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1 **Chapter -8**

2
3 **Indian Ocean Dipole influence on Indian summer monsoon**
4 **and ENSO: A review**

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22

Abstract

23 The Indian Ocean Dipole (IOD) is one of the dominant modes of variability of the tropical
24 Indian Ocean and it has been suggested to have a crucial role in the teleconnection between
25 the Indian summer monsoon and El Niño Southern Oscillation (ENSO). The main ideas at
26 the base of the influence of the IOD on the ENSO-monsoon teleconnection include the
27 possibility that it may strengthen summer rainfall over India, as well as the opposite, and also
28 that it may produce a remote forcing on ENSO itself. In the future, the IOD is projected to
29 increase in frequency and amplitude with mean conditions mimicking the characteristics of
30 its positive phase. Still, state-of-the-art global climate models have large biases in
31 representing mean state and variability of both IOD and ISM, with potential consequences
32 for their future projections. However, the characteristics of the IOD and ENSO are likely to
33 continue in a future warmer world, with a persistence of their linkage.

34

35 **Keywords:** Indian Ocean Dipole, Indian summer monsoon rainfall, ENSO, remote forcing,
36 air-sea coupling, coupled climate models, biases, projections

37

38 **8.1 Introduction**

39 The Indian summer monsoon (ISM) is one of the main components of the South Asian
40 summer monsoon, representing the largest source of moisture and precipitation over the
41 tropical sector (Webster et al., 1998). The ISM is highly variable and its variability is partly
42 modulated by external factors, the El Niño Southern Oscillation (ENSO) being one of the
43 most important. The remote connection between ENSO and ISM is known since the
44 beginning of the nineteenth century and it has been largely investigated in the past (Walker,
45 1924; Sikka, 1980; Rasmusson and Carpenter, 1983; Kirtman and Shukla, 2000, among
46 others). Schematically, during warm ENSO episodes the rising limb of the Walker circulation
47 over West Pacific shifts eastward in response to a warming of the eastern Pacific, causing
48 descent of air to the west of it and aiding decreased monsoon rainfall over India (Goswami,
49 1998; Lau and Wang, 2006). It has been natural also to explore the possible influence of the
50 neighboring Indian Ocean on the ISM variability, with many studies pointing out significant
51 connections (Rao and Goswami, 1988; Ashok et al., 2001, 2004; Gadgil et al., 2004, 2005,
52 2007; Krishnan et al., 2003; Terray et al., 2005, 2007; Cherchi et al., 2007; Izumo et al.,
53 2008; Boschat et al., 2011, 2012; Cherchi and Navarra, 2013; Shukla and Huang, 2016,
54 among many others).

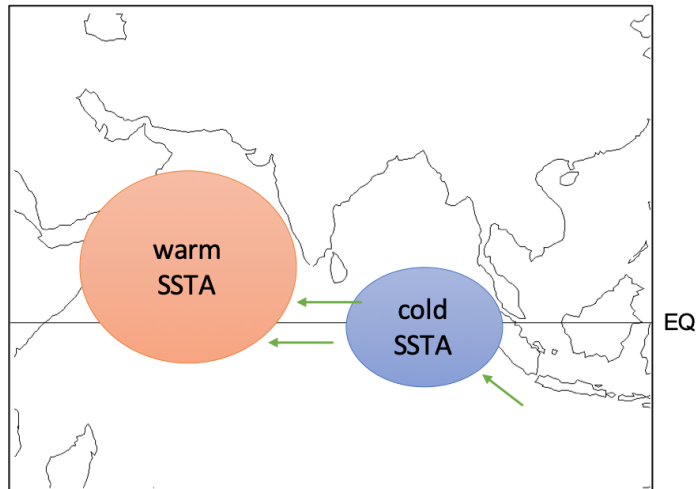
55 The Indian Ocean Dipole (IOD) was discovered at the end of the 90s (Saji et al.,
56 1999; Webster et al., 1999) and it is recognized as one of the dominant modes of variability
57 of the tropical Indian Ocean Sea Surface Temperature (SST). Toward the end of the 20th
58 century, weakening in the strength of the ENSO-monsoon relationship have been
59 documented (Kumar et al., 1999; Kinter et al., 2002) and, since its discovery, the IOD has
60 been identified as one potential element that modulates the ENSO-monsoon connection
61 (Ashok et al., 2001; Li et al., 2003). Contrasting literature about its active or passive role has

62 been produced since then and the debate is still open (Ashok et al., 2001; Li et al., 2003;
63 Meehl et al., 2003; Ashok et al., 2004; Wu and Kirtman, 2004; Cherchi et al., 2007; Krishnan
64 et al., 2011; Cherchi and Navarra, 2013; Krishnaswamy et al., 2015; Chowdary et al., 2015;
65 Srivastava et al., 2019, to mention a few).

66 This chapter intends to provide an updated review on the current understanding about
67 the influence of the IOD on the ISM and its teleconnection with ENSO. In particular, the
68 chapter is organized as follows: Section 8.2 is dedicated to the description of the IOD, while
69 Section 8.3 is focused on the processes at work in IOD influencing the monsoon and its
70 relationship with ENSO, also from a modelling point of view. Section 8.4 reviews the
71 literature about past evidence, present case studies and future projections about the topic, and
72 Section 8.5 is dedicated to the discussion of the results reviewed, highlighting some of the
73 associated challenges and related future perspectives. Finally, Section 8.6 collects the main
74 conclusion derived from the review.

75 **8.2 Some salient features of the Indian Ocean Dipole**

76 The IOD is characterized by a zonal dipole in the tropical Indian Ocean with positive SST
77 anomalies in the western equatorial Indian Ocean (50°-70°E, 10°S-10°N) and negative SST
78 anomalies toward Sumatra (90°-110°E, 10°S-EQ) in its positive phase (Fig. 8-1; Saji et al.,
79 1999). The formation of the IOD relies on the Bjerknes feedback, requiring background
80 surface easterlies and thermocline shallowing in the eastern part along the Equator (Fig. 8-1;
81 Schott et al., 2009). These conditions set in boreal spring and persist until autumn, explaining
82 the IOD development in boreal summer, its peak toward autumn and its rapid termination
83 before winter, because of the monsoon wind swing (Schott et al., 2009).



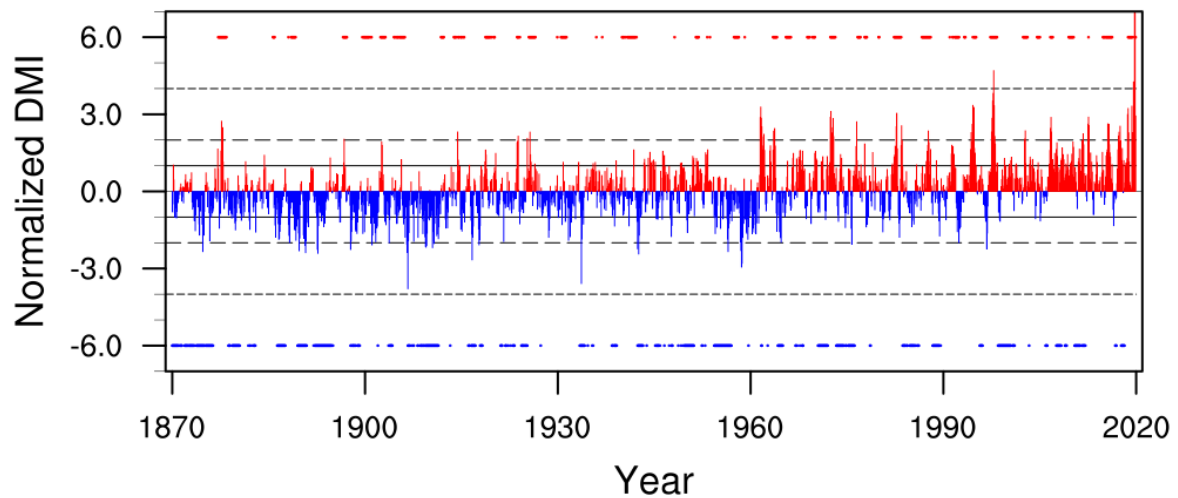
84

85 **Fig. 8-1:** Schematic of the Indian Ocean Dipole in its positive phase with warm SST anomalies on the
 86 west and cold SST anomalies toward the coast of Sumatra. The green arrows indicate the direction of the
 87 prevailing corresponding surface winds.

88

89 Before the discovery of the IOD as one of the dominant modes of variability of the
 90 tropical Indian Ocean, earliest suggestions of inherent coupled dynamics in the basin
 91 identified periods of anomalous easterlies in the central Indian Ocean concurrent with
 92 anomalous cold (warm) SST in the eastern (western) part, that occurred during boreal fall in
 93 the absence of ENSO (Reverdin et al., 1986). Significant feedback mechanisms at play
 94 between zonal SST and pressure gradient, equatorial easterly wind anomalies and
 95 precipitation anomalies in the western equatorial Indian Ocean were later identified
 96 (Hastenrath et al., 1993). An east-west seesaw in sea level anomaly in the tropical Indian
 97 Ocean was noticed and correlated with thermocline depth changes (Murtugudde et al., 1995).
 98 This east-west sea level dipole has been associated with the east-west SST gradient and
 99 strongly correlated with the Indian summer monsoon rainfall (Murtugudde et al., 1998). The
 100 SST gradient was shown to be in phase with equatorial zonal wind anomalies (Murtugudde
 101 and Busalacchi, 1999).

102 Since its discovery there has been strong debate and controversy whether the IOD is
103 an intrinsic mode of Indian Ocean coupled variability, or whether and how it is driven by
104 external forcing, like ENSO (e.g., Hastenrath, 2002; Dommenges and Latif, 2002; Yamagata
105 et al., 2003; Murtugudde et al., 2003; Behera et al., 2003). Many studies claim that most of
106 the IOD variability is driven by ENSO (Allan et al., 2001; Baquero-Bernal et al., 2002;
107 Huang and Kinter, 2002; Dommenges, 2011; Zhao and Nigam, 2015; Zhao et al., 2019),
108 while others suggest that IOD is a self-sustained mode of oscillation (Ashok et al., 2003;
109 Yamagata et al., 2004; Behera et al., 2006). About 20% of the IOD events seem to co-occur
110 with ENSO (Fig. 8-2; Saji, 2018). IOD events that could be categorized as with or without
111 the influence of ENSO have systematic differences in their temporal evolution and spatial
112 distribution, including periodicity, strength and formation processes (Behera et al., 2006;
113 Hong et al., 2008). Modeling studies confirm that IOD events are often triggered by ENSO,
114 but they also demonstrate that IOD events can exist without ENSO by means of dedicated
115 sensitivity experiments in which ENSO is removed by different nudging techniques (Lau and
116 Nath, 2004; Fischer et al., 2005; Behera et al., 2006; Wang et al., 2016; 2019; Cretat et al.,
117 2017, 2018). In the absence of ENSO, the interannual IOD variability is mostly biennial
118 (Behera et al., 2006; Cretat et al. 2018), while in years of co-occurrences ENSO affects the
119 periodicity, strength, and formation processes of IODs (Cretat et al. 2018).



120

121 **Fig. 8-2:** Normalized monthly IOD index (Standard deviation) defined as anomalous SST
 122 gradient between the western equatorial Indian Ocean (50°E-70°E and 10°S-10°N) and the southeastern
 123 equatorial Indian Ocean (90°E-110°E and 10°S-0°N). Anomalies (with respect to 1981-2010 mean) have
 124 been downloaded from https://psl.noaa.gov/gcos_wgsp/Timeseries/DMI/, but 2019 values have been
 125 integrated from JAMSTEC repository (<http://www.jamstec.go.jp/virtualearth/general/en/index.html>). Red
 126 and blue markers along 6 and -6 std correspond to NINO3.4 anomalies (1981-2010 mean removed) larger
 127 than 0.5°C. NINO3.4 anomalies have been downloaded from
 128 https://psl.noaa.gov/gcos_wgsp/Timeseries/Nino34/. As indicated in the respective websites, both IOD and
 129 NINO3.4 values are computed from the HadISST1 dataset (Rayner et al., 2003).

130

131 Through changes in the atmospheric circulation, the IOD exerts its influence on,
 132 among others, the Southern Oscillation (Behera and Yamagata, 2003), the summer climate
 133 condition in Europe (Behera et al., 2012), East Asia (Guan and Yamagata, 2003; Guan et al.,
 134 2003; Chen et al., 2019) and streamflows in the western part of Indonesia (Sahu et al., 2012),
 135 as well as on rainfall over Africa (Black et al., 2003; Manatsa and Behera, 2013; Endris et al.,
 136 2019), Sri Lanka (Zubair et al., 2003), Australia (Ashok et al., 2003; Ummenhofer et al.,
 137 2013; Dey et al., 2019; Hossain et al., 2020), and Brazil (Chan et al., 2008; Taschetto and

138 Ambrizzi, 2012; Bazo et al., 2013). In the following we focus on the IOD influence on the
139 summer monsoon rainfall over India and linkages with ENSO, which are both subject of
140 important controversies in the literature.

141 **8.3 IOD and the ENSO-monsoon teleconnections: processes at work**

142 To influence the ISM and its ENSO teleconnection, IOD events should develop from boreal
143 summer (June to August). As described by Schott et al. (2009), this is in fact the case for
144 many IOD events. However, some IODs peak in the peak monsoon season (Du et al., 2013),
145 though a few others develop later. According to that, strong interactions between IOD and
146 ISM can be expected during boreal summer and the withdrawal phase of the monsoon.

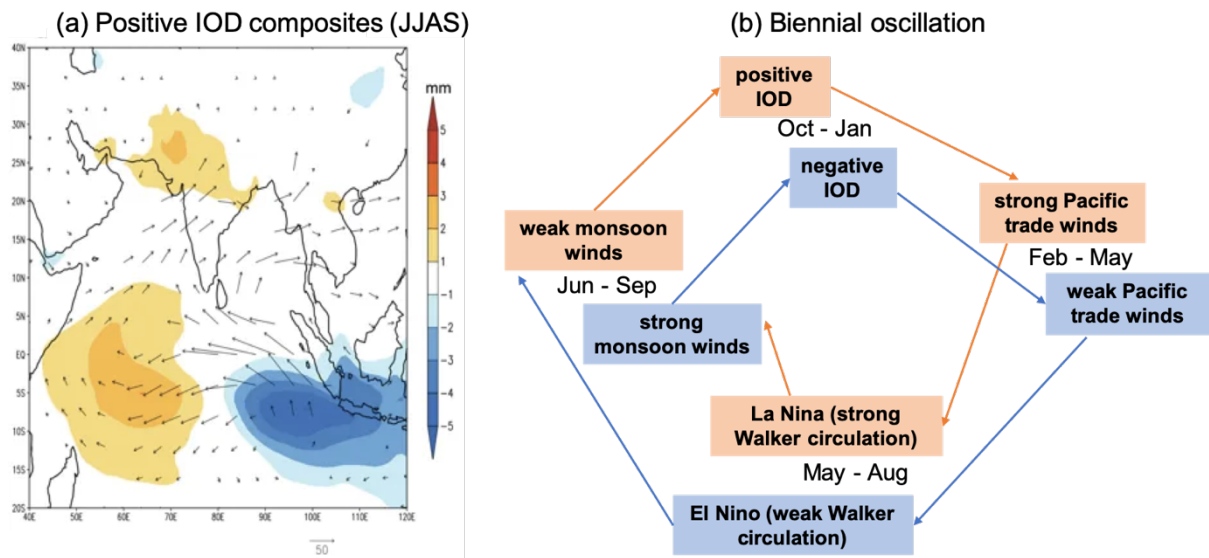
147 Some authors suggested a direct influence of the IOD on the ISM rainfall (ISM_R)
148 through moisture transport over the western Indian Ocean or modifications of the local
149 Hadley cell, with enhanced ascendance and a northward shift of its uplift branch during
150 positive IOD events, both enhancing ISM rainfall (Fig. 8-3a; Ashok et al., 2001, 2004;
151 Gadgil et al., 2004; Behera et al., 2005; Ashok and Saji, 2007; Behera and Ratman, 2018).
152 Recent investigations also reveal that early IODs (i.e., peaking in July) or prolonged IODs
153 (i.e., lasting longer, more than 8 months) have excess of evaporation from the Arabian Sea
154 and stronger cross-equatorial flow, leading to enhanced monsoon activity with decreased
155 numbers of break spells (Anil et al., 2016). Others suggest that positive IOD events during
156 boreal fall normally follow weak ISMs, and vice versa, as ISM circulation during boreal
157 summer can also induce an equatorial anomalous SST gradient in the Indian Ocean during
158 the following boreal fall (Loschnigg et al., 2003, Meehl et al., 2003; Terray et al., 2005,
159 2007). In this framework, ENSO, ISM and IOD appear as strongly inter-related components
160 of the Tropospheric Biennial Oscillation (TBO) in the tropics (Fig. 8-3b; Meehl and
161 Arblaster, 2002; Meehl et al., 2003; Li et al., 2006; Drbohlav et al., 2007; Webster and

162 Hoyos, 2010). The development of the IOD during boreal summer and autumn can lead to
163 SST warming in the tropical southwest Indian Ocean via ocean dynamics during the next
164 boreal winter and spring (Xie et al., 2002; Chowdary and Gnanaseelan 2007; Du et al., 2009;
165 Chowdary et al., 2009). This warming can further influence the ISM onset in the following
166 year, especially for IOD events co-occurring with El Niño in the Pacific Ocean and followed
167 by a basin-wide Indian Ocean warming (Annamalai et al., 2005; Yang et al. 2007; Hong et al.
168 2010).

169 The ISMR response is not necessarily spatially coherent to the IOD phases (Behera
170 and Ratman, 2018). The anomalous moisture transports to India associated with a positive
171 IOD strengthen the monsoon trough and rainfall through an intensified monsoon-
172 Hadley circulation (Behera et al., 1999; Ashok et al., 2001; Anil et al., 2016), with below
173 normal rainfall to the south and to the north of the trough. During positive IODs, the north-
174 south precipitation (heating) gradient over the eastern Indian Ocean dominates over the one
175 in the equatorial Indian Ocean, resulting in a regional meridional circulation with uplift over
176 the monsoon trough and sinking in the eastern lobe (Annamalai et al., 2003). On the other
177 hand, in a negative IOD event, a regional Walker circulation and the moisture distribution
178 favor moisture divergence (convergence) in the eastern (western) part of India. This gives
179 rise to a zonal dipole in the rainfall anomalies with abundant rainfall on the western part and
180 scanty rainfall on the east. The resulted regional asymmetry is a unique feature associated
181 with the ISMR response to IOD but it is not well simulated by coupled General Circulation
182 Models (CGCMs), though regional model experiments with different physical
183 parameterization schemes may provide few combinations able to realistically reproduce the
184 asymmetric response to the two phases of the IOD (Behera and Ratman, 2018).

185 IOD can influence the ENSO-ISM relationship indirectly (Ashok et al., 2001; Behera

186 and Yamagata, 2003; Cai et al., 2011; Weller and Cai, 2013), by modulating ENSO events in
 187 the tropical Pacific itself (Luo et al., 2010; Izumo et al., 2010; Cai et al., 2019). Considering
 188 IOD events as triggered by ENSO, they may counteract its simultaneous influence on ISM
 189 (Ashok et al., 2001, 2004; Ashok and Saji, 2007; Ummenhofer et al., 2011; Lau and Nath,
 190 2003, 2012; Krishnaswamy et al., 2015). More recently IOD has been suggested as potential
 191 trigger of ENSO, adding more complexity to the emerging picture (Luo et al., 2010; Izumo et
 192 al., 2010; Zhou et al., 2015; Jourdain et al., 2016; Wieners et al., 2017a,b; Wang et al., 2019;
 193 Cai et al., 2019). In fact, the Indian Ocean (either IOD and Indian Ocean Basin-wide
 194 Warming) has a highly significant impact on both the variability and predictability of ENSO,
 195 as evidenced by de-coupled experiments using different coupled climate models (Luo et al.,
 196 2010; Izumo et al., 2010; Santoso et al., 2012; Terray et al., 2016; Kajtar et al., 2016; Wang
 197 et al., 2019). For example, IOD events co-occurring with ENSO may fasten its phase
 198 transition (Kug and Kang, 2006; Luo et al., 2010; Izumo et al., 2010; Kug and Ham, 2012).
 199



200
 201 **Fig. 8-3:** (a) Positive IOD composite of specific humidity (mm, shaded) and moisture flux (integrated
 202 up to 300 hPa, kg/m/s, vectors) anomalies averaged in summer (JJAS). The figure is taken from Behera

203 and Ratnam (2018); (b) schematic of the IOD influence on ISM rainfall embedded in the TBO (adapted
204 from the scheme in Webster and Hoyos (2010)).

205 In the absence of ENSO, IOD still exists (usually known as “pure IOD”) and its
206 variability is mainly driven by the eastern Indian Ocean in a suite of coupled climate model
207 simulations nudging the tropical Pacific SSTs toward an SST climatology estimated from
208 observations or a control simulation (Cretat et al., 2017, 2018), as consistently seen in other
209 coupled model studies (Gualdi et al., 2003; Fischer et al., 2005; Behera et al., 2005, 2006;
210 Luo et al., 2010; Izumo et al., 2010; Wang et al., 2016, 2019). In the nudged experiments, the
211 strong diabatic heating associated with enhanced rainfall over the eastern IOD lobe
212 modulates the local Hadley circulation and induces a negative (positive) rainfall anomaly in
213 the northern Indian Ocean during boreal summer during negative (positive) IOD events, as
214 suggested from observational studies (Fig. 8-4; Cretat et al., 2017). Such changes in the local
215 Hadley circulation are attenuated in the presence of ENSO because ENSO-induced changes
216 in the (zonal) Walker circulation dominate (Cretat et al., 2017). It has also been found that
217 rainfall anomalies over India associated with these pure IODs are modest and not statistically
218 significant, especially at the beginning of the monsoon, although the simulated SST
219 variability in the eastern Indian Ocean is overestimated (Fischer et al., 2005; Terray et al.,
220 2012; Cretat et al., 2017, 2018). However, pure IODs promote a quadrupole rainfall pattern
221 linking the tropical Indian Ocean and the Western North Pacific, and induce important zonal
222 shifts of the Walker circulation in the absence of ENSO (Fig. 8-4), in agreement with earlier
223 findings (Li et al., 2006; Chowdary et al., 2011). The circulation patterns with and without
224 ENSO largely differ, confirming potential opposite effects between IOD and ENSO (e.g.,
225 Ashok et al., 2001; Lau and Nath, 2012; Pepler et al., 2014).

226 Pure IOD events may also help sustaining the TBO: a stronger-than-observed biennial

227 spectrum of the IOD is found after removing ENSO's impacts (Behera et al., 2006; Cretat et
228 al., 2018). Moreover, coupled ocean-atmosphere interactions in the Indian Ocean can sustain
229 its own TBO without ENSO (Cretat et al., 2018). First, subsurface ocean dynamics play a key
230 role in the biennial anomalies during boreal winter (Rao et al., 2002, 2009; Schott et al.,
231 2009; McPhaden and Nagura, 2014; Delman et al., 2016) with a sudden reversal of
232 thermocline anomalies in the eastern equatorial Indian Ocean forced by intra-seasonal
233 disturbances reminiscent of the Madden-Julian Oscillation (MJO; Rao and Yamagata, 2004;
234 Han et al., 2006). Second, tropical-extra-tropical interactions within the Indian Ocean appear
235 to be the main trigger of IODs in the absence of ENSO (Cretat et al., 2018). In nudged
236 experiments, both the power spectra of the ISMR and IOD indices during boreal summer
237 shift toward increased biennial variability compared to the control simulation, which may be
238 more consistent with a possible coupling of the IOD with ISM in the absence of ENSO, but
239 in a regional TBO framework (Cretat et al., 2018). However, this TBO framework is again
240 mainly based on the strong influence of the ISM circulation on the Indian Ocean SSTs
241 despite the absence of ENSO in the nudged experiments.

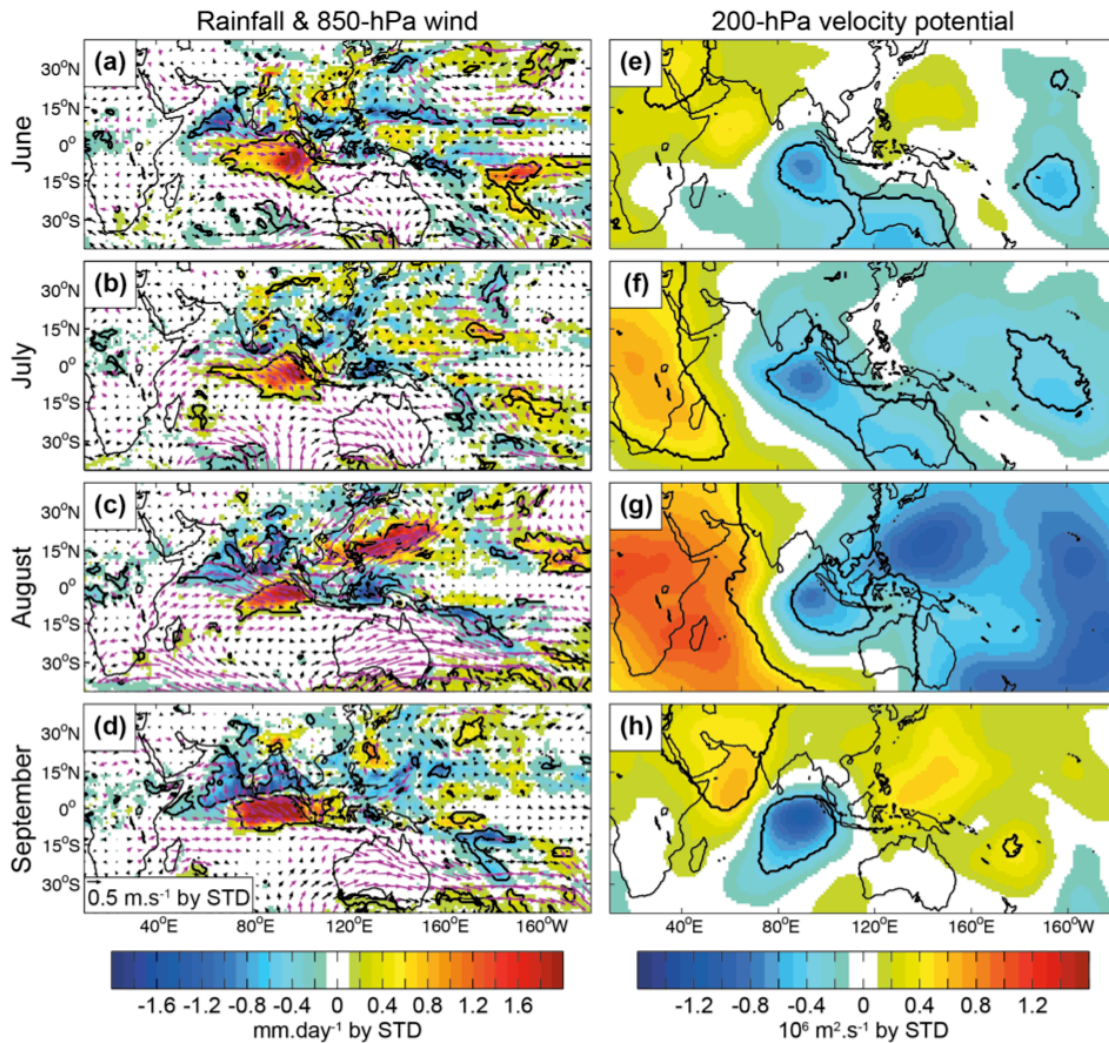
242 In synthesis, these recent modeling studies do suggest that IOD exists without ENSO,
243 but the exact relationships between IOD and ISM remain elusive, even in the absence of
244 ENSO.

245 **8.4 Past, present and future IOD influence on the ENSO-monsoon teleconnection**

246 At long time scales, the influence of IOD on ISM seems opposite to the effect of ENSO, and
247 the IOD-ISM rainfall relationship seems to vary complementarily to that between ENSO and
248 ISM (Ashok et al., 2001; Krishnaswamy et al., 2015). In fact, the IOD-ISM relationship has
249 strengthened in the recent decades (Ashok et al., 2001, 2004; Ashok and Saji, 2007; Izumo et

250 al., 2010; Ummenhofer et al., 2011) due to non-uniform warming of the Indian Ocean (Ihara
251 et al., 2007; Cai et al., 2009), while the ENSO–ISM relationship has weakened (Kumar et al.,
252 1999; Ashrit et al., 2001; Ihara et al., 2007).

253 On longer IOD records, changes in frequency and teleconnections have been
254 identified (e.g. Abram et al., 2008; Kayanne et al., 2006; Abram et al., 2020). Coral proxy
255 records from Lake Victoria in Kenya suggest that the influence of ENSO has decreased over
256 the western Indian Ocean in recent decades (Nakamura et al., 2009). A mode shifts in IOD
257 variability related to the warming trend in the western Indian Ocean has raised the mean SST
258 to a threshold value that encourages tropical convections (Nakamura et al., 2009). A recent
259 reconstruction of the last millennium indicates clustering of positive IOD events with
260 extreme IOD variability and a persistent tropical Indo-Pacific climate coupling (Abram et al.,
261 2020). The frequency and strength of IOD events exceptionally increased during the
262 twentieth century associated with enhanced upwelling in the eastern pole of the IOD, likely
263 making more direct the influence of the IOD on the Asian monsoon (Abram et al., 2008).
264 These processes and associated changes in the Walker circulation, linked to global warming,
265 may precondition the mean state to trigger frequent positive IOD events, together with
266 intense short rains in East Africa.



267

268 **Fig. 8-4:** June to September (a-d) rainfall (shading), 850-hPa wind (vectors), and (e-h) 200-hPa
 269 velocity potential anomalies regressed onto normalized boreal fall (i.e., SON) SST anomalies over the
 270 eastern IOD pole (90°E–110°E, 10°S–0°) when ENSO is removed. Significant anomalies at the 90% level
 271 are shown with black contours for rainfall and 200-hPa velocity potential anomalies, and with purple
 272 vectors for 850-hPa wind anomalies. Positive 200-hPa velocity potential anomalies correspond to
 273 abnormal upper-level mass flux convergence. Adapted from Cretat et al. (2017).

274

275 In the 20th century, the ENSO-IOD correlation was strongly positive and significant
 276 since mid-60s (Cherchi and Navarra, 2013), with ENSO and IOD almost independent before
 277 1970 (Yuan and Li, 2008). A recent weakening of the coupling has been identified during

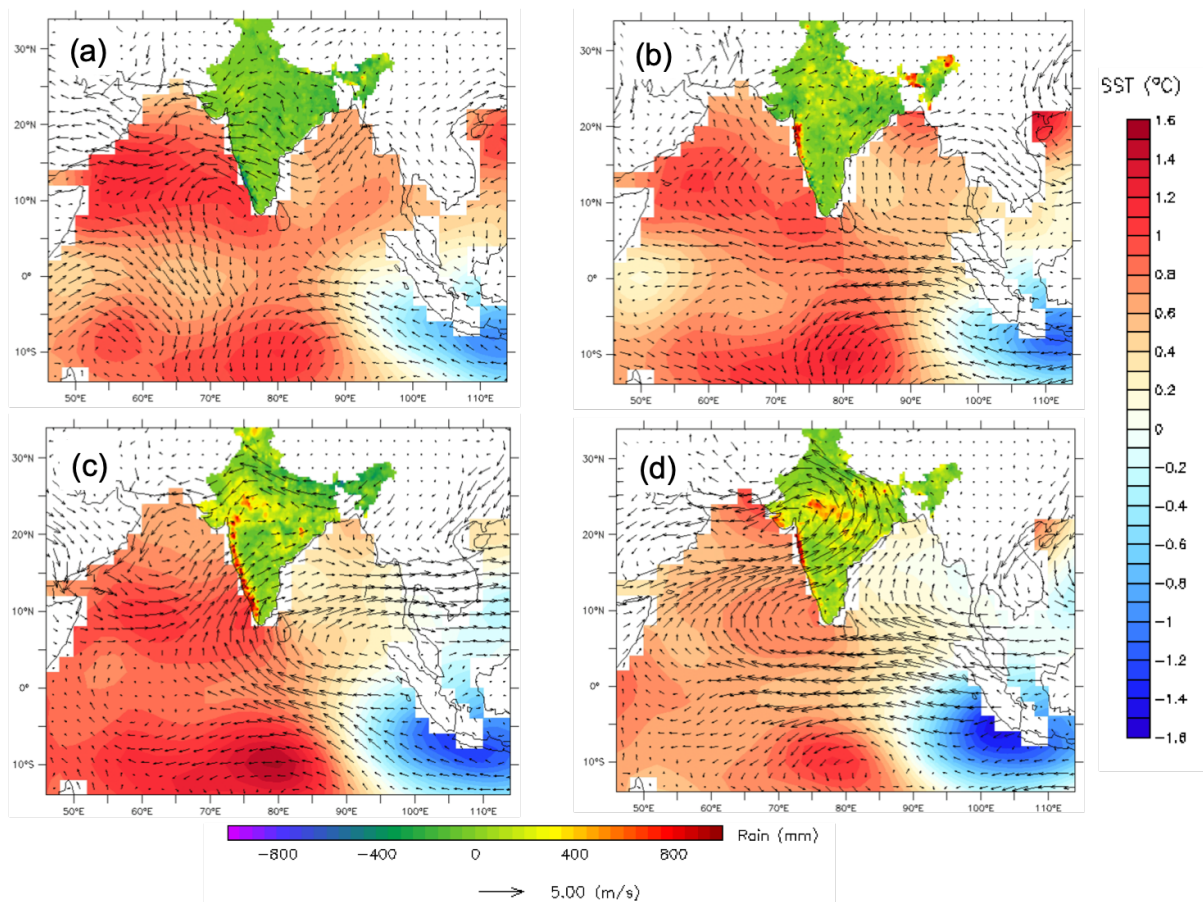
278 1999-2014 compared to the previous two decades (i.e., 1979-1998), associated with different
279 spatial patterns in ENSO evolution during boreal spring and summer (Ham et al., 2017). The
280 stronger/weaker correlation may correspond with either strong or weak ENSO-monsoon
281 relationship and with strong or weak IOD-monsoon relationship, with differences arising
282 from the relationship between Indian monsoon rainfall and SST in other ocean basins rather
283 than the Indo-Pacific sector alone (Cherchi and Navarra, 2013).

284 The IOD-ENSO-ISMR relationship appears to work differently with different seasons
285 (Agrawal et al 2017). The connection between ISM and IOD is mostly confined in the
286 summer and autumn, while that with ENSO is stronger and extends more in time (Cherchi
287 and Navarra, 2013). In fact, the evolution of the correlation between ISMR and monthly
288 NINO3.4 is maximum in August-November, remaining strong and stable until March of the
289 following year (Gershunov et al., 2001).

290 The 1997 El Niño was one of the strongest events occurred in the 20th century, but
291 ISMR was slightly above normal (Srinivasan and Nanjundiah, 2002), and this has been
292 attributed to the influence from the Indian Ocean (Slingo and Annamalai, 2000; Sreejith et
293 al., 2015). Similarly, the influence of IOD helped nullify the effect of ENSO on the monsoon
294 during 1997 (Saji et al., 1999; Webster et al., 1999). Positive IOD events, as those occurred
295 during 2007-2008, co-existed with La Niña episodes (e.g. Ashok et al., 2003; Cai et al.,
296 2009). A very strong positive IOD event occurred in the summer of 1994, as a clear coupled
297 ocean-atmosphere phenomenon of the Indian Ocean (Behera et al., 1999). The 1994 event
298 lasted more than 8 months from March to October and positively influenced the ISMR that
299 recorded 265 mm/month, a value 19% above the climatological mean (Guan and Yamagata,
300 2003).

301 In 2019 the monsoon onset was delayed by about 7 days over India with the June

302 rainfall recording a deficit of about 33% with respect to the climatological mean (Gadgil et
303 al., 2019; <http://www.imd.gov.in>). According to the India Meteorology Department (IMD),
304 subsequent to this monsoon onset, the further northward progression of the monsoon
305 remained slow due to the formation of a very severe cyclone over central eastern Arabian Sea
306 (i.e., the cyclone VAYU that formed during 10-17th June 2019,
307 https://mausam.imd.gov.in/imd_latest/contents/season_report.php). El Niño weakened in July
308 while a strong positive IOD started to develop (Fig. 8-5b) and rainfall picked up strength
309 during the latter stages of the monsoon season (late July to September; Fig. 8-5b-d),
310 remaining above normal. For that year, the seasonal rainfall recorded for India has been
311 quantified at 110% of the long period average as defined by IMD, with the September
312 rainfall being 152%. The positive IOD that occurred during late summer in 2019 was one of
313 the strongest in the recent times (Fig. 8-2), with its predictability linked with the existence of
314 the El Niño Modoki in the Equatorial Pacific (Doi et al., 2020) and to a strong pressure
315 dipole between the Australian High and the South China Sea/Philippine Sea region (Lu and
316 Ren, 2020). The exceptional intensity of the event remains even after the IOD index is
317 detrended (not shown). This very recent case illustrates how the interactions between ISM,
318 IOD, and ENSO are subtle and complex, and highly influenced by internal dynamical
319 processes.



320

321 **Fig. 8-5:** 2019 case from (a) June to (d) September for SST ($^{\circ}\text{C}$), 850 hPa winds (m/s), and
 322 precipitation (mm) anomalies with respect to 1980-2010 mean climatology. SST data is taken from
 323 ERSSTv5 (Huang et al. 2017), 850 hPa wind vectors are obtained from ECMWF ERA5 reanalysis
 324 (Hersbach and Dee, 2016) and precipitation from daily gridded rainfall dataset over India (Pai et al. 2014).
 325

326 In the Coupled Model Intercomparison Project Phase 5 (CMIP5) models a correct
 327 representation of the coupled processes (i.e., Bjerknes feedback) in the equatorial Indian
 328 Ocean is a necessary condition for realistic monsoon simulations (Annamalai et al., 2017). At
 329 the same time, a too weak rainfall over the Arabian Sea in model may generate a warm SST
 330 bias over the western equatorial Indian Ocean that in the following fall may amplify the error
 331 toward an IOD-like SST bias via the Bjerknes feedback (Li et al., 2015). The unrealistic

332 present-day IOD-ISMR correlation simulated by the majority of CMIP5 models may also be
333 related to an overly strong control by ENSO (Li et al., 2017), likely leading to an
334 underestimation of the projected future ISMR increase. Still, CMIP5 models project an
335 increase in ISMR in a warmer climate with a reasonably strong consensus among models
336 (Jayashankar et al., 2015). CMIP5 models' projections tend to exhibit a positive IOD-like
337 pattern in the tropical Indian Ocean with weaker (stronger) warming in the east (west) and an
338 easterly wind trend (Zheng et al., 2010; 2013). The response is driven by the projected
339 weakening of the Walker circulation in a warmer climate in the majority of models (Vecchi
340 et al., 2006; Kociuba and Power, 2015).

341 In future projections, surface moisture increase dominates the changes in rainfall
342 associated with the IOD, while IOD related SST changes dominate the corresponding
343 changes in the circulation, decreasing at a rate of $13.7\%/^{\circ}\text{C}$ (Huang et al., 2019). The
344 ensemble spread in the IOD amplitude change is large (Ng et al., 2018), and it is related to
345 that of the ENSO amplitude change (Hui and Zheng, 2018). The large spread in the IOD
346 response to increasing Greenhouse Gases (GHGs) with significant variations in the amplitude
347 and skewness of the dipole and in climatological zonal SST gradient is due to small
348 differences in the mean thermocline depth induced by internal climate variability via the
349 positive Bjerknes feedback (Ng et al 2018).

350 The frequency of extreme IOD events is projected to increase under global warming
351 conditions (Cai et al., 2014, also see Chapter-21), with a persistence of the ENSO-IOD
352 linkage in a warmer future world (Stuecker et al., 2017). The characteristics of the ENSO-
353 IOD are likely to continue in the future, and given that ENSO and its predictability are
354 modulated on decadal timescales (i.e., Wittenberg, 2009; Wittenberg et al., 2014;

355 Karamperidou et al., 2014), the same should be expected for the IOD (Stuecker et al., 2017).

356 CMIP6 models largely improved in the simulation of the spatial and temporal pattern
357 of the ISM (Gusain et al., 2020), especially over the Western Ghats and the foothills over the
358 Himalayas, whereas a majority of the CMIP5 models underestimated rainfall over central and
359 northern India (Jain et al., 2019). A subset of CMIP6 models (Table 1) confirms CMIP5
360 results, with a tendency toward larger IOD amplitude at the end of the 20th century and in the
361 future, at least under the most extreme CMIP6 scenario SSP5-8.5 (Fig. 8-6a). SSP stands for
362 Shared Socioeconomic Pathways with 5 representing an economic vision of the future with
363 relatively optimistic trends for human development but assuming an energy-intensive, fossil-
364 fuel economy, and 8.5 corresponding to the forcing (in W/m^2) by 2100 (O'Neill et al., 2016).
365 In summer (JJAS mean), SST regressed onto the IOD index project larger IOD-related
366 anomalies over the Pacific and Indian Oceans (Fig. 8-6b,c). In the projection, precipitation
367 and winds regressed onto the IOD index show modest positive anomalies over India and
368 weaker easterlies along the Equator because of a weaker negative pole (Fig. 8-6d,e). This
369 figure is just a flavor of the complex relationship as projected in the new generation of
370 coupled climate models. For example, the methodology applied does not fully disentangle
371 how the mean state changes and its role. A more systematic analysis and comparison of
372 CMIP5 and CMIP6 experiments would be needed to fully understand differences and
373 potential improvements. This is outside the scope of this chapter but it is under investigation
374 in separated ongoing researches.

375

376 **Table 8.1:** List and some characteristics of CMIP6 models used. More details about the
377 models can be found at <https://pcmdi.llnl.gov/CMIP6/>

Model Name	Institute/Country	Atmosphere	Ocean Resolution	Earth System
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		Resolution (km)	(km)	Model
ACCESS-CM2	CSIRO-ARCCSS/Australia	250	100	no
ACCESS-ESM1-5	CSIRO-ARCCSS/Australia	250	100	yes
BCC-CSM2-MR	BCC/China	100	50	no
CESM2	NCAR/US	100	100	yes
CESM2-WACCM	NCAR/US	100	100	yes
CNRM-CM6-1	CNRM-CERFACS/France	250	100	no
CNRM-CM6-1-HR	CNRM-CERFACS/France	100	25	no
CNRM-ESM2-1	CNRM-CERFACS/France	250	100	yes
CanESM5	CCCma/Canada	500	100	yes
EC-Earth3	EC-Earth-Consortium/Europe	100	100	no
EC-Earth3-Veg	EC-Earth-Consortium/Europe	100	100	no
FGOALS-f3-L	CAS/China	100	100	no
FGOALS-g3	CAS/China	250	100	no
GFDL-ESM4	NOAA-GFDL/US	100	50	yes
HadGEM3-GC31-LL	MOHC-NERC/UK	250	100	no
INM-CM4-8	INM/Russia	100	100	no
INM-CM5-0	INM/Russia	100	50	no
IPSL-CM6A-LR	IPSL/France	250	100	yes
KACE-1-0-G	NIMS-KMA/South Korea	250	100	no
MCM-UA-1-0	UA/US	250	250	no
MIROC-ES2L	MIROC/Japan	500	100	yes
MIROC6	MIROC/Japan	250	100	no
MPI-ESM1-2-HR	MPI-M DWD DKRZ/Germany	100	50	yes
MPI-ESM1-2-LR	MPI-M AWI/Germany	250	250	yes
MRI-ESM2-0	MRI/Japan	100	100	yes
NESM3	NUIST/China	250	100	no
NorESM2-LM	NCC/Norway	250	100	yes
NorESM2-MM	NCC/Norway	100	100	yes
UKESM1-0-LL	MOHC NERC NIMS- KMA NIWA/UK	250	100	yes

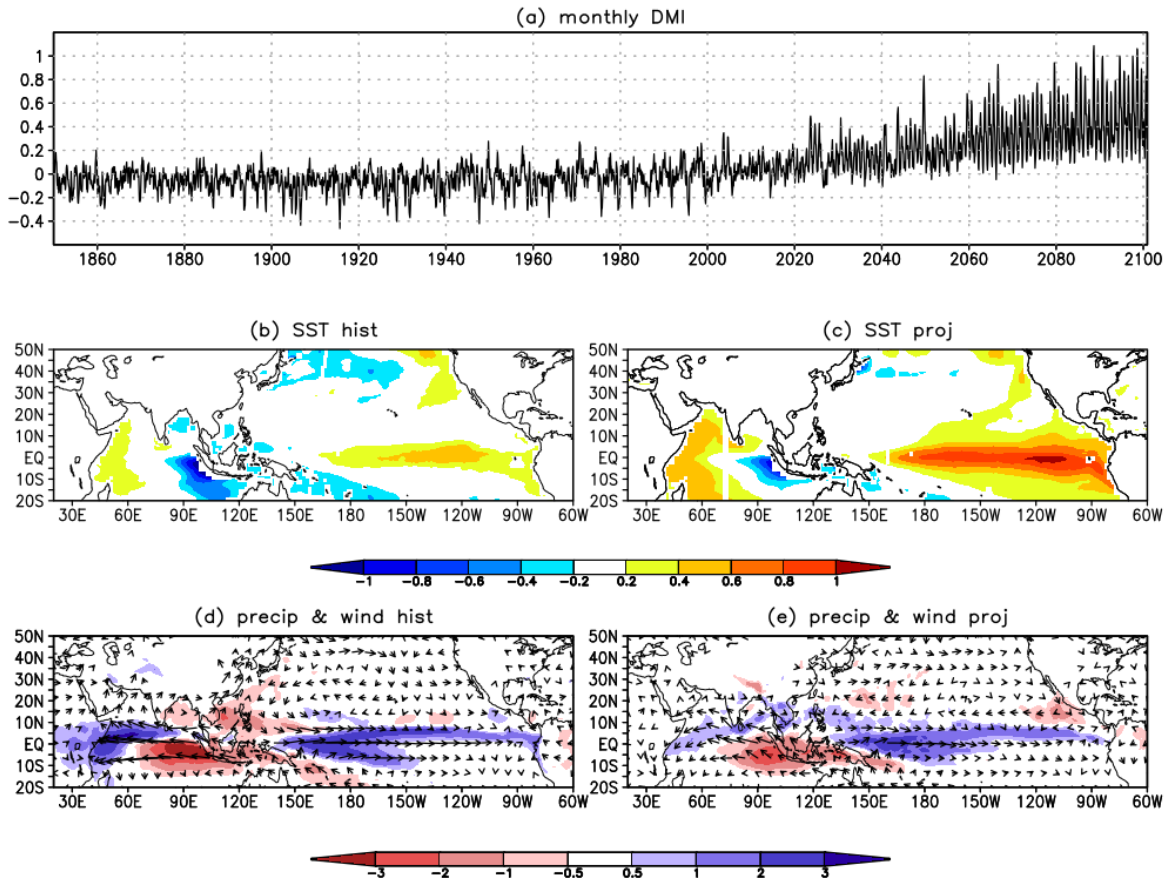
378

379

380 **8.5 Challenges and future perspectives**

381 The main challenges in having a complete picture of the influence of IOD on the
382 ENSO-ISM teleconnection remains related to a full understanding of the IOD itself and to a
383 full and agreed understanding of how the IOD is related to ENSO on one side, and its pure
384 (i.e., independent from ENSO) relationship with ISM on the other. While some progress have
385 been made in recent decades in simulating ENSO variability (Bellenger et al., 2014), many
386 important issues remain open due to the unavailability of long term observational record and

387 the biases in state-of-the-art climate models affecting ISM and the Indian Ocean
388 simulations(Li et al., 2015; Annamalai et al., 2017).



389
390 **Fig. 8-6:** (a) Monthly DMI index (anomalies, °C) for 20th and 21st centuries from a subset of CMIP6
391 models (Table 8.1). The index has been computed as in Fig. 8-2 (same areas difference and anomalies with
392 respect to the 1980-2010 mean climatology). The index has been computed for each model and then
393 averaged to obtain the ensemble mean. (b,c) SST (°C, shaded) and (d,e) precipitation (mm/day, shaded)
394 and 850 hPa wind (m/s, vectors) regressed on the DMI index (values for 1°C change in the index) for JJAS
395 mean during the historical period and the future projection, respectively. One member for each model has
396 been considered. For the 21st century, the SSP5-8.5 scenario has been used. In panels c and e, the time-
397 series have been detrended before computing the regression, to keep out the trend from the related
398 variability.

399

400

401 An exhaustive analysis of the IOD recorded past events would allow a categorization
402 of the main characteristics of the processes at play, but the observed record remains short to
403 have a statistically robust assessment. On the other hand, in state-of-the-art coupled climate
404 models the simulation of the IOD, ISM, and related characteristics (including mean state and
405 variability) still has large biases thus precluding a complete understanding of the processes at
406 work. For example, it is not clear whether the weak IOD-monsoon relationship simulated in
407 the models (Fig. 8-6) is realistic or not, due to the exaggerated IOD variability or to the
408 overly strong control of ENSO simulated by current global climate models. Moreover, it has
409 to be clarified whether poor simulations of other factors, like IOD-induced cross-equatorial
410 flows, may be important. Similarly, a complete understanding of the coupling and feedback
411 processes between the developing phase of the IOD and the ISMR, including the possible
412 feedback on the development of the IOD in the subsequent season, is still missing. As a
413 consequence, much more in-depth observational and modeling studies (including model's
414 improvements) are clearly needed to understand the IOD effect on ISM and the relative roles
415 of remote versus local forcing on the ISM-ENSO relationship as well as the role of internal
416 atmospheric processes in modulating that relationship. In future climate projections, it would
417 be useful to understand how changes in the simulated IOD properties contribute to the
418 relative importance of thermodynamic and dynamic monsoon processes at play in global
419 warming frameworks.

420 For all the points above, crucial keys are the collection of as much as possible
421 observations during known IOD events, and how IOD and related properties are simulated in
422 state-of-the-art climate models. The need for more observations in the Indian Ocean is
423 particularly important because of its changes within the last warm decades (Hermes et al.,

424 2019). For the simulation of IOD and related models' performance, more efforts should be
425 dedicated to reducing systematic biases in coupled climate models and/or in performing ad-
426 hoc sensitivity experiments in a large set of different coupled models to clarify the dynamics
427 involved in IOD formation and related teleconnections. One possibility is to design
428 coordinated international efforts with specific common experiments, likely following current
429 CMIP frameworks (Zhou et al., 2016) or the CORE-II experiments (Rahaman et al., 2020).

430 **8.6 Conclusions**

431 The IOD is one of the dominant modes of SST variability of the tropical Indian Ocean. It is
432 recognized as having important teleconnections worldwide but here it has been considered in
433 terms of its influence on the ISM and its relationship with ENSO. In particular, the literature
434 focused on its active or passive role has been reviewed, evidencing how the influence of the
435 IOD on the ISM rainfall can be interpreted as having a direct impact through moisture
436 transport over the western Indian Ocean or modifications of the local Hadley cell, or
437 alternatively in the framework of the tropospheric biennial oscillation. Recently, more
438 literature is available on the role of the IOD independently from ENSO, or even as a trigger
439 of ENSO itself. Still, combining modelling and observational studies, the precise relationship
440 between IOD and ISM remains elusive, with or without ENSO.

441 Considering major ENSO and/or IOD events, it has been recorded how in 1997 the
442 failure of the negative relationship between ENSO and the ISM rainfall was associated with a
443 positive IOD event developing that summer, or how the strong IOD of 1994 has been
444 responsible for a stronger than normal monsoon that summer. Recently, the strongest IOD
445 event recorded in 2019 and its evolution within the summer evidenced how the interactions
446 between ISM, IOD, and ENSO are complex, and highly influenced by internal dynamical
447 processes.

448 CMIP5 and CMIP6 models agree in projecting stronger IOD events in the future (also
449 see Chapter 21), but how this project on atmospheric anomalies over the Indo-Pacific region
450 may not be fully consistent. A systematic analysis of the two model intercomparisons sets is
451 needed to fully understand the possible differences.

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465

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955

956 **Figure captions:**

957 **Fig. 8-1:** Schematic of the Indian Ocean Dipole in its positive phase with warm SST anomalies on the
958 west and cold SST anomalies toward the coast of Sumatra. The green arrows indicate the direction of the
959 prevailing corresponding surface winds.

960 **Fig. 8-2:** Normalized monthly IOD index (std) defined as anomalous SST gradient between the western
961 equatorial Indian Ocean (50°E-70°E and 10°S-10°N) and the southeastern equatorial Indian Ocean (90°E-
962 110°E and 10°S-0°N). Anomalies have been downloaded from
963 https://psl.noaa.gov/gcos_wgsp/Timeseries/DMI/, but 2019 values have been integrated from JAMSTEC
964 repository (<http://www.jamstec.go.jp/virtualearth/general/en/index.html>). Red and blue markers along 6
965 and -6 std correspond to NINO3.4 anomalies (1981-2010 mean removed) larger than 0.5°C. NINO3.4
966 anomalies have been downloaded from https://psl.noaa.gov/gcos_wgsp/Timeseries/Nino34/. As indicated
967 in the respective websites, both IOD and NINO3.4 values are computed from the HadISST1 dataset
968 (Rayner et al., 2003).

969 **Fig. 8-3:** (a) Positive IOD composite of specific humidity (mm, shaded) and moisture flux (integrated
970 up to 300 hPa, kg/m/s, vectors) anomalies averaged in summer (JJAS). The figure is taken from Behera
971 and Ratnam (2018); (b) schematic of the IOD influence on ISM rainfall embedded in the TBO (adapted

972 from the scheme in Webster and Hoyos (2010).

973 **Fig. 8-4:** June to September (a-d) rainfall (shading), 850-hPa wind (vectors), and (e-h) 200-hPa
974 velocity potential anomalies regressed onto normalized boreal fall (i.e., SON) SST anomalies over the
975 eastern IOD pole (domain: 90°E–110°E, 10°S–0°) when ENSO is removed. Significant anomalies at the
976 90% level are shown with black contours for rainfall and 200-hPa velocity potential anomalies, and with
977 purple vectors for 850-hPa wind anomalies. Positive 200-hPa velocity potential anomalies correspond to
978 abnormal upper-level mass flux convergence. Adapted from Cretat et al. (2017).

979 **Fig. 8-5:** 2019 case from (a) June to (d) September for SST (°C), 850 hPa winds (m/s) and precipitation
980 (mm) anomalies with respect to 1980-2010 mean climatology. SST data is taken from ERSSTv5 (Huang et
981 al. 2017), 850 hPa wind vectors are obtained from ECMWF ERA5 reanalysis (Hersbach and Dee, 2016)
982 and precipitation from daily gridded rainfall dataset over India (Pai et al. 2014).

983 **Fig. 8-6:** (a) Monthly DMI index (anomalies, °C) for 20th and 21st centuries from a subset of CMIP6
984 models (Table 1). The index has been computed as in Fig. 2 (same areas difference and anomalies with
985 respect to the 1980-2010 mean climatology). The index has been computed for each model and then
986 averaged to obtain the ensemble mean. (b,c) SST (°C, shaded) and (d,e) precipitation (mm/day, shaded)
987 and 850 hPa wind (m/s, vectors) regressed on the DMI index (values for 1°C change in the index) for JJAS
988 mean during the historical period and the future projection, respectively. One member for each model has
989 been considered. For the 21st century, the SSP5-8.5 scenario has been used. In panels c and e, the
990 timeseries have been detrended before computing the regression, to keep out the trend from the related
991 variability.