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Deciphering the desiccation trend of the South Asian monsoon hydroclimate in a warming world

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31 **Abstract**

32 Rising propensity of precipitation extremes and concomitant decline of summer-monsoon rains
33 are amongst the most distinctive hydroclimatic signals that have emerged over South Asia since
34 1950s. A clear understanding of the underlying causes driving these monsoon hydroclimatic
35 signals has remained elusive. Using a state-of-the-art global climate model with high-resolution
36 zooming over South Asia, we demonstrate that a juxtaposition of regional land-use changes,
37 anthropogenic-aerosol forcing and the rapid warming signal of the equatorial Indian Ocean is
38 crucial to produce the observed monsoon weakening in recent decades. Our findings also show
39 that this monsoonal weakening significantly enhances occurrence of localized intense
40 precipitation events, as compared to the global-warming response. A 21st century climate
41 projection using the same high-resolution model indicates persistent decrease of monsoonal rains
42 and prolongation of soil drying. Critical value-additions from this study include (a) realistic
43 simulation of the mean and long-term historical trends in the Indian monsoon rainfall (b) robust
44 attributions of changes in moderate and heavy precipitation events over Central India (c) a 21st
45 century projection of drying trend of the South Asian monsoon. The present findings have
46 profound bearing on the regional water-security, which is already under severe hydrological-
47 stress.

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55 **1. Introduction**

56 Countries in South-Asia are endowed with bountiful summer-monsoon (June-September)
57 precipitation which is the principal water-supply source for sustaining agricultural operations,
58 power-generation, ecosystems and livelihoods of over 1.6 billion inhabitants and pivotal for the
59 regional hydrological cycle (Bookhagen and Burbank 2010; Hasson et al. 2013). Whilst the
60 inter-annual variability of the South-Asian Monsoon (SAM, also called ‘Indian monsoon’)
61 rainfall is tightly linked to El Nino-Southern Oscillation (Mishra et al. 2012), several studies
62 have reported significant declining trends in the SAM rainfall (e.g. Chung and Ramanathan
63 2005; Bollasina et al. 2011; Krishnan et al. 2013; Singh et al. 2014 and others) and associated
64 increase of aridity in recent decades (Kumar et al. 2013). The 144-year (1871-2014) time-series
65 of all-India summer monsoon rainfall (AISMR) reveals 10 occurrences of monsoon-droughts
66 during the first half (1871-1943), while the second half (1944-2014) witnessed 15 monsoon-
67 droughts¹ – i.e., a clear increase of drought incidence during the second half relative to the first.
68 The summer monsoon precipitation averaged over the Indian region (70-90°E – 10-28°N)
69 declined by ~7% during 1951-2005, significantly over the west coast and Indo-Gangetic plains
70 of north and east-central India as evidenced from observed rainfall datasets (e.g. Guhathakurtha
71 and Rajeevan 2008; Rajendran et al. 2012; Krishnan et al. 2013; Table 1). The decreasing trend
72 of SAM rainfall is corroborated by a significant weakening of monsoon low-level southwesterly
73 winds, the upper-tropospheric tropical easterlies from the outflow aloft, the large-scale monsoon
74 meridional overturning circulations (Rao et al. 2004; Joseph and Simon 2005; Sathiyamoorthi

¹There were 25 cases of monsoon droughts during 1871-2014 corresponding to the years 1873, 1877, 1899, 1901, 1904, 1905, 1911, 1918, 1920, 1941, 1951, 1965, 1966, 1968, 1972, 1974, 1979, 1982, 1985, 1986, 1987, 2002, 2004, 2009 and 2014. A monsoon-drought over India is defined when the AISMR deficiency exceeds one standard-deviation of long-term climatological mean. The June to September climatological seasonal total of AISMR is 848 mm and the standard-deviation is about 10% of the mean (see <http://www.tropmet.res.in>).

75 2005; Abish et al. 2013; Fan et al. 2010; Krishnan et al. 2013) and a significant increase in the
76 duration and frequency of ‘monsoon-breaks’ (dry spells) over India since 1970s (e.g. Ramesh
77 Kumar et al. 2009; Turner and Hannachi 2010). The weakening of SAM circulation is also borne
78 out from the declining frequency of the Bay of Bengal (BOB) monsoon depressions - the
79 primary rain-producing synoptic-scale monsoon disturbances (e.g. Rajeevan et al. 2000;
80 Krishnamurti et al. 2013; Singh et al. 2014). Concomitant with the declining monsoonal rains,
81 the Indian region additionally experienced a substantial rise in the frequency of daily
82 precipitation extremes in the post-1950s (Goswami et al. 2006; Rajeevan et al. 2008). Even as
83 these regional hydroclimatic changes tend to exacerbate the already diminishing water levels, the
84 underlying reasons are still unclear (Rodell et al. 2009).

85
86 Increase of atmospheric moisture content in response to global warming is expected to
87 enhance precipitation over broader regions of Asia and other monsoonal areas (e.g. Kitoh et al.
88 1997; Meehl and Arblaster 2003; May 2011; Wang et al. 2014). This is consistent with the
89 observed increasing trend in the intensity of “global monsoon” precipitation during the past three
90 decades, which has been attributed to enhanced moisture convergence and surface evaporation
91 due to increasing surface temperature (Hsu et al. 2012; Lau et al. 2013; Kitoh et al. 2013; Wang
92 et al. 2014). However, the Coupled-Model-Intercomparison-Project (CMIP) models show wide
93 inter-model spread in the simulated precipitation changes over South Asia, so that assessments of
94 regional hydroclimatic response to climate change have remained ambiguous (e.g. Kripalani et
95 al. 2007; Annamalai et al. 2007; Turner and Slingo 2009; Sabade et al. 2011; Fan et al. 2010;
96 Hasson et al. 2013; Saha et al. 2014). While most models suggest a greater likelihood of
97 enhanced monsoon precipitation due to global warming, they indicate a likely weakening of

98 large-scale monsoon and tropical circulation (e.g. Kitoh et al. 1997; Douville, 2000; Ueda et al.
99 2006; Cherchi et al. 2011; Rajendran et al. 2012; Krishnan et al. 2013).

100

101 Greenhouse-gas (GHG) and anthropogenic-aerosol forcing are generally regarded to
102 exert opposing influences on the SAM, with the former being conducive for precipitation
103 enhancement (e.g. Bollasina et al. 2011). Recent studies have attributed the observed declining
104 trend of SAM to anthropogenic-aerosol forcing (e.g. Ramanathan et al. 2005, Chung and
105 Ramanathan 2006, Bollasina et al. 2011). The impact of anthropogenic aerosols on surface
106 cooling over the Indian region is also seen in surface temperature observations, particularly
107 during the winter and pre-monsoon seasons (Krishnan and Ramanathan, 2002). Krishnamurti et
108 al. (2013) suggested that rapid increase of anthropogenic aerosol emissions over China and South
109 Asia in recent decades have enhanced the potential for disrupting organized convection in the
110 BOB monsoon depressions through aerosol indirect effects. Owing to the strong internal
111 variability of the Asian monsoon, aerosol impacts on monsoon precipitation are detectable
112 mostly on broader continental scales (e.g. South and Southeast Asia) and not as much on smaller
113 spatial scales (Salzmann et al. 2014). While the direct effects of aerosols on radiation and climate
114 are relatively better understood, there are considerable uncertainties concerning aerosol indirect
115 effects on tropical clouds and monsoon precipitation systems (e.g. Stevens and Feingold 2009;
116 Gautam et al. 2009; Turner and Annamalai 2012). A comparison of CMIP Phase-5 (CMIP5)
117 simulations between models with and without aerosol indirect effects showed large differences in
118 the SAM precipitation response during the second half of the 20th century (Guo et al. 2015). Also
119 the CMIP5 models show poor skills in simulating the recent declining trend of SAM rainfall,
120 with large ambiguities in future projections of monsoon rainfall (e.g. Chaturvedi et al. 2013;
121 Saha et al. 2014; Sharmila et al. 2014). Therefore, reliable attribution of the recent decline of

122 SAM rainfall yet remains a challenging problem given the large decadal-scale natural variations
123 of the monsoon system (Sinha et al. 2015).

124
125 Besides increasing GHGs and anthropogenic-aerosols, the South-Asian region also
126 underwent prominent changes in other forcing elements. Historical analysis of land-use change
127 in South and Southeast-Asia indicates that large-scale deforestation, agricultural expansion and
128 wetland clearance led to massive decline (~47%) of forests/woodlands in the last century while
129 agricultural land-area nearly doubled as that of 1880 (Flint and Richards 1991; Ramankutty et al.
130 2006). Another distinct regional signal is the rapid warming of equatorial Indian Ocean SST
131 (IOSST) by 0.5°-1°C during the last few decades (Mishra et al. 2012; Alory and Meyers 2009;
132 Swapna et al. 2013). The acceleration of the equatorial IOSST trend, which is in excess of the
133 global-warming signal, involves changes in the Indian Ocean circulation in response to
134 weakening of the summer monsoon cross-equatorial flow in the recent decades (Swapna et al.
135 2013). The rapid IOSST warming also favors enhancement of near-equatorial convection and
136 warming of the equatorial troposphere (Krishnan et al. 2006; Abish et al. 2013).

137
138 The objective of this study is to investigate the basis for the recent SAM changes and
139 draw insights into the mechanisms involved. For this purpose, we perform a suite of long-term
140 simulations in this study by employing a state-of-the-art variable resolution global climate model
141 (LMDZ4), developed at the Laboratoire de Météorologie Dynamique, France, having high-
142 resolution telescopic zooming (horizontal grid size ~35 km) over South-Asia. The use of high-
143 resolution overcomes the inadequacies and stringent limitations of coarse-resolution models in
144 representing surface topography and regional processes - particularly those involving land-use
145 change and aerosol-forcing. This is important from the viewpoint of regional-scale feedbacks

146 (eg., planetary-albedo) on the long-term behavior and stability of the Indian monsoon (Lenton et
147 al. 2008). Unlike limited-area models, another advantage of using a global variable-resolution
148 model is that it excludes imposition of time-dependent lateral boundary-conditions of 3D
149 atmospheric fields from global simulations. Furthermore, high-resolution is essential for
150 understanding changes in intense precipitation events (rain rate $> 100 \text{ mm day}^{-1}$) over the SAM
151 region, which are poorly represented in the CMIP models. A recent study by Sabin et al. (2013)
152 showed that the high-resolution LMDZ4 general circulation model (GCM) produces realistic
153 simulations of the mean SAM precipitation, organized moist convective processes and monsoon
154 synoptic disturbances. Datasets and model description are discussed in the following section.

155

156 **2. Datasets and Model**

157 *2.1 Datasets*

158 The observed precipitation datasets used for the analysis include (a) The India Meteorological
159 Department (IMD) high-resolution gridded ($0.25^\circ \times 0.25^\circ$) daily rainfall data for the period 1901-
160 2010 over the Indian region (Pai et.al., 2014) (b) The Asian Precipitation Highly Resolved
161 Observational Data Integration Towards Evaluation of Water Resources (APHRODITE) gridded
162 ($0.5^\circ \times 0.5^\circ$) rainfall over the Monsoon Asia land region (60°E – 150°E , 15°S – 55°N) for the period
163 1951-2007 (Yatagai et al. 2012) and (c) The Global Precipitation Climatology Project (GPCP
164 Version 2) gridded ($2.5^\circ \times 2.5^\circ$) rainfall dataset for the period 1979-2009 (Adler et al. 2003). We
165 also use the observed gridded ($0.5^\circ \times 0.5^\circ$) monthly surface temperature dataset during 1901-2010
166 from Climate Research Unit (CRU) (<http://badc.nerc.ac.uk/view/badc.nerc.ac.uk>); monthly sea
167 surface temperature (SST) and sea level pressure (SLP) during 1886–2005 based on the
168 HadISST (Rayner et al. 2003) and HadSLP (Allan and Ansell 2006) datasets, respectively.

169 Additionally, we used monthly mean atmospheric winds from the National Center for
170 Environmental Prediction (NCEP) Reanalysis for the period 1950–2005 (Kistler et al. 2001).

171

172 *2.2 Brief description of LMDZ4 GCM*

173 The LMDZ4 GCM is the atmospheric component of the Institut Pierre Simon Laplace (IPSL)
174 coupled model (Hourdin et al. 2006; Dufresne et al. 2013) and has been used to produce climate
175 change simulations for the Intergovernmental Panel for Climate Change (IPCC). It includes a
176 comprehensive treatment of atmospheric physical processes and land-surface components and
177 incorporates GHG-forcing and explicit representation of aerosol-radiation-cloud interactions.
178 The dynamical part of LMDZ4 is based on a finite-difference formulation of the primitive
179 equations of meteorology discretized on a sphere with stretchable grid (Z in LMDZ stands for
180 zoom) and uses the hybrid σ -p coordinate with 19 vertical layers (Sadourny and Laval 1984;
181 Hourdin et al. 2006). The energy and water cycles of soil and vegetation, terrestrial carbon cycle,
182 vegetation composition are simulated by a land-surface component known as Organizing Carbon
183 and Hydrology in Dynamic Ecosystems (ORCHIDEE) (Krinner et al. 2005). Land surface is
184 described as a mosaic of twelve plant functional types and bare soil. We designed a specific grid-
185 configuration by including high-resolution telescopic zooming (~35 km in longitude and
186 latitude) centered at 15°N, 80°E to simulate finer details of the SAM (Sabin et al. 2013).

187

188 *2.3 IPSL-CM5 Coupled Model*

189 The IPSL-CM5 coupled model couples the LMDZ atmosphere-land surface model to an ocean-
190 sea ice model (Dufresne et al. 2013). The ocean GCM (NEMOv3.2) ensures better representation
191 of bottom bathymetry, stream flow and friction at the bottom of the ocean, improved mixed-layer
192 dynamics, double diffusion, tracer diffusion and several key ocean processes including

193 prognostic interaction between incoming shortwave radiation into the ocean and the
194 phytoplankton. The IPSL-CM5 coupled model simulations have been generated for two
195 atmospheric grids. The low resolution (LR) grid has 96 x 95 points corresponding to a resolution
196 of 3.75° longitude x 1.875° latitude and the Medium Resolution (MR) version has 144×143
197 points, with a resolution of 2.5° longitude x 1.25° latitude.

198

199 **3. Design of LMDZ4 experiments**

200 The 20th century experiments consist of one set (HIST1, HISTNAT1) of long simulations for the
201 period 1886-2005, and another set (HIST2, HISTNAT2) for a shorter period 1951–2005. The
202 HIST1 and HIST2 experiments include both natural (varying solar-irradiance and volcanic-
203 aerosols) and anthropogenic (GHG, aerosols, ozone and land-use changes) forcing, while
204 HISTNAT1 and HISTNAT2 include natural forcing only. Each of these experiments is a single
205 member realization. We also performed a future projection following the Representative-
206 Concentration-Pathway-4.5 (RCP4.5) scenario spanning 2006-2095. To distinguish the effects of
207 GHG and non-GHG forcing, we performed two sets of decadal time-slice experiments
208 (HIST1_GHG and HIST1_PIGHG) for 1951-1960, 1961-1970, 1971-1980, 1981-1990 and
209 1991-2000 respectively². A two-year spin-up is performed before starting the decadal time-slice
210 integrations so as to avoid possible shocks from restarts (e.g. the 1951-1960 experiment was
211 initiated from January 1949). The HIST1_GHG time-slices account for observed time-varying
212 GHG-forcing only, but land-use and aerosol distributions are set to the 1886 values. The
213 HIST1_PIGHG employs varying land-use and aerosol distributions, but fixed GHG

²The high-resolution LMDZ4 simulations with zooming over South Asia are computationally intensive and time-consuming. Therefore, the HIST1_GHG and HIST1_PIGHG time-slice simulations across different decades were performed in parallel.

214 concentration corresponding to 1886. To confirm the robustness of results, we employed two
215 cumulus convection parameterization schemes for the historical simulations. The (HIST1 and
216 HISTNAT1) simulations are based on the Emanuel scheme (Emanuel, 1993), while the (HIST2
217 and HISTNAT2) simulations use the Tiedtke scheme (Tiedtke, 1989). All the other experiments
218 used Emanuel convection. A summary of all the experiments is provided in Table 2.

219

220 *3.1 Forcing used in the LMDZ4 experiments*

221 Global CO₂ concentration is directly prescribed in the simulations from 1886 to 2095 for
222 computing radiative budget. CO₂ concentration for the period 1886-2005 is derived from the
223 Law Dome ice core, the Mauna Loa and the NOAA global-mean records. From 2006 onwards,
224 CO₂ emissions and concentrations are projected by Integrated Assessment Models and Carbon
225 Cycle – Climate models (Dufresne et al. 2013). In the RCP4.5 scenario, the radiative forcing at
226 the end of 2100 is 4.5 Wm⁻² and CO₂ concentration stabilizes at 543 ppmv in 2150. The
227 concentration of other GHGs like CH₄, N₂O, CFC-11 and CFC-12 are directly prescribed in the
228 model radiative code based on the recommended CMIP5 datasets (Dufresne et al. 2013).

229

230 Land-use changes employed in the simulations are based on the harmonized dataset,
231 derived from yearly global land-cover maps at 0.5°x0.5° grid resolution for the period (1500–
232 2100), that smoothly concatenates historical reconstructions with future projections (Hurtt et al.
233 2011). Time-varying distributions of SST used in the simulations are derived from the IPSL-
234 CM5 coupled-model projections. The SST boundary-forcing in LMDZ4 experiments is basically
235 constructed by superposing the SST anomalies from the IPSL-CM5A-LR simulations on the
236 observed climatological mean SST from AMIP (Taylor et al. 2013). This approach retains the
237 realism of climatological mean SST and also overcomes the limitation of prescribing raw SSTs

238 warranted by limited-area models. Note that the SST anomalies for the different LMDZ4
239 experiments (i.e., HIST, HISTNAT, RCP4.5 HIST_GHG) come from the corresponding
240 IPSL-CM5A-LR simulations (Table.2). The same procedure is also used for specifying
241 sea-ice boundary conditions.

242

243 Time-varying 3-dimensional distributions of ozone, natural aerosols (e.g., sea-
244 salt, dust) and anthropogenic aerosols (e.g., sulfates, black carbon, particulate organic
245 matter) and gaseous reactive species are prescribed in the LMDZ4 experiments. These
246 fields come from the IPSL-CM5A-LR simulations coupled to the Interaction with
247 Chemistry and Aerosol (INCA) model (Szopa et al. 2013). In addition, the LMDZ4
248 simulations are directly forced by time-varying total solar irradiance (Lean et al. 2009)
249 and volcanic radiative-forcing for past periods. For the future, it is assumed that the solar
250 cycles repeat identical to the last cycle (cycle 23) with solar irradiance values from 1996
251 to 2008. The volcanic forcing is held constant for the future scenarios (Dufresne et al.
252 2013). Note that the present LMDZ4 simulations take into account the impact of
253 prescribed aerosols on precipitation through large-scale radiative forcing, but the aerosol
254 transports are not explicitly computed. Fully coupled / interactive aerosol-climate
255 simulations with the zoom configuration are currently beyond reach, here we wanted to
256 focus on the importance of climate change on the regional-scale SAM and to understand
257 the mechanisms involved. This is the reason for choosing the high-resolution zoomed
258 model configuration at the expense of ensemble realizations and aerosol coupling.
259 Furthermore, a recent study indicates that differences in the LMDZ simulations between
260 prescribed aerosols versus interactively computed aerosols have little impact in terms of
261 radiation and radiative forcing (Deandreis et al. 2012).

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4. Analysis of observed and simulated changes in SAM during recent decades

4.1 Long-term trends in seasonal monsoon rainfall

We employed two approaches to determine long-term trends in the seasonal monsoon rainfall and their statistical-significance. The first method estimates trends using linear least-square fit and a two-tailed student's t-test for statistical-significance (P values). Table 1 provides a summary of observed and simulated trends in the June-September (JJAS) seasonal rainfall averaged over the Indian land region 70-90°E, 10-28°N during 1951–2005 and for the 21st century following RCP4.5. The second method, based on the locally weighted polynomial regression (LOESS) technique, fits the data locally in segments and does not require specification of a global function for fitting a model to the data (Cleveland and Devlin 1988). The fit of local polynomial to each subset of the data uses weighted linear least-squares of first-degree. The “bandwidth”, which determines the data-length for fitting each local first-degree polynomial, is set to 35 years. The LOESS trends are estimated for the observed and simulated SAM rainfall time-series, including simulations from 21 CMIP5 models (Table 3). Finally, the local (in time) standard-deviations estimated from the CMIP5 trends are computed and plotted to assess the significance of monsoon rainfall trends.

It is interesting to note that the HIST1 and HIST2 simulations capture both the observed mean and decreasing trend of monsoon precipitation during later part of the 20th century (Figs.1a,2,3,4). The coarse-resolution IPSL-CM5A-LR (Fig.2d) and most of the CMIP5 coupled models fail to simulate the decreasing trend of SAM precipitation (e.g. Saha et al. 2014). Note that HIST1 and HIST2 show a significant ($P < 0.01$) reduction of monsoon rainfall over the Indian land region during 1951-2005 by ~16% and ~9%, respectively (Table 1, Figs.2-3). The

286 magnitude of the simulated rainfall trends is higher, particularly in HIST1, relative to the
287 observed rainfall decline (~7%). The prominence of the monsoon precipitation decline also
288 manifests in the LOESS trends in observations and LMDZ4, which clearly lie outside the range
289 of the CMIP5 trends (Fig.4). While the overall shape of the time-varying LOESS trends in
290 HIST1 is similar to that of the observations, the simulated amplitude is relatively larger than the
291 observed (Fig.4). Also notice that the declining trend, which starts from 1950s in observations
292 and late 1930s in HIST1, continues till the end of the 20th century. Furthermore, attribution of the
293 declining SAM rainfall to human influence is clearly supported by the absence of statistically-
294 significant trends in HISTNAT1 and HISTNAT2 (Fig.3, Table.1). The RCP4.5 projection from
295 LMDZ4 indicates a further decline of SAM rainfall and persistence of drought conditions during
296 coming decades (Fig.1a), which is contrary to the precipitation increase projected by the
297 ensemble-mean of CMIP5 models with large inter-model spread (Chaturvedi et al. 2013, Saha et
298 al. 2014, Sharmila et al. 2014).

299

300 *4.2 Long-term trends in aridity*

301 With rising surface temperatures and declining seasonal monsoon rains, aridity tends to be more
302 severe due to higher rates of potential evapotranspiration (Kumar et al. 2013; Vicente-Serrano, et
303 al. 2013; Thornthwaite, 1948). We examined aridity variations using the Standardized
304 Precipitation-Evapotranspiration Index (SPEI), from LMDZ4 simulations and observations
305 computed using the IMD rainfall (Pai et al. 2014) and CRU surface air-temperature (Fig.1b).
306 SPEI is an indicator of the spatio-temporal variability of droughts that considers the combined
307 effects of precipitation and temperature (via., evaporative demand) on droughts (Kumar et al.
308 2013; Vicente-Serrano et al. 2013). It is based on a difference between monthly precipitation (P
309 in mm) and monthly Potential Evapotranspiration (PET in mm), calculated for different time-

310 scales. SPEI is particularly suited for studying the effects of long time-scale droughts (> 3
311 months) on hydrological and ecological systems (Vicente-Serrano et al. 2013). The calculation of
312 PET is based on the Thornthwaite's method, which requires only data of monthly mean
313 temperature. The difference $D = P - PET$ is then accumulated for various time-scales,

$$D_n^k = \sum_{i=0}^{k-1} (P_{n-i} - PET_{n-i}), \quad n \geq k$$

314
315 where (k months) is the accumulation timescale and 'n' is the calculation index. The D values
316 are undefined for $k > n$. In calculating SPEI, a log-logistic probability distribution function is
317 used to fit the data-series of D and standardized (Vicente-Serrano et al. 2013). Values of SPEI <
318 -0.5 represent moderate-droughts (Kumar et al. 2013), and SPEI < -1.5 indicates severe-drought
319 conditions. A marked increase in the propensity of monsoon-droughts is seen during the post-
320 1950s in HIST1 and HIST2 (Fig.1b), which is congruent with observations (Kumar et al. 2013).
321 We have also noted that the SPEI index at 12-month and 24-month time-scales in HIST1 and
322 HIST2 exhibits clear decreasing trends during (1951-2005), which are absent in the HISTNAT1
323 and HISTNAT2 simulations (figures not shown).

324

325 *4.2 Long-term trends in frequency daily precipitation extremes*

326 A significant rise in the frequency of heavy-precipitation events (intensity $\geq 100 \text{ mm day}^{-1}$) over
327 Central India is reported to have occurred during recent decades (Goswami et al. 2006). Using
328 daily rainfall from observations and LMDZ4 simulations, we counted the number of heavy-
329 precipitation events over Central India ($74.5^\circ - 86.5^\circ\text{E}$, $16.5^\circ - 26.5^\circ\text{N}$) having rainfall intensity \geq
330 100 mm day^{-1} . The counts were determined for the JJAS season (122 days) of each year, so as to
331 produce year-wise time-series of frequency-count of heavy-precipitation events over Central
332 India (Fig.1c). An assessment of linear trends in these time-series during 1951-2005 and their

333 statistical significance is presented in Table.4 (see Auxiliary Fig.A1). Significant ($P < 0.01$)
334 increases in the frequency of heavy-precipitation occurrences are seen in observations (~30%)
335 and the HIST1 (~30%) and HIST2 (~42%) simulations; but not in HISTNAT1 and HISTNAT2.
336 Also the future projection under RCP4.5 shows further increase in the frequency of such heavy-
337 precipitation events. It must be mentioned that the robust increase of heavy-precipitation
338 occurrences in the monsoon region in HIST1 and HIST2 (Fig.1c) is an important value-addition
339 from the use of high-resolution, especially given that the statistics of monsoon precipitation
340 extremes and their changes are rather poorly captured by the CMIP models (Turner and Slingo
341 2009, Chaturvedi et al. 2012). The rising trend of heavy precipitation occurrences in the
342 backdrop of a weakening large-scale SAM will be taken up for discussion later.

343

344 **5. Attribution of monsoon weakening to anthropogenic forcing**

345 Figure 5 shows difference maps (HIST1 minus HISTNAT1) of mean JJAS rainfall and 850 hPa
346 winds for the period 1951-2005. The widespread and significant ($P < 0.05$) decrease of the
347 simulated monsoon rainfall over the Indo-Gangetic plains and the mountainous west coast
348 provides a basis for attributing the recent SAM weakening to anthropogenic-forcing (Fig.5). A
349 clear weakening of the large-scale monsoon circulation is evident from the anticyclonic pattern
350 with anomalous easterly winds extending far into the Arabian Sea. The precipitation
351 enhancement and associated circulation anomalies over the Himalayan foothills, near-equatorial
352 Indian Ocean, downstream areas of Indo-China and Southeastern China (Fig.5) are reminiscent
353 of ‘monsoon-breaks’ over India (e.g. Krishnan et al. 2000). Interestingly, the downstream
354 precipitation enhancement seen over Southeastern China in the LMDZ4 simulations (Fig.5) and
355 was not captured in the aerosol-forced response of Bollasina et al. (2011). Consistent with
356 observed pattern of rainfall trends (see Chung and Ramanathan, 2006), the simulations also

357 display anomalous rainfall deficits over northeastern China and West-African Sahel, with the
358 latter being linked to SST-warming in the equatorial Indian and Atlantic Oceans (Giannini et al.
359 2003) and SST variations in the tropical Pacific (Semazzi et al. 1988). A further reduction of
360 monsoon rainfall over north-central parts of India is noted in the RCP4.5 projection in the near-
361 future (see Auxiliary Fig.A2).

362

363 *5.1 Distinguishing the mean monsoon response to GHG and non-GHG forcing*

364 To differentiate the effects of GHG versus non-GHG forcing, we examined the HIST1_GHG and
365 HIST1_PIGHG experiments. We compared the JJAS mean rainfall and 850 hPa winds from
366 these two simulations with HISTNAT1 for the 5 decadal time-slices (1951-1960), (1961-1970),
367 (1971-1980), (1981-1990) and (1991-2000). An anomalous decrease of monsoon rainfall over
368 the Indian landmass and weakening of monsoon circulation is noted in the (HIST1_PIGHG
369 minus HISTNAT1) difference maps for the 5 decadal time-slices (figure not shown); whereas the
370 (HIST1_GHG minus HISTNAT1) difference showed anomalous intensification of monsoon
371 circulation and rainfall over the Indian subcontinent (see Fig.6b). Figure.6 illustrates the
372 decomposition of the SAM response to GHG and non-GHG forcing using difference maps
373 composited for the period (1951-2000). For example, the difference $\delta(\text{No_GHG}) = (\text{HIST1}$
374 $\text{minus HIST1_GHG})$ considers the effect of all forcing, except GHG, including the SST response
375 and internal variability. In this case, a prominent weakening of the monsoon circulation and
376 rainfall reduction over India can be noted in HIST1 relative to HIST1_GHG (Fig.6a). On the
377 other hand the effect of GHG-forcing, including the SST response to GHG-forcing, is borne out
378 in the difference $\delta(\text{GHG}) = (\text{HIST1_GHG minus HISTNAT1})$ map (Fig.6b) which shows a
379 prominent intensification of monsoon rainfall and circulation. The difference $\delta(\text{GHG_Atmos}) =$
380 $(\text{HIST1 minus HIST1_PIHG})$ represents the atmospheric radiative effect of GHG-forcing with

381 imposed SSTs. In this case, a slight weakening of monsoon flow and small rainfall decrease is
382 seen (Fig.6c), indicative of a possible increase of static-stability due to rapid increase of moisture
383 in a warming world. This point will be discussed again later. The difference map in Fig.6d
384 [$\delta(\text{GHG_SST}) = \delta(\text{GHG})$ minus $\delta(\text{GHG_Atmos})$] indicates the effect of SST changes only due
385 to GHG increase plus the internal variability of SSTs. An intensification of monsoon circulation
386 and rainfall is prominently seen in this case, which is apparently linked to the pronounced SST
387 warming ($\sim 1^\circ\text{C}$) in the north-central Arabian Sea and north Bay of Bengal in HIST1_GHG
388 (figure not shown). Also the contrasting monsoonal responses in HIST1_GHG and
389 HIST1_PIGHG are indicative of the impact of non-GHG forcing on the recent SAM weakening.

390

391 **6. Non-GHG forcing and SAM weakening**

392 Several studies have linked the recent SAM weakening to increases in anthropogenic aerosols
393 (eg., Ramanathan et al. 2005, Chung and Ramanathan, 2006, Meehl et al. 2008, Bollasina et al.
394 2011, Ganguly et al. 2012, Salzmann et al. 2014, Sanap et al. 2015). The aerosol direct-radiative
395 effect which causes reduction of surface insolation (“solar-dimming”) through scattering and
396 absorption can weaken the SAM through a variety of mechanisms viz., decrease of meridional
397 sea-surface-temperature (SST) gradient between equator and 25°N (Ramanathan et al. 2005;
398 Meehl et al. 2008), inter-hemispheric energy imbalance (Bollasina et al. 2011), weakening of
399 tropospheric-temperature gradient and vertical wind-shear (Ganguly et al. 2012), decrease of
400 water vapor availability (Salzmann et al. 2014), surface cooling over the Indian subcontinent
401 (Sanap et al. 2015) and so on. Here, we discuss on the temporal evolution of the major regional-
402 forcing elements and the SAM response from the (HIST1 + RCP4.5) simulations (Fig.7).

403

404 *6.1 Regional land-use change*

405 The South and Southeast Asian region experienced significant land-use and land-cover changes
406 during the 19th and 20th centuries (Flint and Richards 1991; Ramankutty et al. 2006; Hurtt et al.
407 2011). The time-evolution of tree-fraction, crop-fraction and planetary albedo averaged over the
408 Indian land region (70°E-90°E; 10°N-28°N) is presented in Fig.7a. In addition, spatial maps of
409 tree-fraction/crop-fraction during 1951-2000 are provided in Figs.8a-b and the corresponding
410 difference [(1891-1930) minus (1951-2000)] maps in Figs.8c-d. We note from Fig.7a that the
411 crop-fraction over India increased by about 45% and tree-fraction declined by about 30% during
412 1886-2005. This is consistent with earlier reports of crop-area change over the region, which
413 expanded by ~45% at the expense of all natural vegetation types (Flint and Richards 1991,
414 Ramankutty et al. 2006). Also note that the RCP4.5 scenario allows for a partial recovery of tree-
415 fraction from 2030s (Fig.7a). From Fig.8, one can notice that the expansion (contraction) of
416 crop-area (tree-area) during (1951-2000) relative to (1891-1930) is seen across various parts of
417 South and Southeast Asia. Interestingly, the time-series of simulated regional planetary-albedo
418 shows an increasing trend from 1886 to mid-1980s, and a near-stationary pattern until 2040s,
419 followed by a slight decrease thereafter (Fig.7a). The overall increase in regional planetary-
420 albedo (including cloud effects) during the period 1886-2005 is 9%. Further, note that the
421 decrease (increase) of crop-fraction (tree- fraction) in the late 21st century proceeds rather
422 gradually as compared to the faster rate of change [with opposite sign] in the early 20th century.
423 This is also reflected in the gradual decrease of the simulated regional planetary-albedo
424 variations in the late 21st century.

425

426 *6.2 Anthropogenic aerosol-forcing*

427 The time-evolution of the simulated anthropogenic aerosol-forcing at the top-of-atmosphere
428 (TOA) averaged over the Indian region is shown in Fig.7b. The TOA aerosol radiative forcing

429 ranges between -0.8 and -1.6 Wm^{-2} during 1960s and mid-2020s, and becomes less negative after
430 2020s (Fig.7b). Although, the RCP4.5 scenario considers abatement of emissions, note that the
431 projected aerosol-forcing (negative) continues till 2050. Spatial maps of the simulated
432 anthropogenic aerosol-forcing at the TOA and the atmospheric absorption (i.e., difference in
433 aerosol-forcing [TOA minus surface]) from HIST1 during (1951-2005) are shown in Fig.9. Note
434 that the simulated TOA aerosol-forcing is about -2 Wm^{-2} over north-central India and larger
435 (about -4 Wm^{-2}) over east-central China (Fig.9a). The simulated atmospheric absorption is high
436 ($3-4 \text{ Wm}^{-2}$) over north India and maximum ($> 6 \text{ Wm}^{-2}$) over east-central China (Fig.9b).
437 Ramanathan et al. (2005) and Chung and Ramanathan (2006) argued that the declining trend of
438 SAM rainfall is largely due to increasing emissions of absorbing aerosols (e.g. black carbon)
439 over the Asian region. On the other hand, Bollasina et al. (2011) propose that the recent SAM
440 weakening is influenced largely by increase of sulfate aerosols (scattering-type) in the NH. An
441 examination of aerosol optical depths at 500 nm from the HIST1 simulation, for both black
442 carbon and sulfate, revealed that both species have large contributions to anthropogenic aerosol-
443 forcing over East Asia, while black-carbon contribution is significant over the South Asian
444 region and sulfate contribution is significant over central Eurasia (figure not shown).

445
446 Both land-use change and aerosol-forcing can influence the SAM through changes in
447 regional planetary-albedo (Lenton et al. 2008). Given the reflectivity differences between
448 absorbing and scattering aerosol-types, increases in absorbing aerosols need not necessarily lead
449 to increase of planetary-albedo. The marginal increase in the simulated regional planetary-albedo
450 after mid-1980s (Fig.7a) is more indicative of radiative effects from absorbing-aerosols. We also
451 note that the increase of atmospheric absorption over the Indian region is associated with
452 increased atmospheric stability in HIST1 during the post-1950s (figure not shown). As discussed

453 before, the main point from Fig.7a is that the combined effects of land-use change and
454 anthropogenic aerosol-forcing apparently yield an overall increase in regional planetary-albedo
455 during the historical period. Increases in planetary-albedo can reduce precipitation through
456 compensating subsidence required to maintain thermal equilibrium (Charney 1975). This albedo-
457 precipitation feedback is substantiated in the simulated decrease of SAM rains during the
458 second-half of the 20th century and in the future (Auxiliary Fig.A2).

459

460 *6.3 Equatorial Indian Ocean SST warming signal*

461 Another major regional signal is the warming of equatorial IOSST, by 0.5°-1°C during the last
462 few decades, and projected to continue through the 21st century (Figs.7b, 10). While drivers such
463 as aerosols or GHGs are independent forcing which can drive SST and atmospheric circulation
464 changes, our focus here is on the fact that changes in surface winds can in turn alter SST through
465 oceanic processes, and are taken into account in the CMIP coupled models (see Fig.10). For
466 example the recent equatorial IOSST warming trend, which is in excess of the global-warming
467 signal (Fig.10c), is a dynamical response of the Indian Ocean to changing monsoonal winds
468 (Swapna et al. 2013). Mishra et al. (2012) analyzed coupled patterns of observed SST and
469 Indian monsoon rainfall variability for the 20th century using maximal covariance analysis. The
470 second mode of variability from their analysis suggests a linkage between the recent decreasing
471 trend of monsoon rainfall over the Indo-Gangetic plains and the equatorial IOSST warming
472 signal and this has also been pointed out by other investigators (e.g., Krishnan et al. 2013, Abish
473 et al. 2013, Swapna et al. 2013, Roxy et al. 2015). The connection between the declining
474 monsoon rains and IOSST warming signal is clearly borne out by the precipitation increase in
475 the near-equatorial region in the HIST1 simulation (Fig.5).

476

477 *6.4 Weakening of SAM circulation and hydrological consequences*

478 The temporal evolution of the SAM response in HIST1 shows a clear increasing trend of sea-
479 level pressure (SLP) over the Indian region during the recent few decades and is supported by the
480 observed SLP trend (Fig.7c). On the other hand, the HISTNAT1 simulation does not exhibit any
481 long-term trend in the SLP variations. Also note that the simulated SLP increase in the 21st
482 century in Fig.7c is consistent with the likely projected decline of SAM rainfall in future.
483 Furthermore, the SAM weakening in the HIST1+RCP4.5 simulation is corroborated by the
484 decreasing trend of the easterly vertical-shear of zonal-winds (Fig.7e).

485
486 A sustained weakening of monsoon in a warming environment has direct hydrological
487 consequences (Kumar et al. 2013). On multi-decadal and centennial time-scales, soil moisture-
488 climate feedbacks can also induce vegetation shifts and contribute to climate change
489 (Seneviratne et al. 2010). Note that the HIST1 simulation indicates depletion of soil-moisture by
490 ~23% during 1886-2095 (Fig.7f). Persistence of soil-moisture deficiencies over the SAM region,
491 an identified hot-spot of land-atmosphere coupling to precipitation (Koster et al. 2004), can in
492 turn reduce evapotranspiration particularly in soil-moisture limited regimes and semi-arid
493 transition zones (e.g. Manabe and Delworth 1990, Koster et al. 2004, Seneviratne et al. 2010)
494 and impact crop-production. In a recent study by Ramarao et al. (2015), we analyzed the land-
495 surface response over the Indian region to changing monsoon precipitation by means of the
496 LMDZ4/HIST1 simulations and noted that a 5% depletion of soil-moisture during (1951-2005)
497 resulted in a decrease of evapotranspiration by nearly 10% (figure not shown).

498
499 *6.5 Changes in moderate and heavy precipitation occurrences over the SAM region*

500 Another critical hydrological ramification comes from the rising trend in the frequency and
501 intensity of daily precipitation extremes, which are expected to increase over several regions of
502 the world in response to global warming (Toreti et al. 2013). It is noted from the HIST1
503 simulation that the surface warming over the Indian region during (1886-2005) is over 2°C
504 (figure not shown) and correspondingly the simulated moisture-content over the Indian region
505 increased by ~24% during this period (Fig.7d). Rising humidity levels over the Indian region can
506 also be confirmed from surface specific-humidity observations, which are available from early
507 1970s (figure not shown). In the SAM context, increases in heavy precipitation (intensity > 100
508 mm day⁻¹) occurrences over Central India have happened at the expense of moderate events (5
509 mm day⁻¹ ≤ rainfall intensity < 100 mm day⁻¹), with the latter having declined significantly since
510 1950s (Goswami et al. 2006, Rajeevan et al. 2008).

511
512 Here, we examine changes in the distribution of heavy and moderate monsoon
513 precipitation due to GHG and regional-forcing elements, respectively. For this purpose, we
514 analyze year-wise frequency-counts of heavy and moderate daily precipitation events over
515 Central India (74.5° – 86.5°E, 16.5°– 26.5°N) during 1951-2000 from HIST1 and HIST1_GHG
516 and compared them to the frequency-count in HISTNAT1 (see Fig.11). Year-wise counts were
517 calculated for each JJAS season (122 days) during the entire 50-year period. Changes in
518 frequency-count were computed relative to the mean-counts of heavy and moderate
519 precipitation-types from HISTNAT1 averaged for the period (1951-2000). Fig.11 shows box-
520 whisker plots based on distributions of changes in yearly counts of moderate and heavy
521 precipitation events over Central India from HIST1 and HIST1_GHG. It is interesting to note
522 that HIST1 captures the contrasting change of the frequency-counts between the moderate and
523 heavy precipitation types, whereas the frequency-count of both categories increases in

524 HIST1_GHG. With regard to changes in the heavy-precipitation category, notice that the
525 median (i.e., second quartile Q_2) is nearly the same for both HIST1 and HIST1_GHG; although
526 the third quartile (Q_3) is significantly higher in HIST1 as compared to HIST1_GHG (Fig.11).
527 This suggests that the surplus of heavy precipitation in HIST1 is basically contributed by very
528 intense precipitation events. Given that increase of moisture is common to both HIST1 and
529 HIST1_GHG, the increase of more number of intense precipitation events in HIST1 is
530 attributable to enhancement of deep localized convection, which is more likely in an atmosphere
531 with weak vertical-shear of the SAM circulation (e.g. Romatschke and Houze 2011).
532 Conversely, a weak SAM circulation with depleted vertical-shear inhibits organization of meso-
533 scale convective systems and suppresses monsoonal rains in the moderate category (Stano et al.
534 2002). These discussions suggest that increase of atmospheric moisture (Fig.7d) and decrease of
535 easterly vertical shear of the SAM circulation (Fig.7e) in HIST1 provide conditions favorable for
536 localized heavy precipitation occurrences at the expense of moderate monsoonal rains.

537

538 **7. Physical mechanism of weakening of SAM**

539 Based on the present results, it is suggested that the physical mechanism for the recent decline of
540 SAM precipitation is rendered through a slowdown of the monsoon meridional overturning
541 circulation in response to the regional-forcing elements. The HIST1 simulations reveal that land
542 use changes and anthropogenic aerosol-forcing over South Asia have significantly enhanced the
543 regional planetary-albedo. Dynamical constraints for maintaining thermal equilibrium are
544 supported by subsidence induced by the near-equatorial precipitation anomalies associated with
545 the IOSST warming signal. Figure.12a shows anomalous subsidence over the Indian landmass,
546 together with near-equatorial ascending motions, indicating weakening of the monsoon

547 meridional overturning circulation in HIST1 relative to HISTNAT1. Further, this suppressed
548 convection over the subcontinent facilitates anomalous interactions of the monsoon and
549 midlatitude circulations and sets up large-scale quasi-stationary anomalies in the middle and
550 upper-troposphere, so that sinking of cold and dry northwesterly winds in turn can suppress
551 precipitation leading to prolonged dry conditions over the subcontinent (Auxiliary Fig.A3, see
552 also Krishnan et al. 2009, Krishnamurti et al. 2010). The equatorial tropospheric warming
553 (Auxiliary Fig.A3) is in part associated with the IOSST warming signal and also due to an
554 overall increase in tropical condensational heating in HIST1. With continued equatorial IOSST
555 warming, the RCP4.5 projection indicates persistence of anomalous ascending (descending)
556 motions in the near-equatorial region (Indian subcontinent) in the coming decades (Fig.12b).

557

558 **8. Value additions from high-resolution**

559 Here, we discuss value additions from the high-resolution SAM simulations using LMDZ4 vis-à-
560 vis the coarse-resolution IPSL-CM5A model.

561

562 *8.1. Resolving the monsoon orographic precipitation over the Western Ghats*

563 The narrow Western Ghats (WG) escarpment is an important topographic feature that runs for
564 over 1600 km almost parallel to the Arabian Sea with highest peaks having altitudes greater than
565 2600 m (Kale 2010; Xie et al. 2006). Impingement of monsoon southwesterly winds forces moist
566 air ascent over the windward slopes of the WG, causing substantial precipitation as much as
567 3000 mm in the monsoon season (Xie et al. 2006), and the associated latent heat of condensation
568 estimated from satellite measurements is around 3-4 K day⁻¹ (Choudhury and Krishnan, 2011).

569 High-resolution is very important to capture the mean monsoon orographic precipitation along
570 the WG and Himalayan foothills, and also to resolve moisture-gradients across the Indo-Pak and

571 Hindu-Kush mountainous region for supporting moist convective processes over the Indo-
572 Gangetic plains (Sabin et al. 2013). Improvements in the WG orographic precipitation in
573 LMDZ4 relative to the IPSL-CM5A models are clearly seen from (Auxiliary Fig.A4). Mean and
574 standard-deviation values of JJAS rainfall averaged over the Western Ghats (72°-76°E, 10°-
575 19°N) and Indian land region (70°-90°E, 10°-28°N) from GPCP, LMDZ4, IPSL-CM5A-LR and
576 IPSL-CM5A-MR are given in Table.5. Also shown is information about 850 hPa monsoon zonal
577 winds, averaged over a broader region (5°N–22°N, 55°E-90°E) around India. Note that the
578 magnitudes of mean precipitation and zonal winds are considerably underestimated in the IPSL-
579 CM5A models, whereas LMDZ4 shows closer agreement with observations (Table.5). It is to be
580 also noted that the variability of monsoon winds and precipitation is stronger in LMDZ4 relative
581 to the IPSL-CM5A models. The differences in mean precipitation over Central India between
582 HIST1 and HIST2 (Auxiliary Fig.A4) are indicative of the strong internal variability of SAM
583 (Salzmann et al. 2014) and the sensitivity of precipitation simulation to choice of cumulus
584 convection scheme (i.e., Emanuel in HIST1 / Tiedtke in HIST2). However, note that the JJAS
585 mean rainfall averaged over the larger land region is comparable in HIST1 (6.9 mm day⁻¹) and
586 HIST2 (6.3 mm day⁻¹). Also both HIST1 and HIST2 show decreasing trends of monsoon
587 precipitation averaged over the larger land region, despite sub-regional scale differences. We
588 have also confirmed the weakening trend of low-level monsoon winds in HIST1, HIST2 and
589 NCEP reanalysis during 1951-2005 (figure not shown). In a recent separate study by Ramarao et
590 al. (2015), we note that the high resolution LMDZ4 simulation significantly improves the
591 climatological surface hydrological balance between Runoff (R) and Precipitation minus
592 Evapotranspiration (P-ET) over the over the Indian land region as compared to the IPSL coarse
593 resolution model. Another prominent feature of the high-resolution simulation is the anomalous

594 precipitation enhancement near the eastern Himalayan foothills and downstream areas of Indo-
595 China (Fig.5), emerging from interplay between large-scale and regional-scale circulation
596 dynamics and mesoscale orographic amplification (Vellore et al. 2014).

597

598 8.2. *Coupling of monsoon precipitation and wind variations under changing climate*

599 Monsoon precipitation variability over the WGs is closely linked to strength of the monsoon
600 low-winds over the Arabian Sea (Krishnamurti and Bhalme 1976; Joseph and Sabin 2008).

601 Strong southwesterly winds (wind speed $> 20 \text{ ms}^{-1}$) favor enhanced precipitation, so that latent

602 heating in turn strengthens the winds and forces upward motions over the WG and the Arakan

603 Yoma in Myanmar (e.g. Xie et al. 2006; Choudhury and Krishnan 2011; Sabin et al. 2013). We

604 examined the coupling between monsoon precipitation and low-level winds using observations,

605 the LMDZ4 and IPSL-CM5A-LR models. We first performed an Empirical Orthogonal Function

606 (EOF) / Principal Component (PC) analysis of JJAS rainfall over the WGs and Peninsular India

607 during (1951-2005) using observations (APHRODITE), the LMDZ4/HIST1 and IPSL-CM5A-

608 LR simulations. Later, the PC1 index of rainfall was regressed on the 850 hPa winds. The

609 leading EOF/PC and regression of wind pattern are shown in Auxiliary Fig. A5. The leading PC

610 time-series shows a decreasing trend in APRHODITE and LMDZ4/HIST1; and wind regression

611 shows a westerly pattern implying a positive correlation between the declining WG rainfall and

612 monsoon low-level circulation during (1951-2005) - a period associated with significant global-

613 warming. In other words, the monsoon wind-precipitation connection is consistently seen in the

614 high-resolution HIST1 simulation, but not in the coarse-resolution ISPL-CM5A-LR counterpart.

615 In fact, the wind and precipitation variations are anti-correlated in the IPSL-CM5A-LR

616 simulation, with the PC1 rainfall time-series not displaying a decreasing trend.

617

618 Our understanding suggests that high-resolution is crucial for capturing the coupled
619 variability of the WG orographic precipitation and large-scale monsoon winds. Basically, the
620 internal variability of SAM rainfall is tied to both the monsoon circulation dynamics, as well as
621 thermodynamical effects arising from radiative-convective feedbacks (Krishnamurti and Bhalme
622 1976). Most of the CMIP coarse-resolution models predominantly exhibit the thermodynamical
623 sensitivity of the tropical/monsoonal response to global-warming, which results in enhancement
624 of precipitation despite damping of the dynamical component (i.e., upward velocity, large-scale
625 winds, horizontal-advection, etc) of monsoon circulation (see Kitoh et al. 1997; Douville et al.
626 2000; Veechi et al. 2006; Ueda et al. 2006; Cherchi et al. 2011). On the other hand, the high-
627 resolution LMDZ4 simulations better capture the wind-precipitation coupling, and thus a
628 weakening of the monsoon flow induced by the regional-forcing leads to precipitation decrease
629 over the Indian region.

630

631 **9. Concluding remarks**

632 Our findings suggest that the collaborative influence of regional land-use change, anthropogenic-
633 aerosol forcing and accelerated IOSST warming signal have conspired to weaken the SAM in
634 recent decades. It must be pointed out that the realism of monsoon simulations in LMDZ4 and
635 new insights gained from this study entail the use of high-resolution, which is crucial to resolve
636 the interactions between the high orographic precipitation along the narrow Western Ghats and
637 southwesterly monsoon winds. The high-resolution simulations also illustrate that the weakening
638 of SAM circulation significantly alters the rainfall distribution over Central India by promoting
639 surplus occurrences of intense localized precipitation events at the expense of moderate events.
640 In this context, it may be mentioned that the coarse-resolution CMIP models have major

641 shortcomings in detecting intense precipitation events and their changes particularly over the
642 SAM region (Chaturvedi et al. 2012) and the tropics in general (Toreti et al. 2013).

643
644 Sensitivity studies using a simple conceptual model of the monsoon (see Zickfeld et al.
645 2005; Lenton et al. 2008) suggest that increase of regional planetary-albedo, due to land-use
646 change and/or aerosol forcing, beyond a critical threshold (~ 0.5) can potentially cause the Indian
647 summer monsoon to pass a tipping-point, thereby completely altering the state of the monsoon.
648 Although projected albedo variations (Fig.7a) are below this threshold, the hydrological-stress
649 associated with declining monsoon rains is already severe (Kumar et al. 2013). It is noted that
650 the projected monsoon rainfall tends to partly recover in later part of the 21st century (Auxiliary
651 Fig.A6), although drier conditions continue to prevail over the subcontinent. This leads to an
652 important question, as to whether the SAM will recover in future. This possibility is still not
653 clear especially in the light of the persistent equatorial IOSST warming signal. Observations and
654 model simulations suggest that surface warming is often amplified in the tropical troposphere
655 due to an overall increase in condensational heating by moist ascending air in regions of
656 convection (Santer et al. 2008). This implies that, in addition to mitigating the effects of
657 anthropogenic-aerosols and land-use changes, a slowdown of equatorial IOSST warming may
658 hold the key for the SAM revival.

659
660 It is realized that the use of stand-alone atmospheric GCM can miss certain aspects of
661 ocean-atmosphere coupling. On the other hand, it is known that the CMIP coupled models have
662 large SST biases which can completely modify the monsoon and tropical circulation, leading to
663 poor skills in capturing regional precipitation variability (Roehrig et al. 2013, He et al. 2014). In
664 fact, atmospheric GCM simulations with imposed SST are reported to be far more realistic in

665 capturing the observed decadal variations of the Sahel monsoon rainfall as compared to the
666 CMIP5 coupled models (Roehrig et al. 2013). While errors due to lack of coupling are mostly
667 related to internal variability, stand-alone atmospheric GCM simulations well reproduce
668 circulation and precipitation response to long-term climate changes (He et al. 2014).

669

670 Bony et al. (2013) argue that robust tropical precipitation changes are associated with
671 weaker radiative cooling in response to rising levels of carbon dioxide concentration, which
672 affects the strength of the vertical component of atmospheric circulation. However, the findings
673 from this study suggest that the decline of SAM circulation and precipitation in the latter half of
674 the 20th century is largely attributable to changes in regional-forcing namely land-use changes,
675 anthropogenic aerosols along with the equatorial IOSST warming signal, in addition to the GHG-
676 forcing. The combined effects of these elements tend to increase the regional planetary-albedo
677 over the Indian subcontinent, resulting in weakening of monsoon meridional overturning
678 circulation, decreased precipitation/convection and soil-moisture reduction. We also realize that
679 remote-forcing (e.g., sulfate aerosol-emissions over Central Europe, long-range transports of
680 pollutants over the Eurasian region, etc) can have implications on the large-scale Asian summer
681 monsoon (e.g., Lelieveld et al. 2002, Cowan and Cai, 2011). However, it must be noted that the
682 present simulations preclude us from separating the roles of the individual forcing components
683 and drawing any inferences about the relative roles of regional and remote forcing on the SAM.
684 Further investigations will also be necessary to better comprehend the role of soil-moisture
685 feedbacks on monsoon precipitation.

686

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694

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931

932 **Figure Captions**

933

934 **Figure 1 | Temporal evolution of the observed and simulated monsoon hydroclimatic**
935 **signals. (a)** Time-series of 5-year running mean of seasonal (June-September) monsoon
936 precipitation (mm day^{-1}) averaged over land-points in the Indian region ($70^{\circ}\text{E}-90^{\circ}\text{E}$, $10^{\circ}\text{N}-28^{\circ}\text{N}$;
937 see inset in panel ‘b’) – based on IMD observations (black line), HIST1 (brown), RCP4.5 (red),
938 HISTNAT1 (blue), CMIP5-Multimodel-Mean (dark grey), IPSL-CM5A-LR (purple) and IPSL-
939 CM5A-MR (green). The grey shading is the inter-model spread, ranging between $\mu + \sigma$ and $\mu - \sigma$,
940 where μ and σ are the mean and standard-deviation based on simulations from 21 CMIP5 models
941 (Table.3) **(b)** Time-series of SPEI averaged over the Indian region **(c)** Temporal variation in the
942 number of heavy-precipitation events (precipitation intensity $\geq 100 \text{ mm day}^{-1}$) during the JJAS
943 season over the land points of central India ($74.5^{\circ}\text{E}-86.5^{\circ}\text{E}$, $16.5^{\circ}\text{N}-26.5^{\circ}\text{N}$; see inset). The black,
944 brown, blue and red lines in ‘b’ and ‘c’ are same as described in ‘a’. Information on linear-trends
945 and statistical significance for seasonal monsoon precipitation and frequency of heavy-
946 precipitation events are given in Table.1 and Table.4 respectively.

947

948 **Figure 2 | Spatial map of changes in the seasonal monsoon rainfall during (1951-2005) from**
949 **observations and high-resolution simulations (a)** APHRODITE **(b)** HIST1 **(c)** HIST2 **(d)**
950 IPSL-CM5A-LR. The monsoon rainfall changes are computed at each grid-cell using least-
951 square linear trends and expressed as $\text{mm day}^{-1}(\text{55 years})^{-1}$. Only values that exceed the 95%
952 confidence level, based on a student’s t-test, are displayed.

953

954 **Figure 3 | Time-series of interannual variations in the seasonal monsoon precipitation (mm**
955 **day^{-1}) averaged over land-points of the Indian domain ($70^{\circ}\text{E}-90^{\circ}\text{E}$, $10^{\circ}\text{N}-28^{\circ}\text{N}$). The time-**
956 **series is shown for the period 1951-2005 based on IMD observations (black line), HIST1 (brown**
957 **solid line), HIST2 (brown dashed line), HISTNAT1 (blue solid line) and HISTNAT2 (blue**
958 **dashed line). Linear least-square trends and their statistical significance are given in Table. 1.**

959

960 **Figure 4 | Temporal variation of LOESS trends in the seasonal monsoon rainfall averaged**
961 **over the land-points of the Indian region ($70^{\circ}\text{E}-90^{\circ}\text{E}$, $10^{\circ}\text{N}-28^{\circ}\text{N}$). Using the locally**
962 **weighted polynomial regression method (LOESS), time-varying trends are estimated for**

963 standardized rainfall time-series based on mean and standard deviation values for the period
964 1951–2005. The LOESS trends are shown for IMD observations (black line), CMIP5 ensemble
965 mean (grey solid line), HIST1+RCP4.5 (red solid line), HISTNAT1 (blue line), IPSL-CM5
966 (green lines). Local standard-deviations, estimated from the trends of the CMIP5 population, are
967 displayed by the grey dashed lines. Contrary to the coarse-resolution CMIP5 and IPSL-CM5
968 models, note that the (HIST1+RCP4.5) high-resolution exhibits a prominent declining trend
969 starting from 1940s. Also note that the LOESS trends from (HIST1+RCP4.5) clearly lie outside
970 the range of the CMIP5 trends.

971
972 **Figure 5 | Attributing the monsoon weakening to anthropogenic influence** Map showing the
973 difference in June-September mean precipitation (mm day^{-1}) and 850 hPa winds (vectors; ms^{-1})
974 between the HIST1 and HISTNAT1 simulations for the period (1951-2005). Grey dots
975 correspond to mean precipitation differences (HIST1 minus HISTNAT1) which exceed 95%
976 confidence level based on a two-tailed student's t-test.

977
978 **Figure 6 | Decomposing the monsoonal response to GHG and regional forcing elements**
979 Composite difference maps of the simulated June-September precipitation (mm day^{-1}) and 850
980 hPa winds (ms^{-1}) (a) $\delta(\text{No_GHG}) = \text{HIST1} - \text{HIST1_GHG}$ (b) $\delta(\text{GHG}) = \text{HIST1_GHG}$
981 minus HISTNAT1 (c) $\delta(\text{GHG_Atmos}) = \text{HIST1} - \text{HIST1_PIGHG}$ (d) $\delta(\text{GHG_SST}) =$
982 $\delta(\text{GHG}) - \delta(\text{GHG_Atmos})$. The composite maps are constructed for the period (1951-
983 2000) using the decadal time-slices.

984
985 **Figure 7 | Temporal evolution of regional=forcing elements and simulated response during**
986 **1886-2095** (a) Tree-fraction (%) in green, Crop-fraction (%) in brown and Planetary albedo in
987 grey (b) Equatorial IOSST anomaly ($^{\circ}\text{C}$) used in HIST1 + RCP4.5 (brown line) and HadISST
988 (1886-2005) (black line). The grey line is the anthropogenic-aerosol radiative forcing (Wm^{-2}) at
989 the top-of-atmosphere (TOA) (c) Sea level pressure (hPa) from HadSLP (1886 – 2005) (black
990 line) and simulations (d) Simulated precipitable water (kg m^{-2}) (e) Vertical shear of zonal winds
991 ($\text{U}_{200} - \text{U}_{850}$) in ms^{-1} from the LMDZ4 simulations and NCEP reanalysis (f) Simulated
992 soil moisture (mm). The SST anomalies are averaged for the equatorial region ($60^{\circ}\text{E}-90^{\circ}\text{E}$, $5^{\circ}\text{S}-$
993 5°N ; see inset in panel 'b') and other variables are averaged over the region ($70^{\circ}\text{E}-90^{\circ}\text{E}$, $10^{\circ}\text{N}-$

994 28°N; see inset in panel ‘d’). The brown lines in panels ‘c’, ‘d’, ‘e’, and ‘f’ correspond to the
995 HIST1 + RCP4.5 experiments and the blue line correspond to HISTNAT1.

996
997 **Figure 8 | Spatial maps of land-use used in the LMDZ4 experiments.** (a) Mean tree-fraction
998 (%) for the period 1951-2000 (b) Same as ‘a’ except for crop-fraction (%) (c) Change in tree-
999 fraction (%) shown by difference [(1891-1930) minus (1951-2000)] map (d) Same as ‘c’ except
1000 for crop-fraction (%). Note the larger spatial coverage of tree area over South and Southeast
1001 Asia and China during (1891-1930) relative to (1951-2000); while the crop area coverage was
1002 less during (1891-1930) relative to (1951-2000).

1003
1004 **Figure 9 | Spatial distribution of mean anthropogenic aerosol forcing from the HIST1**
1005 **experiment during 1951-2005** (a) Anthropogenic aerosol forcing (Wm^{-2}) at the top-of-
1006 atmosphere (TOA) (b) Atmospheric absorption (Wm^{-2}) due to anthropogenic aerosols [i.e.,
1007 aerosol-forcing @ TOA minus aerosol-forcing @ Surface]. The mean aerosol forcing is
1008 computed for the JJAS season from the HIST1 simulation during the period 1951-2005.

1009
1010 **Figure 10 | Tropical Indian Ocean SST warming trend during (1951-2005)** (a) Spatial
1011 pattern of linear trend of SST ($^{\circ}C$ per 55 years) from the IPSL-CM5A-LR simulation (b) Time-
1012 series of equatorial Indian Ocean SST (IOSST in $^{\circ}C$) anomalies averaged over the region ($5^{\circ}S$ -
1013 $5^{\circ}N$, $60^{\circ}E$ - $90^{\circ}E$) from HadISST (black line), IPSL-CM5A-LR (green line), IPSL-CM5A-MR
1014 (purple), ensemble mean of CMIP5 models (red line). The grey shading shows the spread of SST
1015 anomalies simulated across the CMIP5 models (c) Time-series of IOSST δ GM anomalies ($^{\circ}C$)
1016 (IOSST δ GM = EQIOSST minus Global Mean SST) for HadISST (black line), IPSL-CM5A-LR
1017 (green line). The rapid warming of IOSST δ GM is apparently linked to weakening of the
1018 summer-monsoon cross-equatorial flow in recent decades (Swapna et al. 2014).

1019
1020 **Figure 11 | Attributing changes in moderate and heavy precipitation types to global and**
1021 **regional forcing.** Box-whisker plot of distributions of yearly count of moderate ($5\text{ mm day}^{-1} \leq$
1022 precipitation intensity $< 100\text{ mm day}^{-1}$) and heavy (precipitation intensity $\geq 100\text{ mm day}^{-1}$)
1023 events over Central India ($74.5^{\circ}E$ - $86.5^{\circ}E$, $16.5^{\circ}N$ - $26.5^{\circ}N$) during the period (1951-2000) from
1024 HIST1 and HIST1_GHG, expressed as percentage departure relative to HISTNAT1. Note that,

1025 for each year the events are counted over 750 grid-cells in the Central India domain and 122 days
1026 of the JJAS season. Year-wise departures in frequency counts are first calculated for both
1027 precipitation categories relative to HISTNAT1. The quartiles are then computed from year-wise
1028 counts for the two precipitation categories in HIST1 and HIST1_GHG. Note that HIST1 displays
1029 an opposite change for the moderate (-) and heavy (+) precipitation categories, but HIST1_GHG
1030 shows positive changes for both categories. For changes in the heavy-precipitation category, the
1031 third quartile in HIST1 is significantly higher by 10% as compared to HIST1_GHG, thus
1032 highlighting the influence of regional forcing on amplifying precipitation extremes.

1033

1034 **Figure 12 | Latitude-pressure sections showing difference maps of meridional overturning**
1035 **anomalies.** Streamlines are constructed using meridional wind (v in ms^{-1}) and vertical velocity
1036 ($\omega \times 150 \text{ hPa s}^{-1}$) averaged over the $70^{\circ}\text{E} - 90^{\circ}\text{E}$ longitudinal band. The anomalies of ω shaded,
1037 such that negative (positive) values correspond to anomalous upward (downward) motions **(a)**
1038 HIST1 (1951 – 2005) minus HISTNAT1 (1951-2005) **(b)** RCP4.5 (2006-2050) – HIST1 (1951
1039 – 2005).

1040

1041 **Auxiliary Figure A1 | Time-series of year-wise count of heavy rainfall events (intensity \geq**
1042 **100 mm day⁻¹) over Central India ($74.5^{\circ}\text{E} - 86.5^{\circ}\text{E}$, $16.5^{\circ}\text{N} - 26.5^{\circ}\text{N}$).** The counts are for the
1043 June-September monsoon season from 1951-2005 based on IMD observations (black line),
1044 HIST1 (brown solid line), HIST2 (brown dashed line), HISTNAT1 (blue solid line) and
1045 HISTNAT2 (blue dashed line). The linear least-square trends and their statistical significance are
1046 presented in Table. 4.

1047

1048 **Auxiliary Figure A2 | Difference maps of precipitation (mm day^{-1} , shaded) and 850 hPa**
1049 **winds (ms^{-1} , vectors) (a) RCP4.5 minus HISTNAT1 (b) RCP4.5 minus HIST1.** The mean of
1050 RCP4.5 is for the period 2006-2060. For HIST1 and HISTNAT1, the means are for the period
1051 1951-2005. Note the persistence of weak SAM circulation and rainfall anomalies in the RCP4.5
1052 projection.

1053

1054 **Auxiliary Figure A3 | Tropospheric temperature (TT) and circulation response to**
1055 **anthropogenic influence:** Map showing the difference in JJAS mean of TT ($^{\circ}\text{C}$) and

1056 tropospheric circulation (vectors: ms^{-1}) between HIST1 and HISTNAT1 for the period (1951-
1057 2005). The temperature and wind fields are vertically averaged between 600 and 200 hPa. Note
1058 that the TT response over the near-equatorial areas is warmer as compared to that of the extra-
1059 tropics (poleward of 30°N). The cyclonic circulation anomaly over West-Central Asia is
1060 associated with cold air advection and subsidence over the Indian subcontinent. The anticyclonic
1061 circulation anomaly over the Indian region indicates weakening of the SAM circulation.

1062

1063 **Auxiliary Figure A4 | Climatological mean monsoon rainfall and 850 hPa winds from**
1064 **observations/reanalysis, LMDZ4 high-resolution simulations, IPSL-CM5A models. a,**
1065 **GPCP and NCEP b, HIST1 c, HIST2 d, IPSL-CM5A-MR e, IPSL-CM5A-LR.** The means are
1066 for the period 1951-2005, except for GPCP rainfall which is for the period 1979-2009. Notice the
1067 severe underestimation of monsoon winds and precipitation, particularly over the Western Ghats
1068 in the IPSL-CM5A models.

1069

1070 **Auxiliary Figure A5 | Coupled variability of monsoon precipitation and low-level winds in**
1071 **observations and simulations.** The first empirical orthogonal function (EOF1) of JJAS
1072 precipitation over western Ghats and west-central peninsular India for the period 1941-2005
1073 from (a) Observations (b) HIST1 (c) IPSL-CM5-LR (d, e, f) corresponding principal component
1074 (PC1) time-series (g, h, i) Pattern obtained by regressing the 850 hPa winds over the Arabian Sea
1075 upon the PC1 time-series of rainfall. Note the decreasing trend of PC1 time-series in
1076 observations and HIST1 high-resolution simulation, but not in the IPSL-CM5-LR model.
1077 Consistent with the decreasing trend of PC1, the regression pattern of westerly winds indicate
1078 weakening of the monsoon flow in NCEP reanalysis and HIST1. In contrast, note that the wind
1079 variations in the IPSL-CM5-LR are anti-correlated with the increasing trend of PC1 time-series
1080 as seen from the easterly anomaly.

1081

1082 **Auxiliary Figure A6 | Spatial map of projected future changes in the seasonal monsoon**
1083 **rainfall.** Least-square linear trend of June-September monsoon rainfall from the RCP4.5
1084 simulation expressed as $\text{mm day}^{-1} (45 \text{ years})^{-1}$ (a) (2006 – 2050) (b) (2051 – 2095). Only
1085 values exceeding the 95% confidence level are displayed.

1086

Figure 1

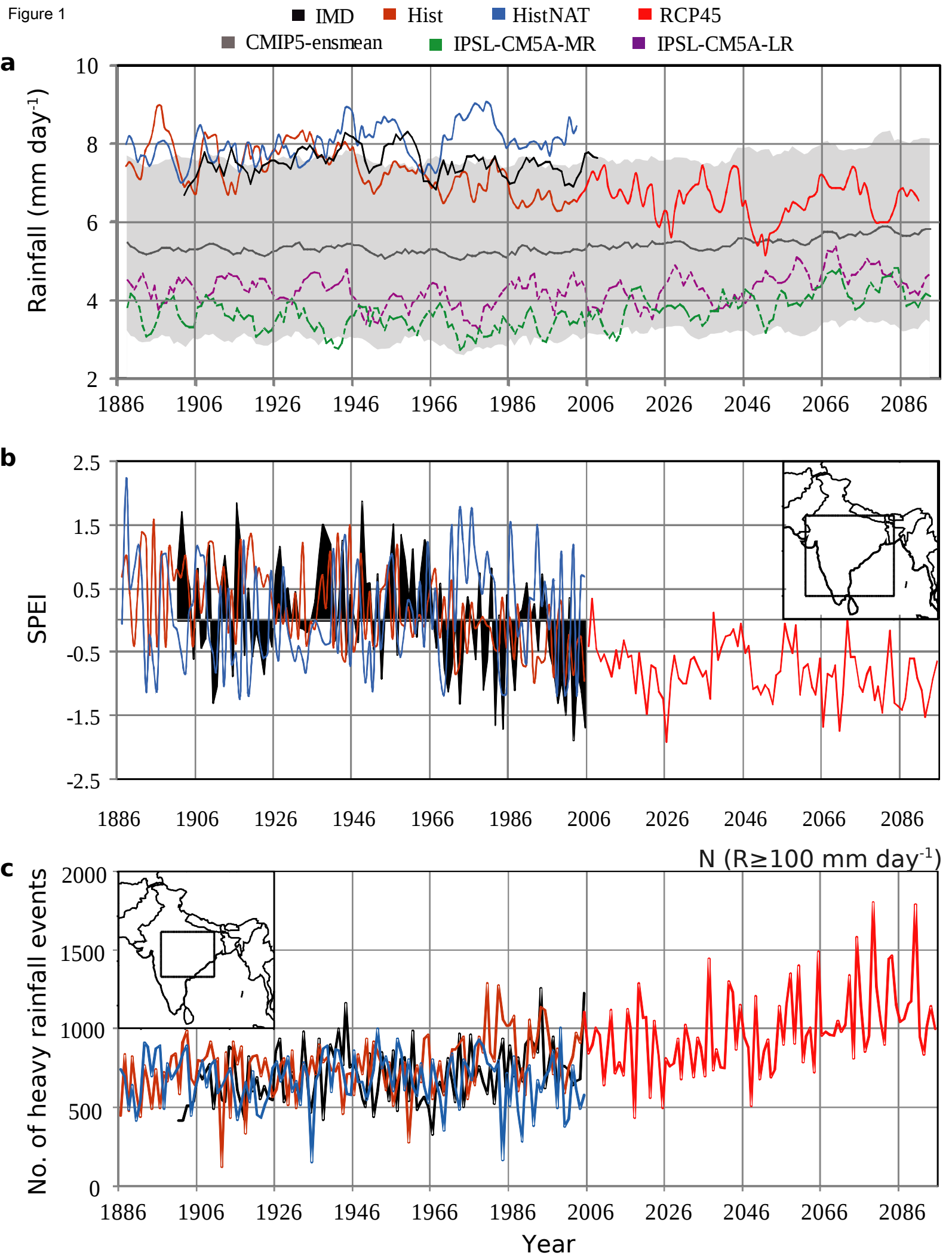


Figure 2

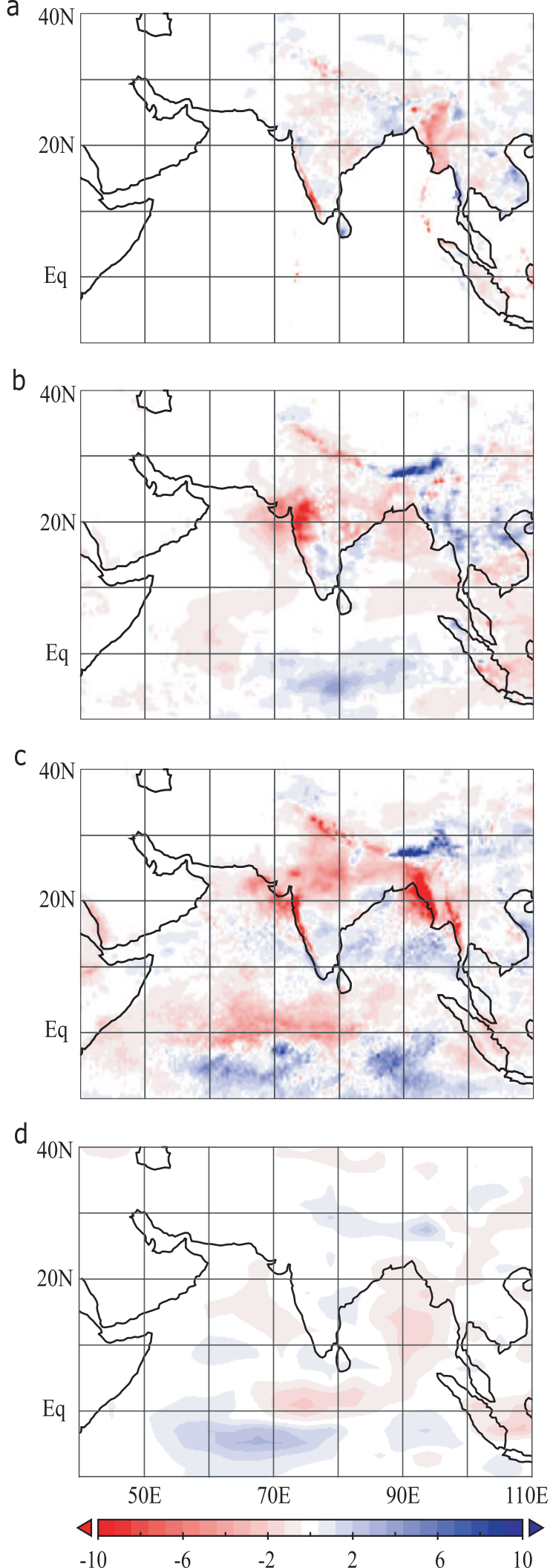


Figure 3

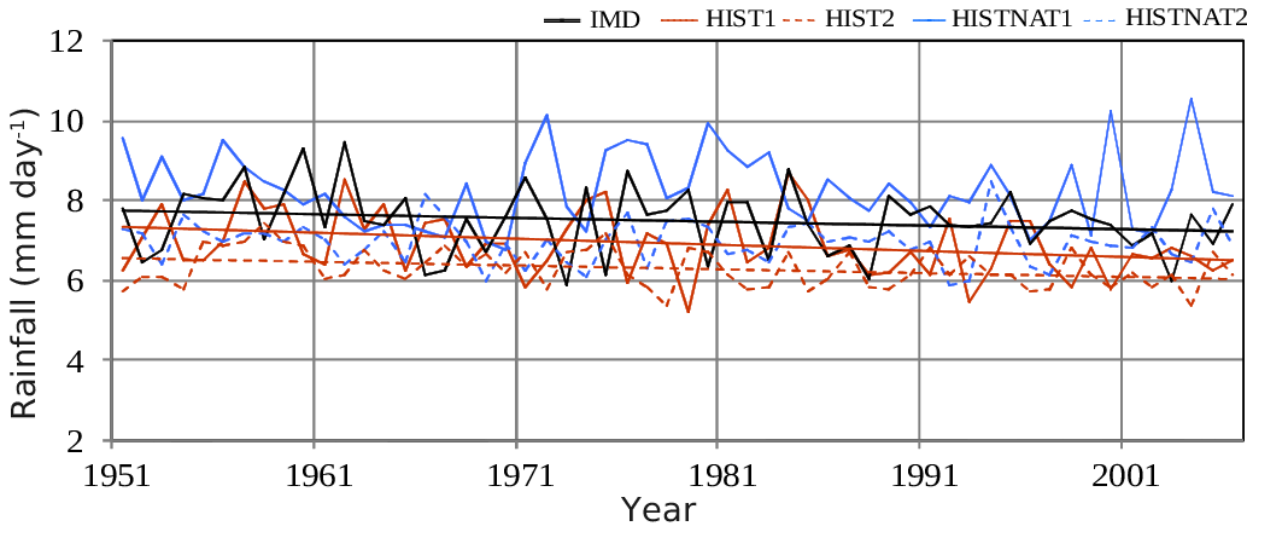


Figure 4

Monsoon rainfall trend from standardized time series - 1860-2095

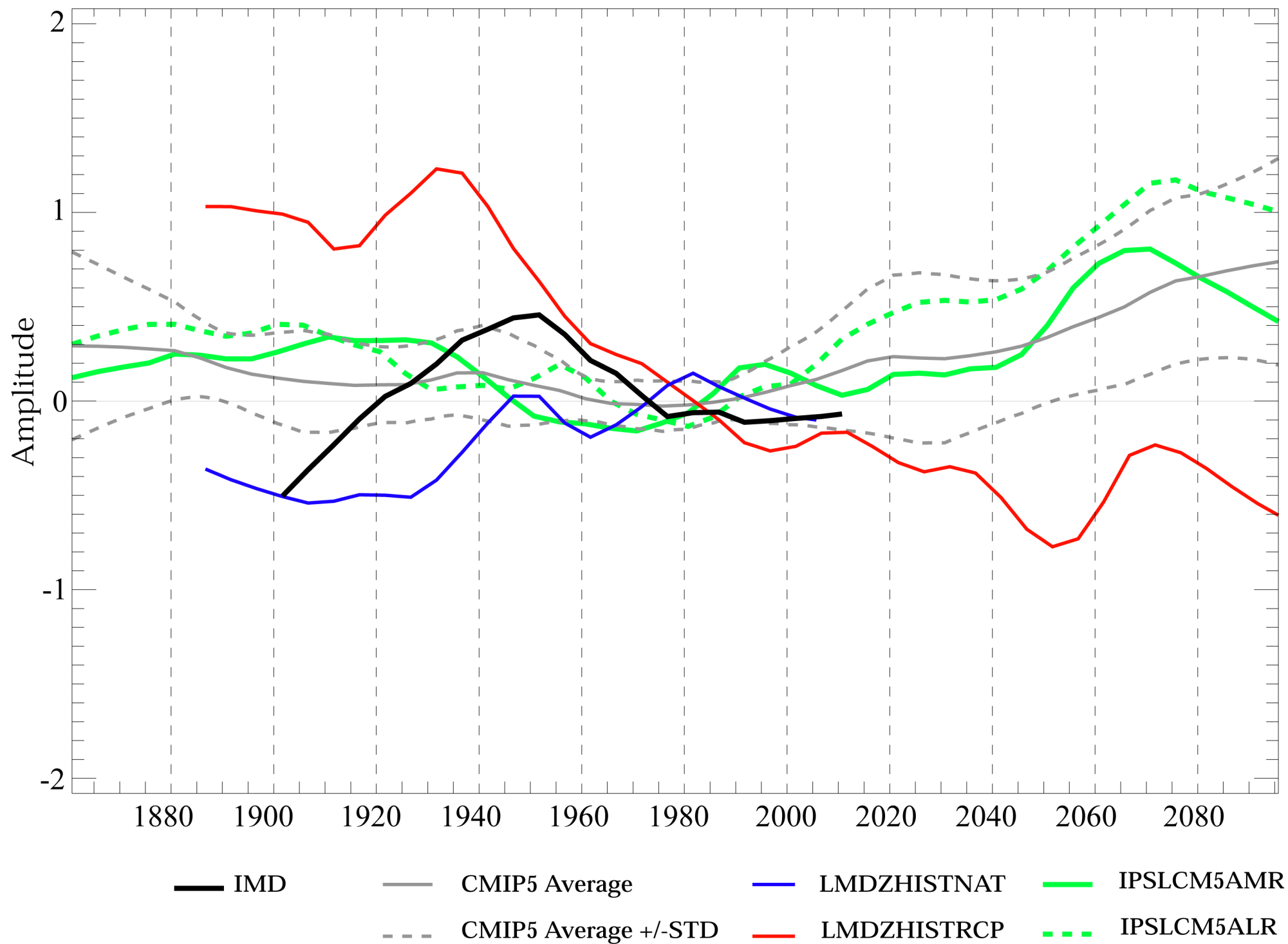


Figure 5

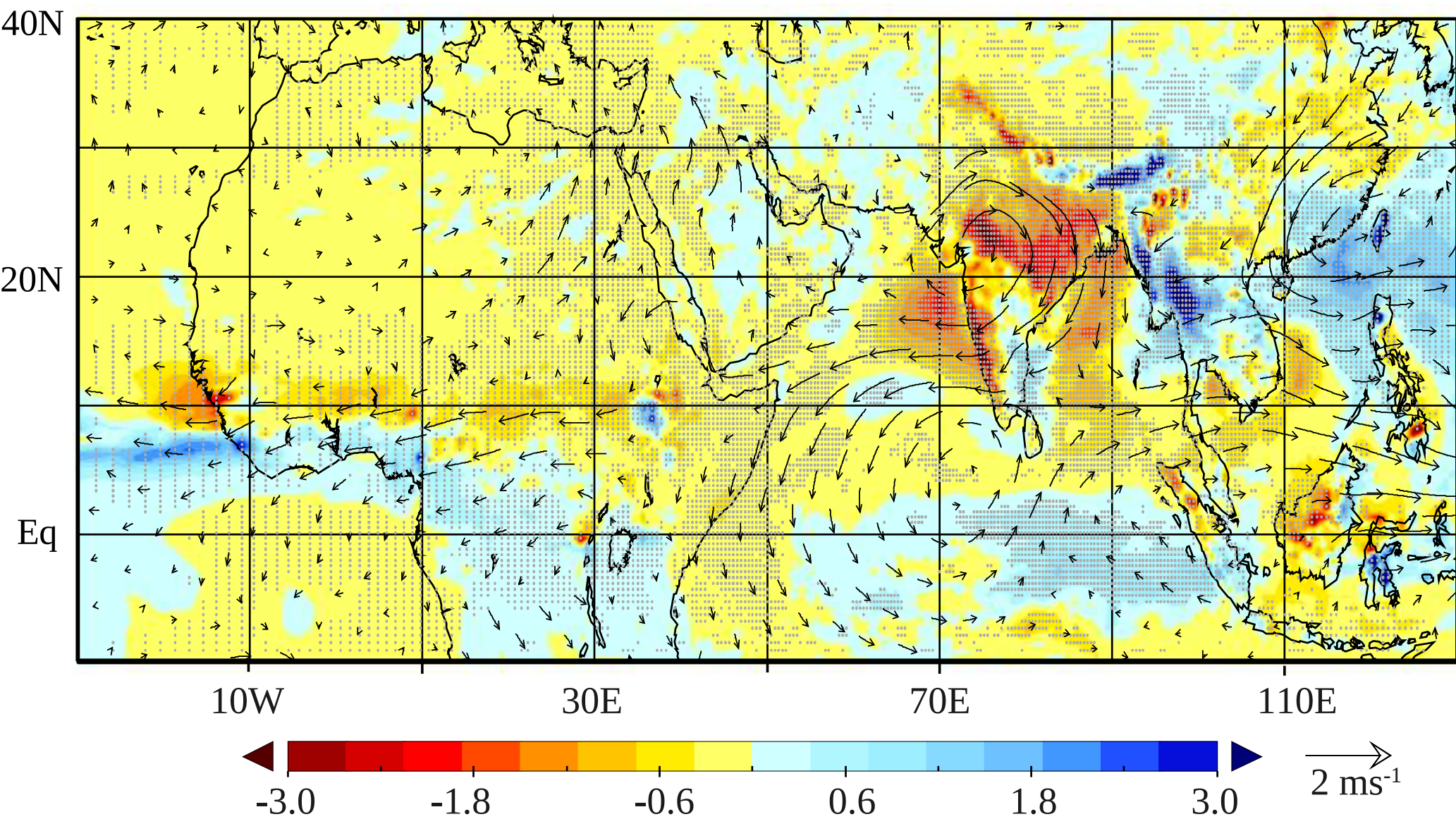


Figure 6

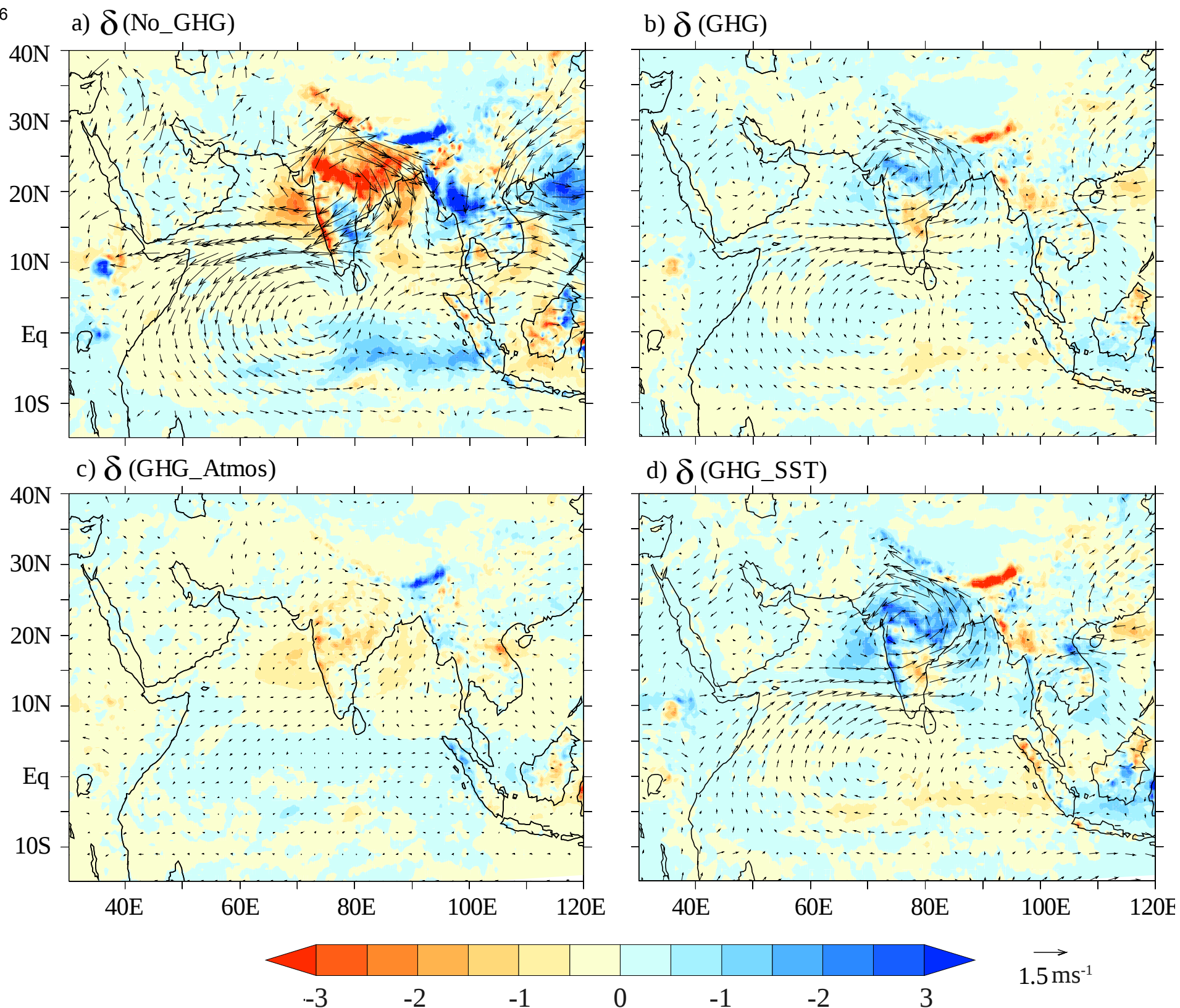


Figure 7

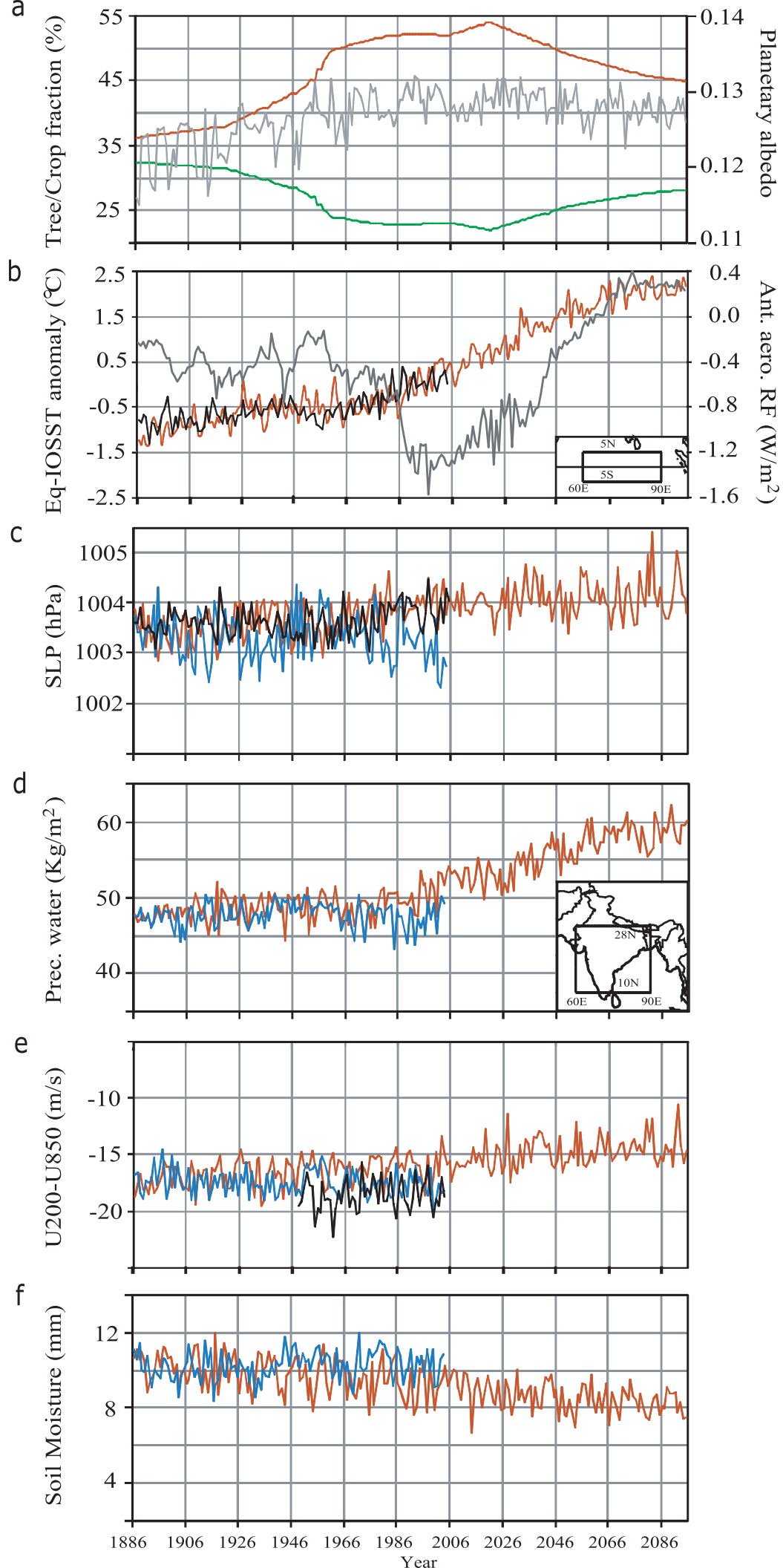


Figure 8

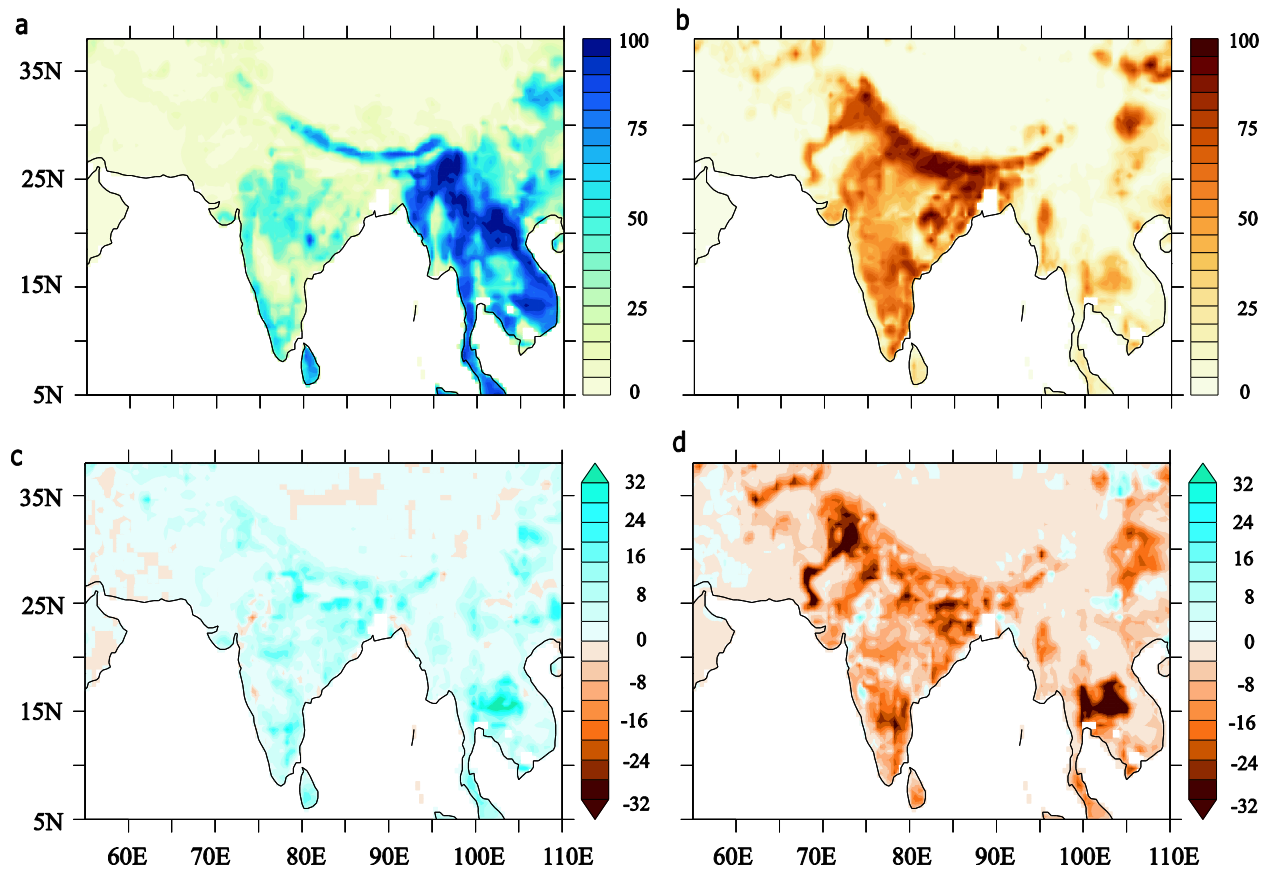
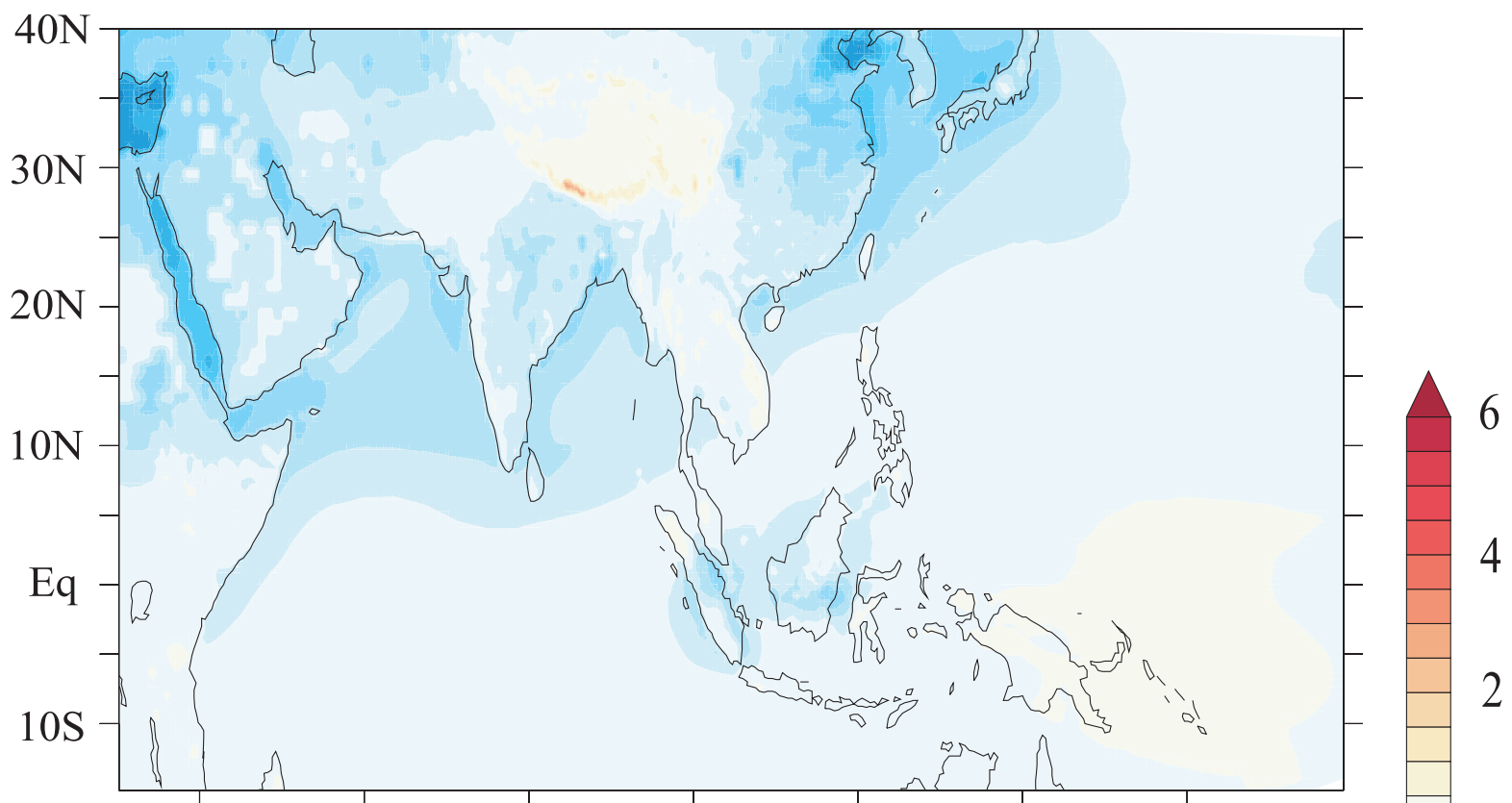


Figure 9

a



b

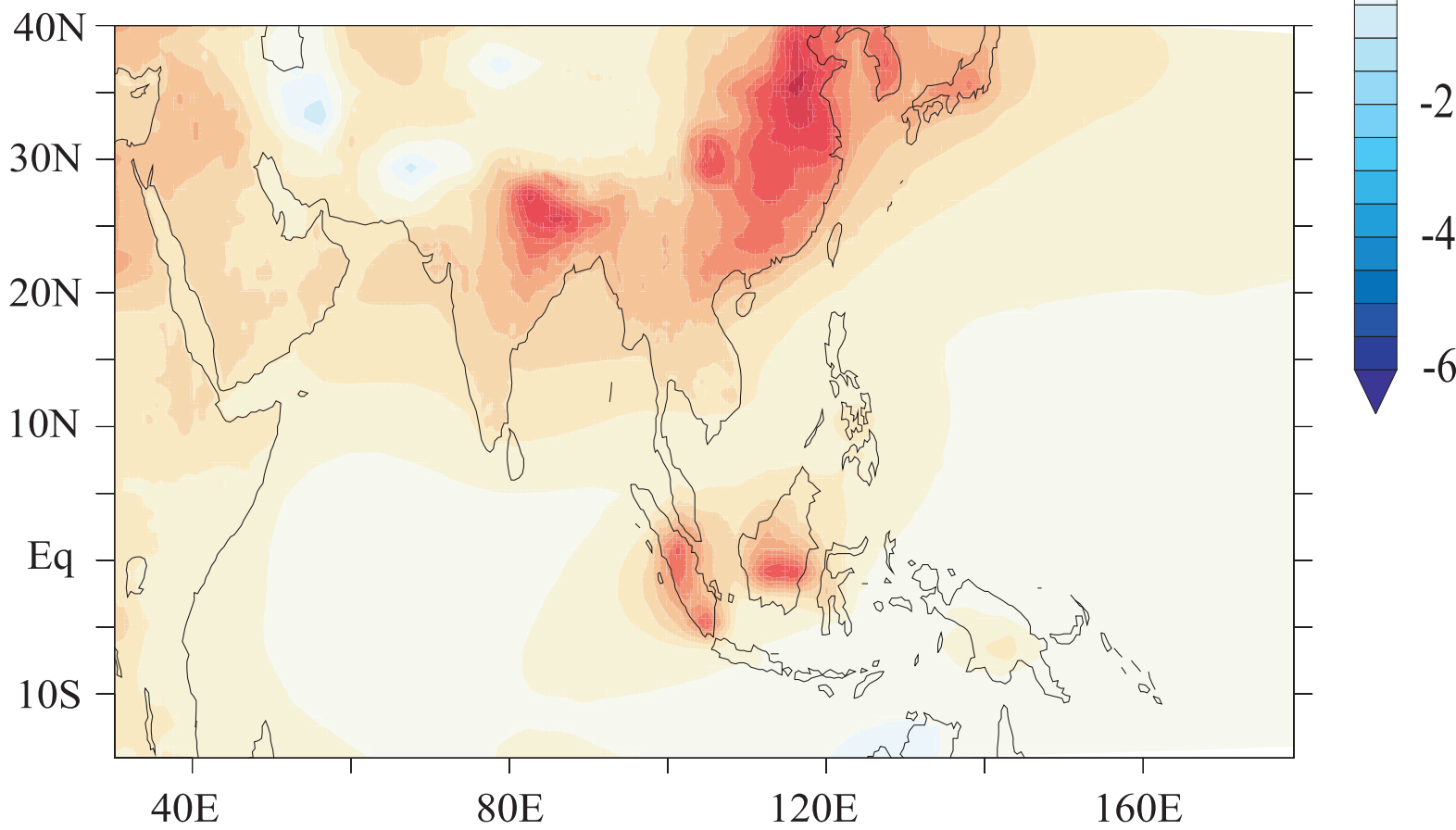


Figure 10

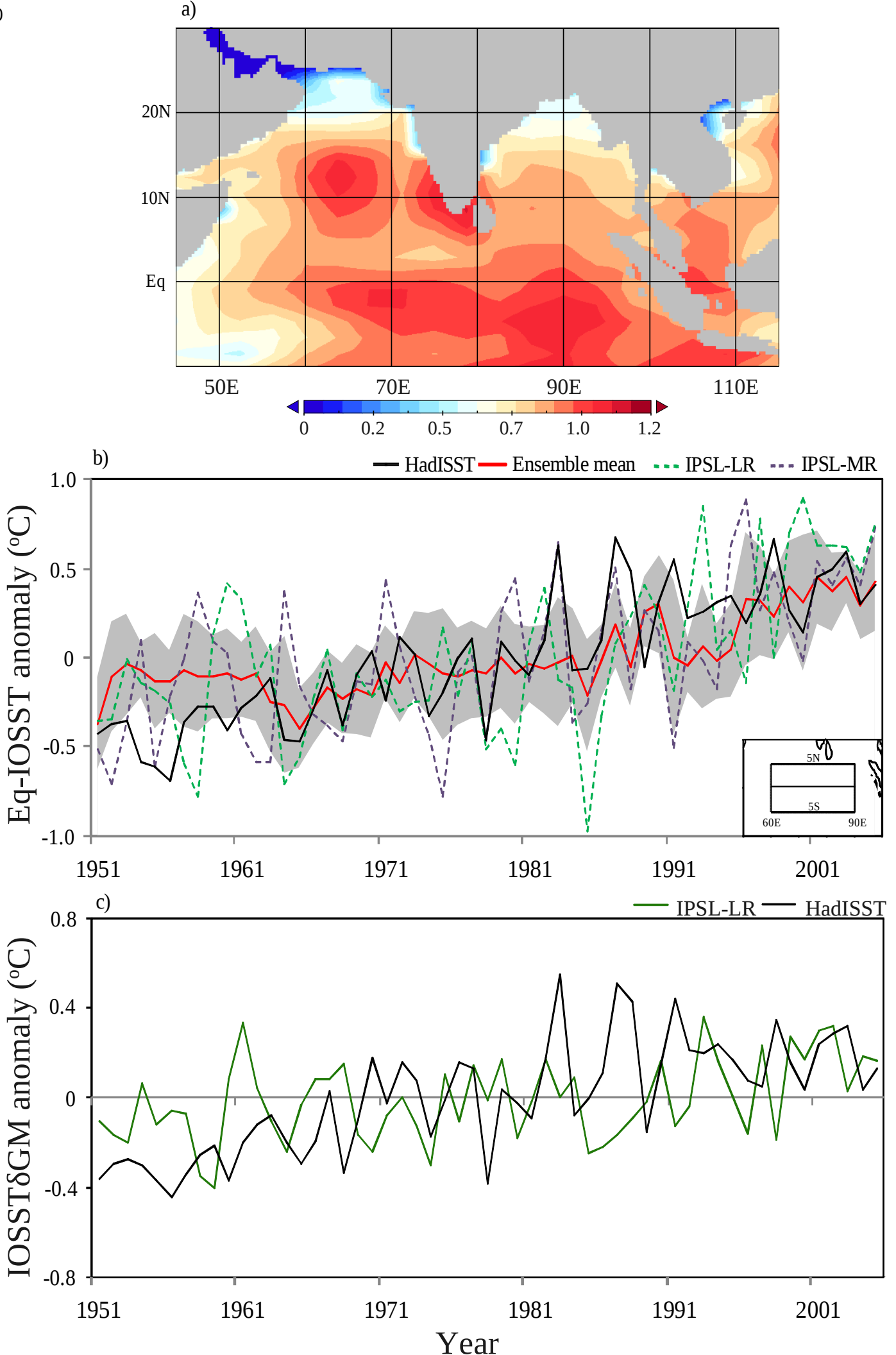


Figure 11

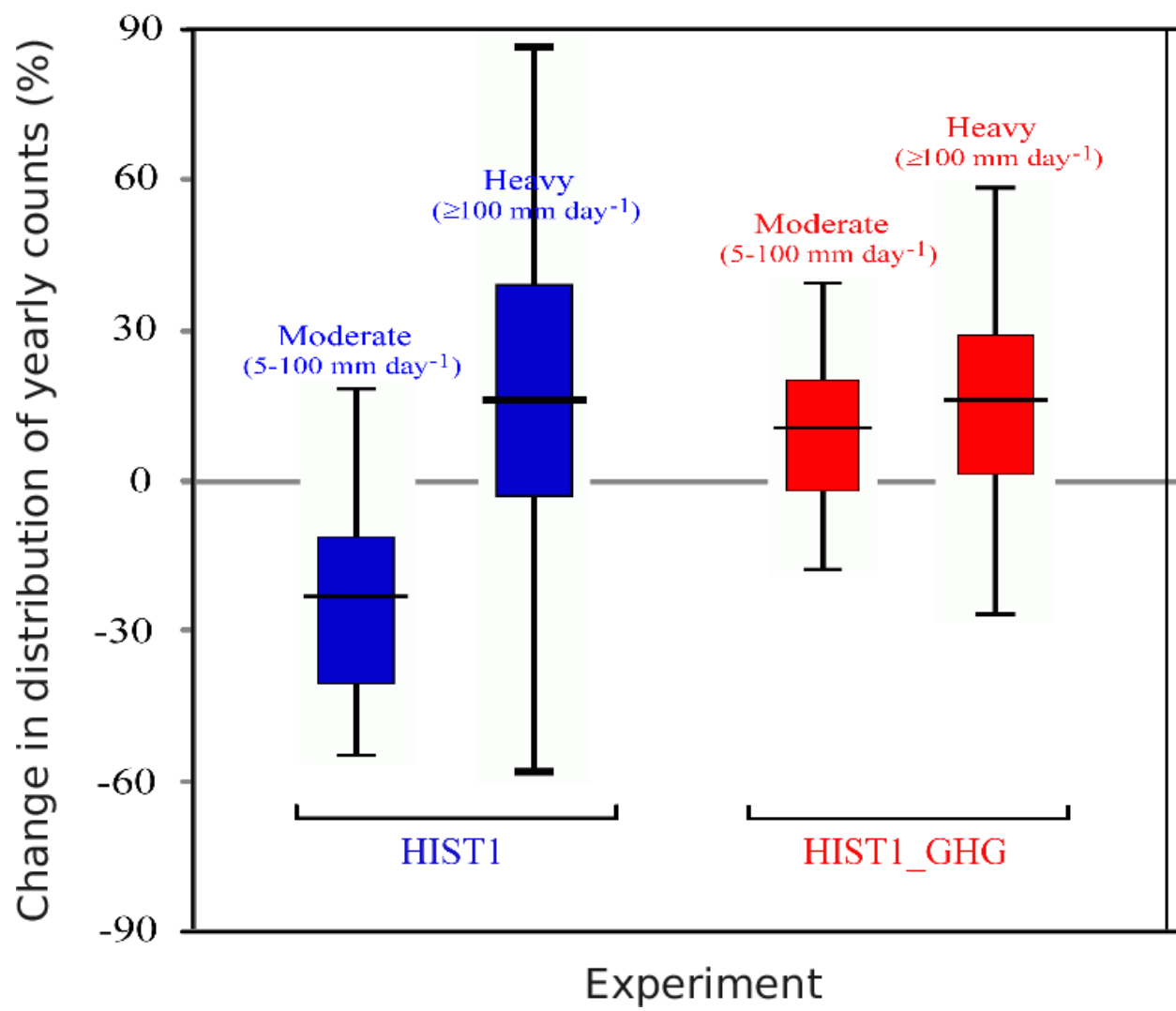
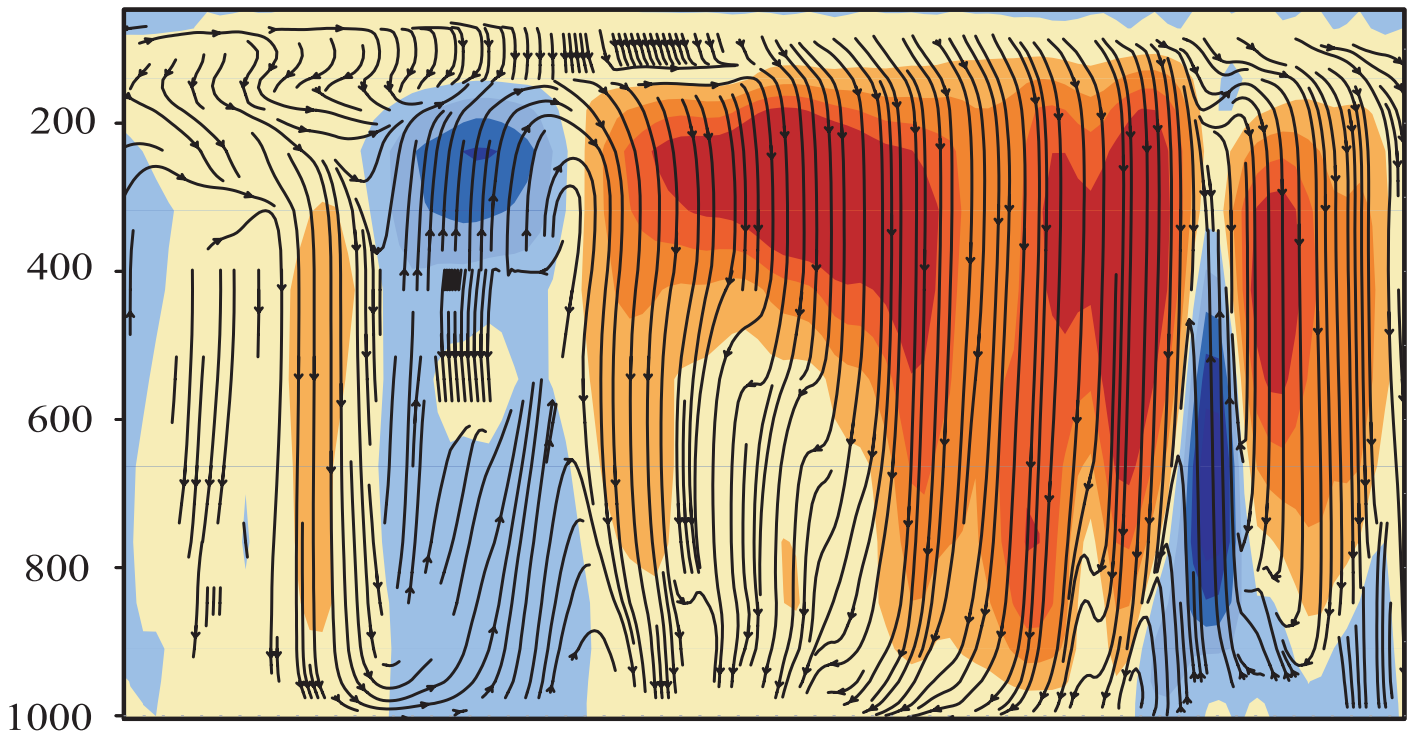
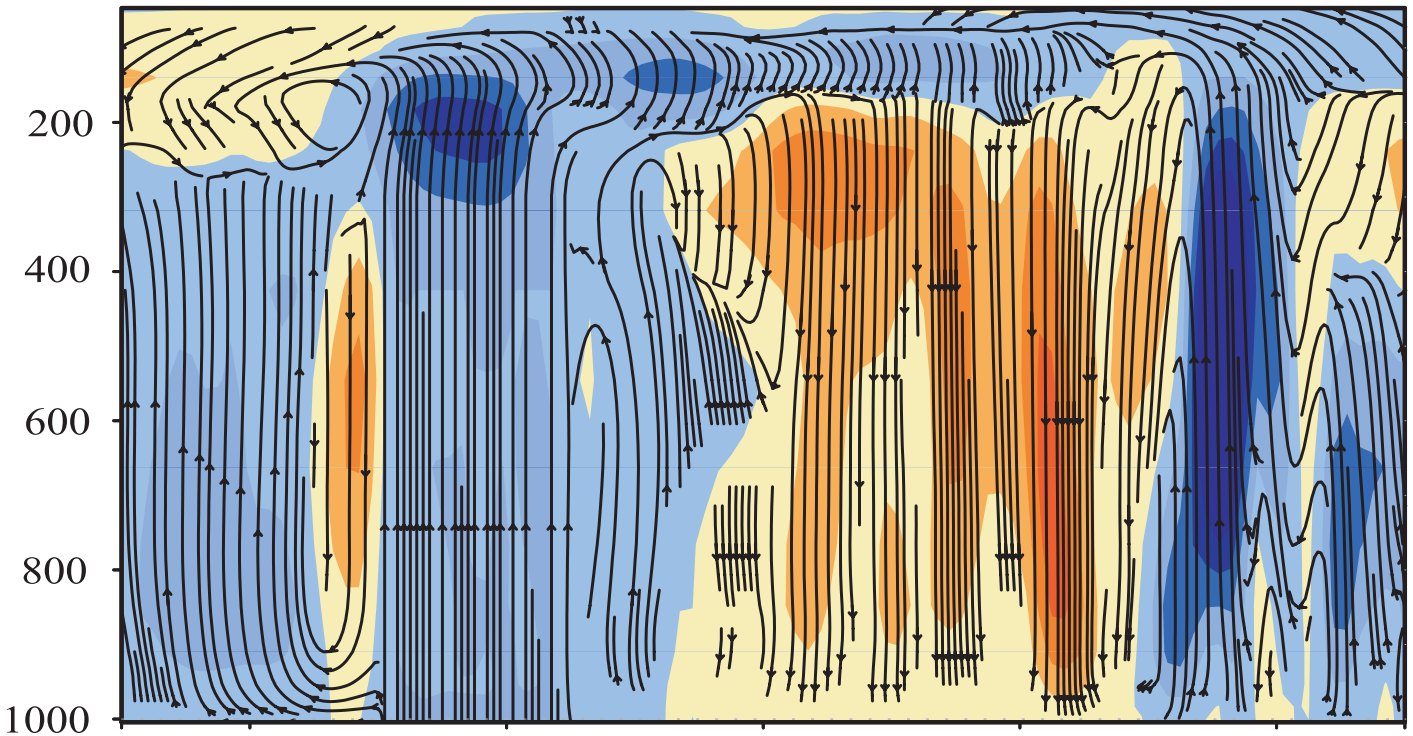


Figure 12

a) Hist1-HistNAT1



b) RCP4.5-Hist1



10S

Eq

10N

20N

30N

Latitude

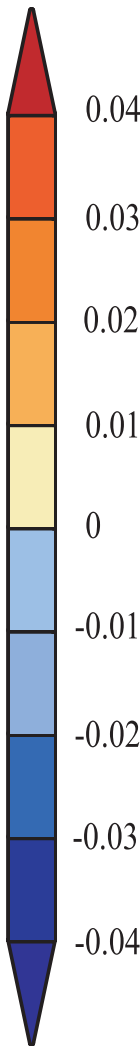


Table.1. Summary of trends of JJAS rainfall averaged over the land points for the Indian region (70-90°E, 10-28°N). A 5-year running mean has been applied on the rainfall time-series.

	Rainfall trend	Mean rainfall (mm day⁻¹)	% change w.r.t mean rainfall	P value based on two tailed student's t-test
IMD dataset (1951-2005)	-0.55 units mm day ⁻¹ (55 years) ⁻¹	7.5	-7%	P < 0.01
HIST1 (1951 – 2005)	-1.1 units mm day ⁻¹ (55 years) ⁻¹	6.9	-16%	P < 0.01
HIST2 (1951-2005)	-0.55 units mm day ⁻¹ (55 years) ⁻¹	6.3	-9%	P < 0.01
HISTNAT1 (1951-2005)	-0.03 units mm day ⁻¹ (55 years) ⁻¹	8.3	-0.3%	P = 0.54 (not significant)
HISTNAT2 (1951-2005)	- 0.1 units mm day ⁻¹ (55 years) ⁻¹	6.9	-1%	P = 0.2 (not significant)
RCP4.5 (2006-2060)	-1.1 units mm day ⁻¹ (55 years) ⁻¹	6.6	-17%	P < 0.01
RCP4.5 (2006-2095)	-0.29 units mm day ⁻¹ (90 years) ⁻¹	6.6	-5%	P < 0.01

Table.2. Summary of the LMDZ4 experimental design

Expt.	Period	Forcing	Cumulus convection	SST forcing
HIST1	Historical: (1886 – 2005)	Natural and Anthropogenic forcings	Emanuel	SST_ANOM_IPSL_CM5A_HIST + SST_AMIP_CLIM
HISTNAT1	Historical: (1886 - 2005)	Natural only	Emanuel	SST_ANOM_IPSL_CM5A_HISTNAT + SST_AMIP_CLIM
HIST2	Historical: (1950 – 2005)	Natural and Anthropogenic forcings	Tiedtke	SST_ANOM_IPSL_CM5A_HIST + SST_AMIP_CLIM
HISTNAT2	Historical: (1950 – 2005)	Natural only	Tiedtke	SST_ANOM_IPSL_CM5A_HISTNAT + SST_AMIP_CLIM
RCP4.5	Future RCP4.5 scenario (2006 – 2095)	Natural and Anthropogenic forcings	Emanuel	SST_ANOM_IPSL_CM5A_RCP4.5 + SST_AMIP_CLIM
HIST1_GHG	Historical (1950 – 2000) Decadal time slice runs for (1951-1960), (1961-1970), (1971-1980), (1981-1990), (1991-2000)	Natural and GHG-only forcings. Land use and aerosol fields are set to 1886 values	Emanuel	SST_ANOM_IPSL_CM5A_HIST_GHG + SST_AMIP_CLIM
HIST1_PIGHG	Historical: Decadal time slice runs for (1951-1960), (1961-1970), (1971-1980), (1981-1990), (1991-2000)	Includes Natural variations, Aerosol forcing and Land-use change. The concentration of GHGs are set to 1886	Emanuel	SST_ANOM_IPSL_CM5A_HIST + SST_AMIP_CLIM

Table.3. List of 21 CMIP5 models used in this study, their sponsor, country and name.

Sponsor and Country	Model name
Beijing Climate Centre Climate System Model, China	BCC-CSM1-1
Meteo-France / Centre National de Recherches Meteorologiques, France	CNRM-CM5
Centro Euro-Mediterraneo sui Cambiamenti Climatici, Italy	CMCC-CM
Commonwealth Scientific and Industrial Research Organisation (CSIRO), Australia	CSIRO-Mk3-6-0
Geophysical Fluid Dynamics Laboratory, National Oceanic and Atmospheric Administration (NOAA), USA	GFDL-CM3
	GFDL-ESM2M
	GFDL-ESM2G
National Aeronautics and Space Administration (NASA) / Goddard Institute for Space Studies (GISS), USA	GISS-E2H
	GISS-E2-R
	GISS-E2_R-CC
Met Office Hadley Centre, UK	HadGEM2-AO
	HadCM3
	HadGEM2-ES
	HadGEM2 CC
Institute for Numerical Mathematics, Russia	INMCM4
Institute Pierre Simon Laplace, France	IPSL-CM5A-MR
	IPSL-CM5A-LR
Centre for Climate System Research (University of Tokyo), National Institute for Environmental Studies and Frontier Research Center for Global Change (JAMSTEC), Japan	MIROC-ESM-CHEM
	MIROC5
	MIROC-ESM
Meteorological Research Institute, Japan	MRI-CGCM3

Table.4. Summary of trends in the frequency of heavy precipitation events over Central India, with intensities ≥ 100 mm day⁻¹, from IMD observations and LMDZ4 simulations

	Trend in the frequency count	Mean frequency count	% change w.r.t mean frequency count	P value based on the two tailed student's t-test
IMD dataset (1951-2005)	430 units (55 years) ⁻¹	1448	30%	P < 0.01
HIST1 (1951 – 2005)	499 units (55 years) ⁻¹	1652	30%	P < 0.01
HIST2 (1951-2005)	638 units (55 years) ⁻¹	1507	42%	P < 0.01
HISTNAT1 (1951-2005)	-34 units (55 years) ⁻¹	1356	-3%	P = 0.2 (not significant)
HISTNAT2 (1951-2005)	+6 units (55 years) ⁻¹	1233	0.5%	P = 0.8 (not significant)
RCP4.5 (2006-2095)	750 units (90 years) ⁻¹	1976	38%	P < 0.01

Table.5: Comparison of JJAS mean and standard deviation of rainfall and 850 hPa zonal winds between observations (GPCP / NCEP reanalysis) and models (LMDZ4, IPSL-CM5A-MR, IPSL-CM5A-LR). Rainfall averages are shown both for the Western Ghats (72°-76°E, 10°-19°N) and the Indian (70°-90°E, 10°-28°N) land regions. The 850 hPa zonal winds are averaged over a broader region (5°N – 22°N, 55°E-90°E) over India and the adjoining Arabian Sea and Bay of Bengal.

	GPCP rainfall / NCEP winds	LMDZ4	IPSL-CM5A-MR	IPSL-CM5A-LR
Mean & (std. dev) of rainfall over the Western Ghats	10.3 mm day ⁻¹ (1.6 mm day ⁻¹)	10.8 mm day ⁻¹ (3.3 mm day ⁻¹)	5.4 mm day ⁻¹ (1.6 mm day ⁻¹)	3.8 mm day ⁻¹ (1.1 mm day ⁻¹)
Mean & (std. dev) of rainfall over the Indian land region	7.6 mm day ⁻¹ (0.8 mm day ⁻¹)	6.5 mm day ⁻¹ (1.5 mm day ⁻¹)	4.3 mm day ⁻¹ (0.8 mm day ⁻¹)	4.5 mm day ⁻¹ (0.7 mm day ⁻¹)
Mean & (std. dev) of zonal wind	9.3 ms ⁻¹ (0.6 ms ⁻¹)	9 ms ⁻¹ (0.8 ms ⁻¹)	7.3 ms ⁻¹ (0.4 ms ⁻¹)	6.1 ms ⁻¹ (0.3 ms ⁻¹)