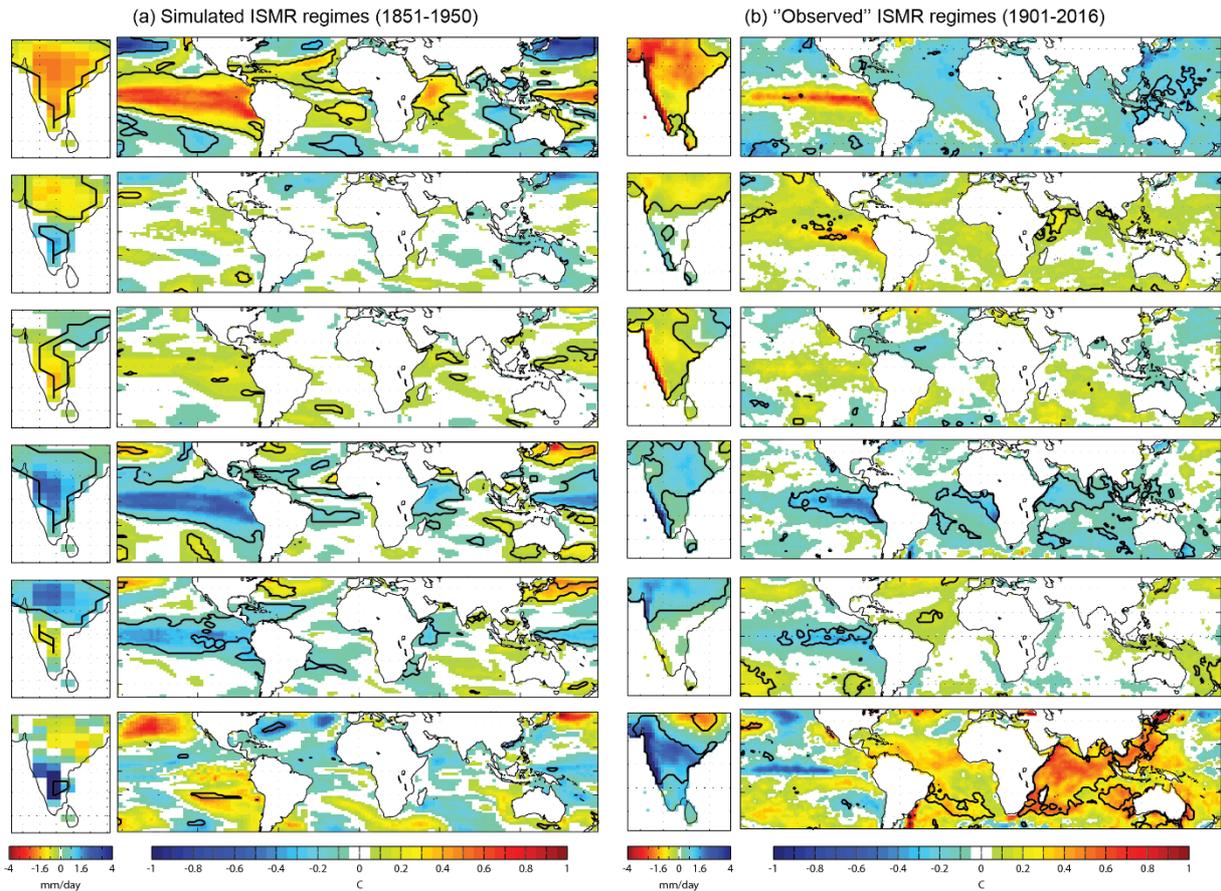
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Figure 8: Same as Fig. 7 but for SST anomalies ($^{\circ}\text{C}$), except significant SST anomalies are contoured in black.



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Figure 9: Same as Fig. 7 but for 850-hPa wind anomalies (vectors; m/s) and associated divergence (shadings; m/s; blue = anomalous convergence; red = anomalous divergence). Wind anomalies are shown in black when at least one component (i.e., zonal or meridian) depict significant changes tested as in Fig. 7.



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1145 **Figure 10:** (a) Ensemble mean rainfall and SST anomalies associated with the six ISMR
 1146 regimes identified by applying the AHC clustering to the last 100 years (i.e., 1851-1950) of
 1147 the TR5AS-Vlr01 and TR6AV-Sr02 ensemble. Significant anomalies at the 95% confidence
 1148 level according to a Student t test are contoured in black. (b) Same as (a) but for the CRU-
 1149 HadISST data for the 1901-2016 period.

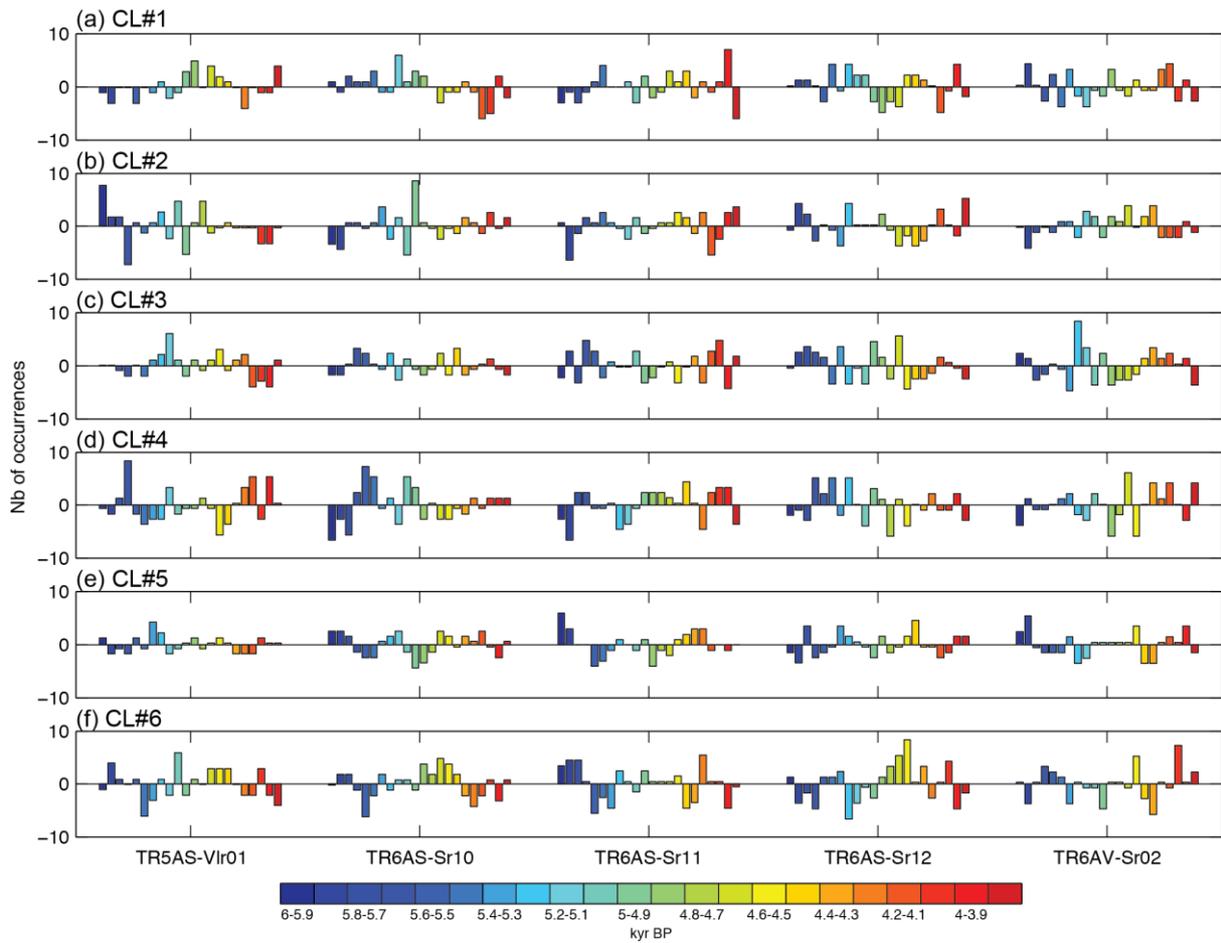


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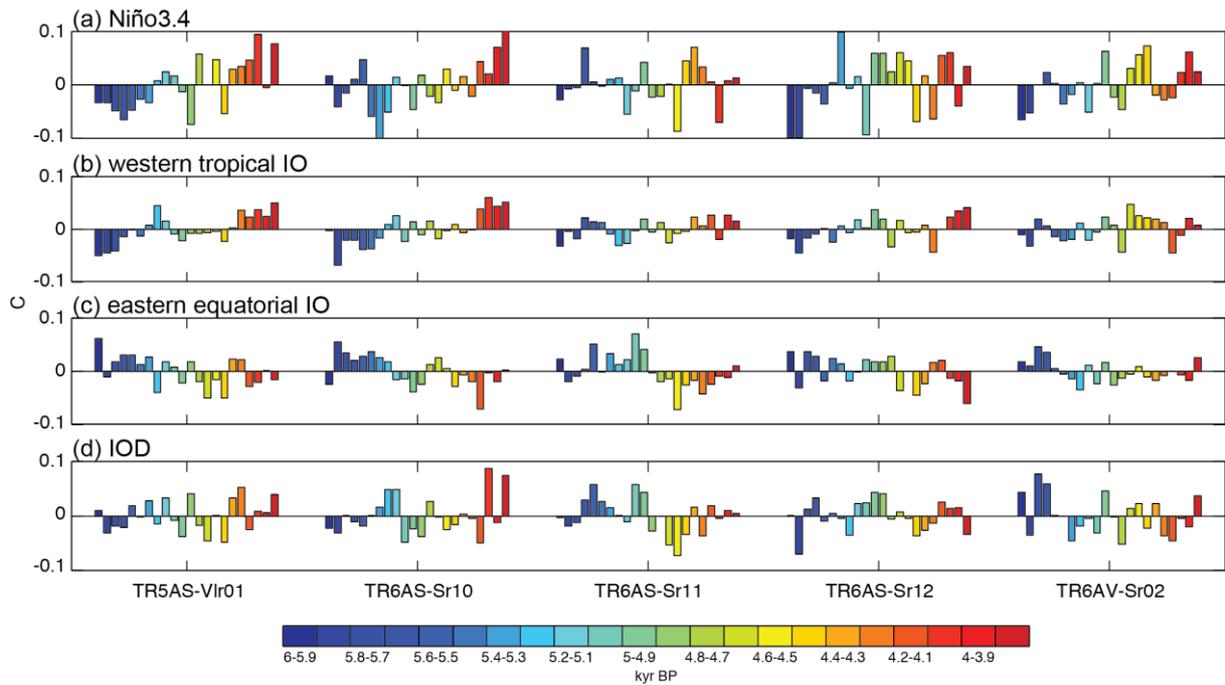
1152 **Figure 11:** Model dependency in the mean anomalous rainfall pattern associated to the six
 1153 ISMR regimes for (a-e) the five simulations. For each ISMR regime, model dependency is
 1154 defined as the departure of the mean anomalous rainfall pattern of each simulation from the
 1155 ensemble mean anomalous rainfall pattern shown in Fig. 7. Only departure significant at the
 1156 95% confidence level according to a Student t test is shown.

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Figure 12: (a-f) Century-to-century evolution in the frequency of occurrence of the six ISMR regimes from 6 to 3.8 kyr BP for the five simulations. The frequency of occurrence is expressed as the departure from its mean value.



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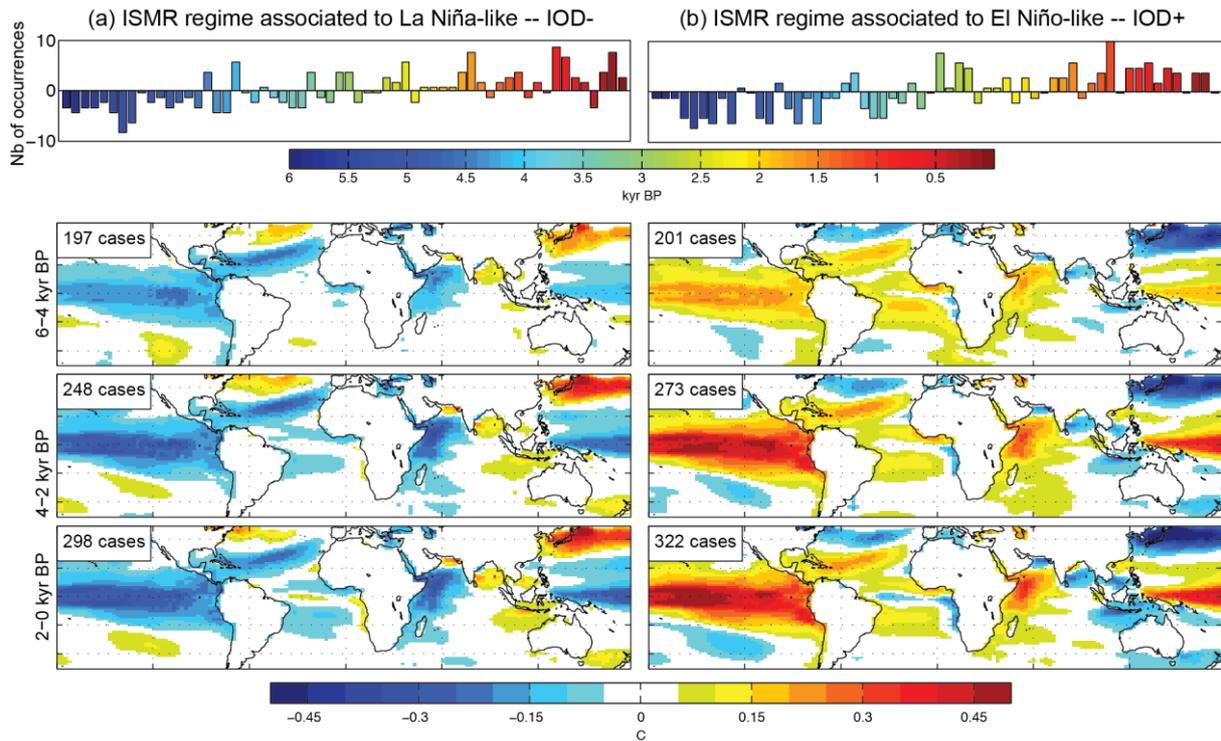
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Figure 13: Century-to-century evolution in boreal summer (JJAS) SST variability of the (a) Niño3.4, (b) western tropical IO (10°S-10°N; 50°-70°E), (c) eastern equatorial IO (10°S-Eq; 90°-110°E) and (d) IOD indices from 6 to 3.8 kyr BP for the five simulations. Variability is defined as standard-deviation of SST anomalies for each 100-yr adjacent window. The variability is displayed as the departure from the mean value of the standard-deviations computed for each 100-yr adjacent window.



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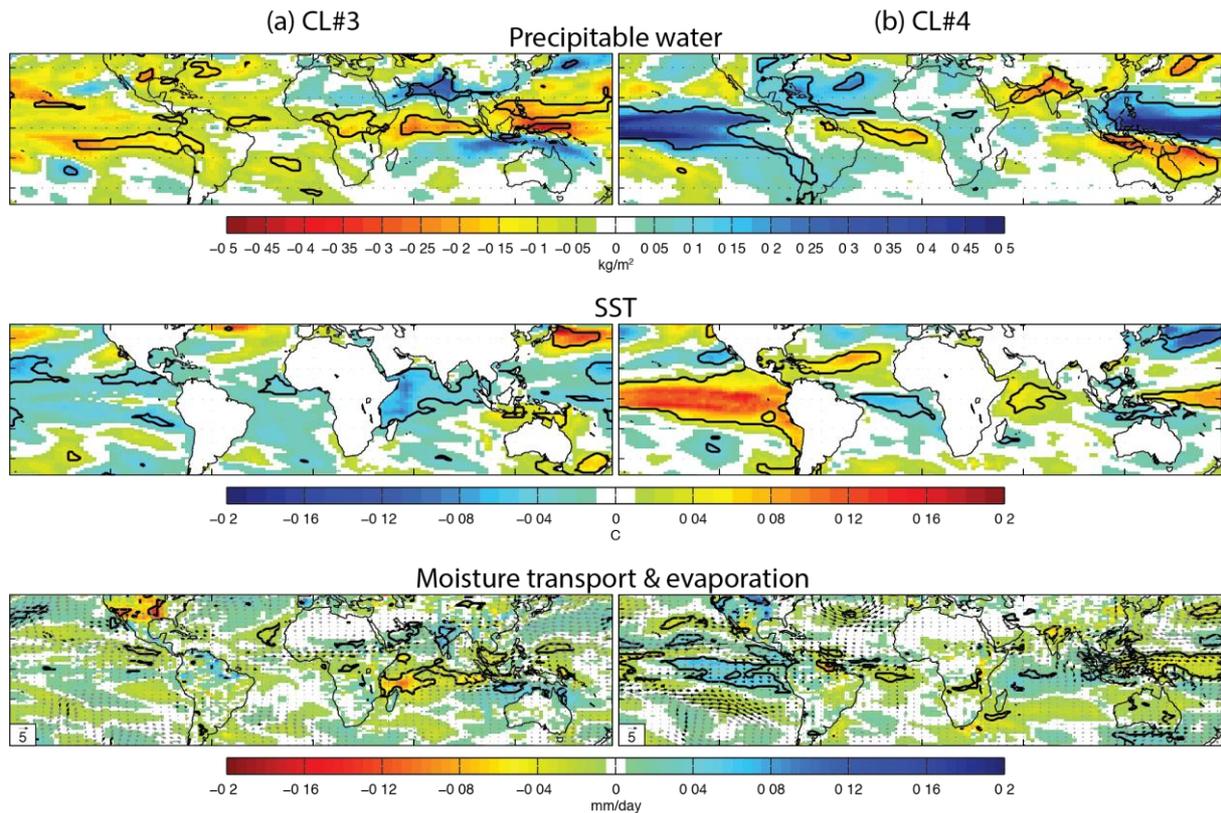
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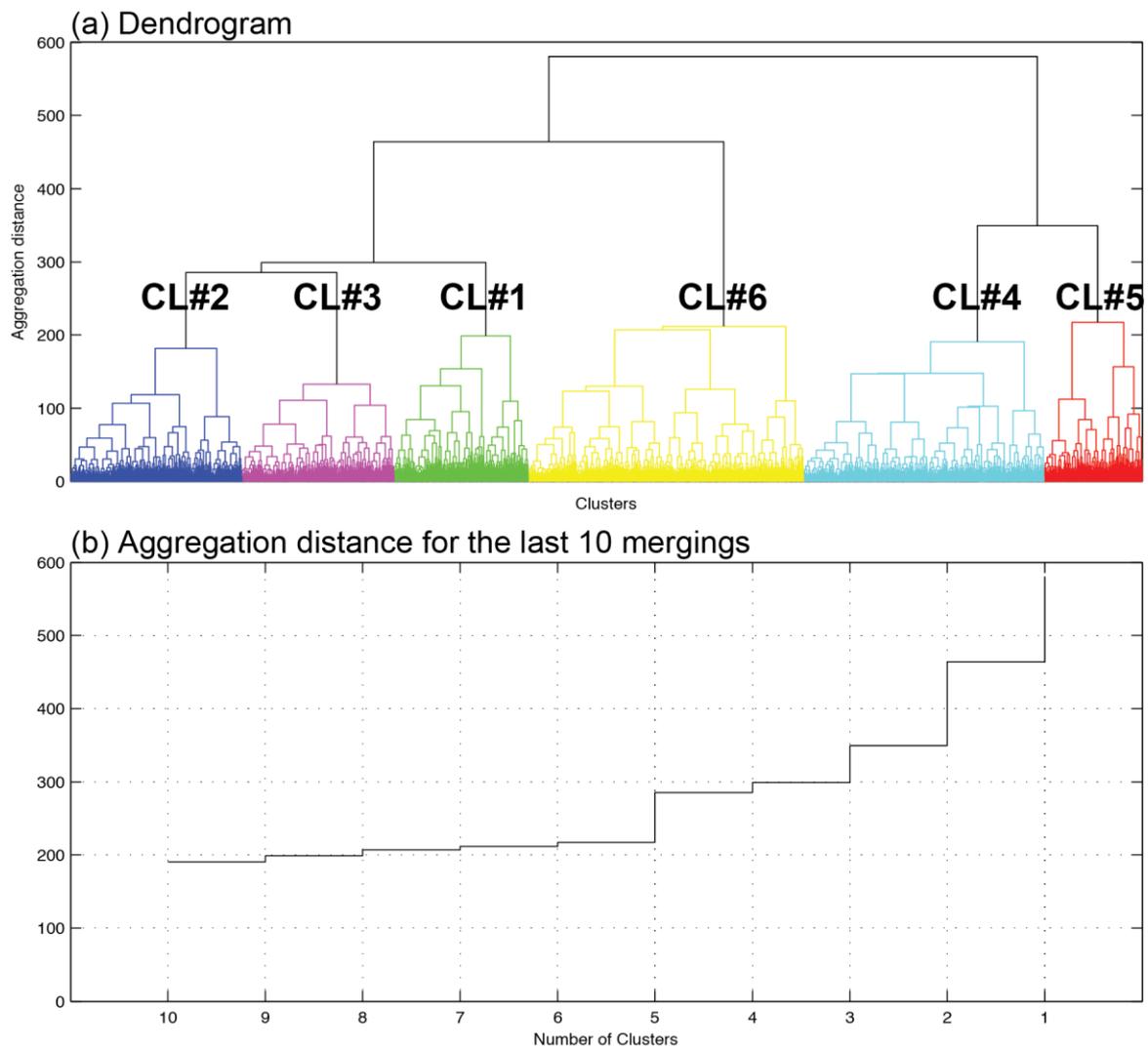
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Figure 14: (a; upper panel) Century-to-century evolution in the frequency of occurrence of the ISMR regime associated to La Niña-like – IOD- collocation from 6 to 0 kyr BP for the TR6AV-Sr02 simulation. This ISMR regime is extracted by applying the AHC clustering to anomalous Indian rainfall patterns from 6 to 0 kyr BP in the two 6000-yr simulation ensemble (TR5AS-Vlr01 and TR6AV-Sr02) with a 6-regime cutoff. The frequency of occurrence is expressed as the departure from its mean value. (a; bottom panels) Number of occurrences of this ISMR regime along each 2000-yr adjacent window and associated mean anomalous SST patterns. Only SST anomalies significant at the 95% confidence level according to a Student t test are shown. (b) Same as (a) but for the ISMR regime associated to El Niño-like – IOD+ collocation.



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Figure 15: (a) 4.9-3.8 minus 6-4.9 kyr BP differences in precipitable water, SST, moisture transport and evaporation ensemble mean anomalies associated to regime #3. Significant anomalies at the 95% confidence level according to a Student two-tailed t test are contoured in black for precipitable water, SST and evaporation and shown with black vectors for moisture transport. (b) Same as (a) but for regime #4.



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Figure A1: (a) Dendrogram tree obtained at the end of the AHC procedure applied to JJAS rainfall anomalies simulated over India by the 5-simulation ensemble from 6 to 3.8 kyr BP. (b) Evolution of the similarity metric for the 10 last mergings. See Appendix A for details.