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1 **Chapter 15**

2 **South Asian Summer Monsoon and subtropical deserts**

3
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Abstract

The relationships between south Asian monsoon and Northern Hemisphere (NH) subtropical deserts have generated a lot of research in the recent times as both systems, despite of contrasting climates, are expected to be severely affected by anthropogenic climate change. This review envisages two pathways for the monsoon-desert relationship. The first one hypothesizes a significant influence of the monsoon on subtropical deserts whereby convection over the Indian Summer Monsoon (ISM) region induces Rossby-wave descent to its west. The second one proposes a very different perspective by emphasizing the potential role of NH deserts on ISM either through the changes of surface heating over the subtropical deserts or by dry air intrusions from arid regions into the monsoon domain. However, most current coupled climate models struggle to simulate realistically these monsoon-desert relationships due to various reasons, as documented in this chapter, calling for advances in coupled models with improved physical parameterizations.

Keywords: Indian summer monsoon, NH subtropical deserts, monsoon-desert relationship, anthropogenic climate change

42 **15.1 Introduction**

43 Monsoon and desert regions coexist at the subtropical latitudes of the African-Asian
44 continent (e.g., Rodwell and Hoskins, 1996; Warner, 2004). The mutual association between
45 these two contrasting climates has been studied in the past as well as in the recent times (e.g.,
46 Ramage, 1966; Charney, 1975; Charney et al., 1977; Shukla and Mintz, 1982; Sud and Fennessy,
47 1982; Smith, 1986a, 1986b; Mooley and Paolino, 1988; Sud et al., 1988; Parthasarathy et al.,
48 1992; Yang et al., 1992; Webster, 1994; Rodwell and Hoskins, 1996, 2001; Sikka, 1997;
49 Claussen, 1997; Bonfils et al., 2000; Douville et al., 2001; Xue et al., 2004; Yasunari et al., 2006;
50 Wang, 2006; Wu et al., 2009; Biasutti et al., 2009; Lavaysse et al., 2009; Xue et al., 2010;
51 Bollasina and Nigam, 2011a, 2011b; Bollasina and Ming, 2013; Tyrlis et al., 2013; Cherchi et al.,
52 2014; Vinoj et al., 2014; Shekhar and Boos, 2017; Sooraj et al., 2019). As desertification is a
53 fundamental process of the ongoing climate change (Cook and Vizzy, 2015; Zhou, 2016; Wei et
54 al., 2017) and climate projections of the south Asian monsoon remain uncertain (e.g., IPCC,
55 2013; Sabeerali et al., 2015; Annamalai et al., 2015; Sooraj et al., 2015; Krishnan et al., 2016;
56 Kitoh, 2017; Singh and Achutarao, 2018; Wang et al., 2020), a renewed interest to understand
57 the monsoon-desert relationships has grown. Furthermore, these two climate systems are affected
58 by severe temperature, rainfall and radiation biases in current climate models (e.g., Bollasina and
59 Ming, 2013; Levine et al., 2013; Sperber et al., 2013; Roehrig et al., 2013; Prodhomme et al.,
60 2014; Sandeep and Ajayamohan, 2015; Samson et al., 2016; Haywood et al., 2016; Terray et al.,
61 2018), and improving their representation is an important aspect to have better climate forecasts
62 and projections (e.g., Cherchi et al., 2014; Terray et al., 2018; Sooraj et al., 2019).

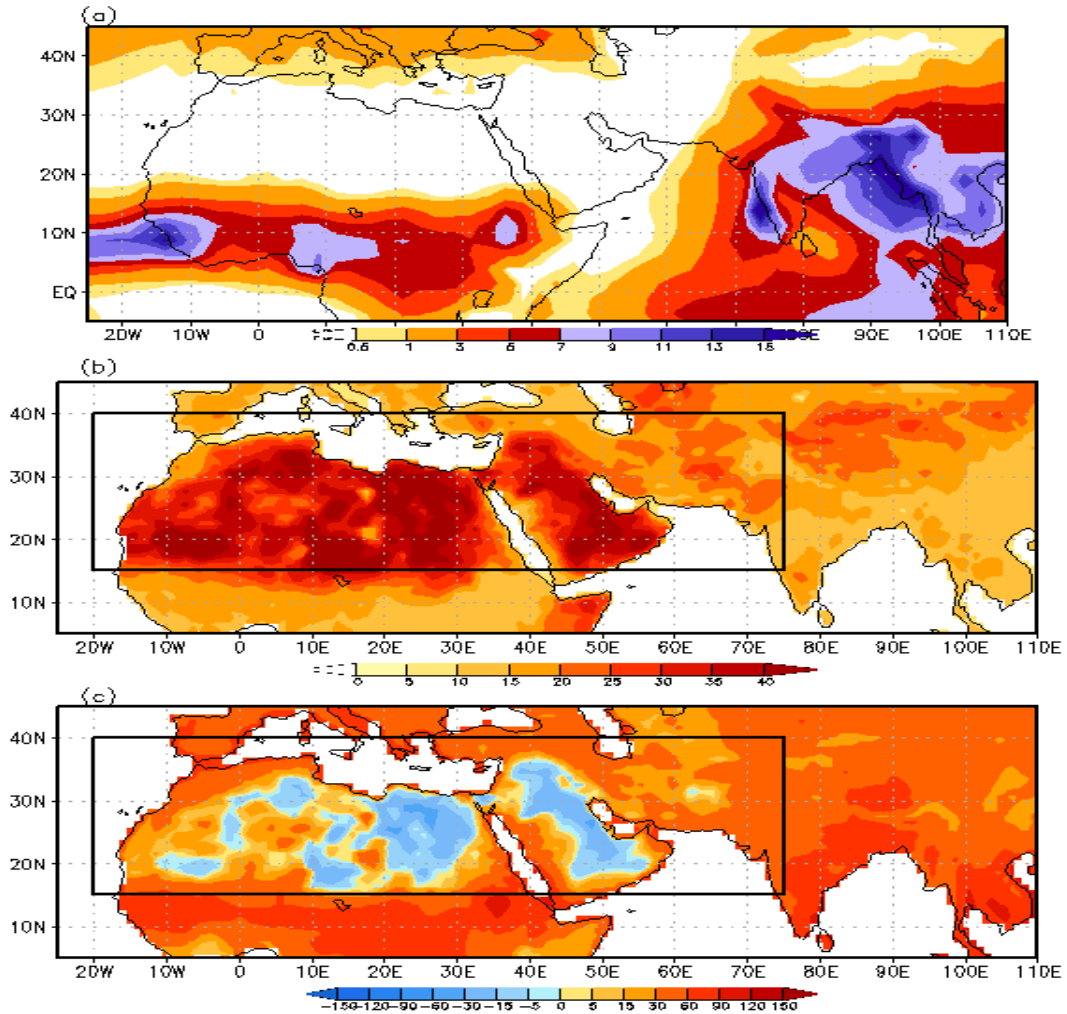
63 Against the backdrop of this, we made an attempt to provide a comprehensive overview
64 of the mutual relationships between these two contrasting climates thereby highlighting the

65 underlying mechanisms. The chapter is organized as follows: Section 15.2 describes the salient
66 climatological characteristics of the monsoon-desert system and highlights the historical
67 background on the existence of subtropical deserts over African-Asian regions at the same
68 latitude than the south Asian monsoon. Section 15.3 focuses on the south Asian monsoon (i.e.,
69 Indian Summer Monsoon, ISM) influence on the hot subtropical deserts, thus presenting the
70 monsoon-desert paradigm using observation, reanalysis and coupled General Circulation Models
71 (GCMs). Section 15.4 reviews the literature on the potential role of deserts in modulating the
72 ISM. Finally, Section 15.5 encapsulates the main highlights as drawn from this Chapter review.

73 **15.2 The Monsoon-Desert system: Background settings**

74 The monsoon-desert system over African-Asian region is characterised by a sharp rainfall
75 gradient during the boreal summer (e.g., from June to September) with heavy rainfall in the
76 monsoon regions of the northern hemisphere (NH), but only sporadic and low rainfall in the
77 adjacent arid regions (Fig. 15.1a). The contrasts between the two climates are further
78 corroborated when considering other important parameters such as surface albedo or net
79 radiation budget at the Top of the Atmosphere (TOA; Fig. 15.1b-c).

80



81

82 **Figure 15.1:** Climatological map of (a) rainfall (mm day^{-1}), (b) land surface albedo (%)
 83 and (c) net radiation budget at TOA (Wm^{-2}), for boreal summer period (June to September). In
 84 (a) rainfall climatology is computed for the 1986-2014 period from Global Precipitation
 85 Climatology Project (GPCP version 2.1; Huffman et al., 2009). In (b) and (c), albedo and
 86 radiation climatology is computed for the 2000-2018 period from the Clouds and the Earth's
 87 Radiant Energy System Energy Balanced and Filled (CERES-EBAF edition 4.0; Kato et al.,
 88 2018). In (b) and (c), the region highlighted in the black rectangle refers to “subtropical deserts”.

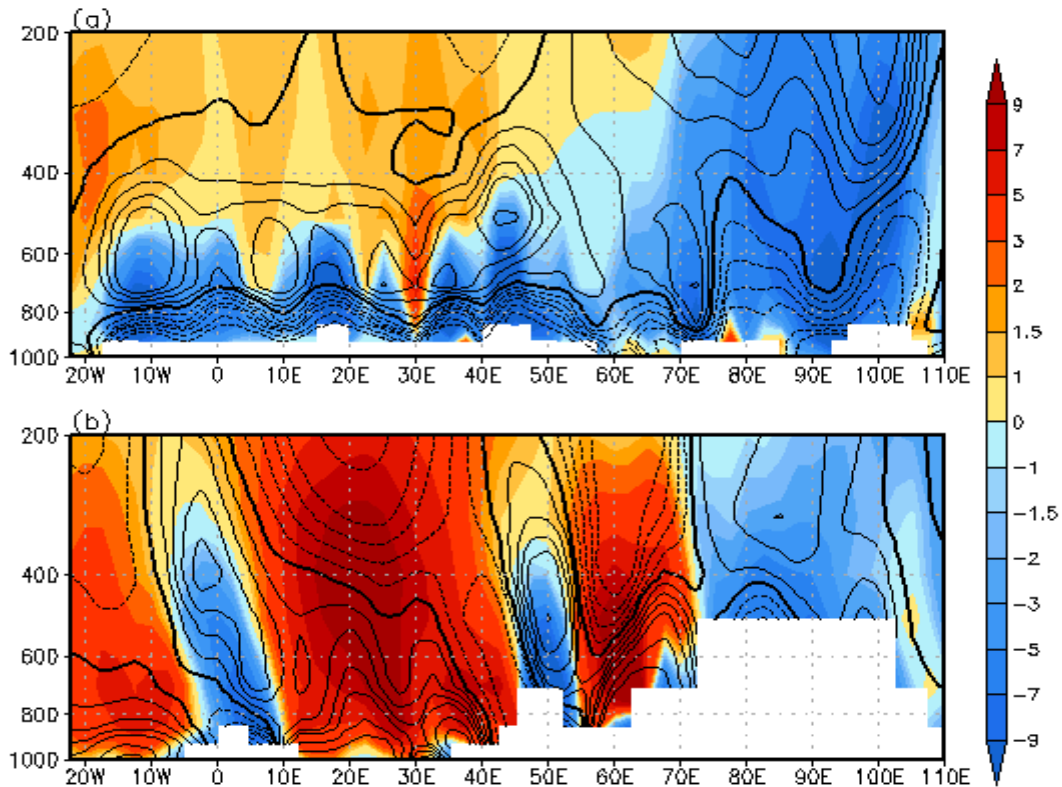
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90 The subtropical desert regions, with bright sandy surface terrains, clear sky conditions,
91 high temperature, reduced soil moisture and lack of vegetation, are characterised by relatively
92 high surface albedo (Fig. 15.1b; Sikka, 1997; Warner, 2004; Terray et al., 2018; Sooraj et al.,
93 2019). Figure 15.1b shows two important examples for subtropical deserts: The Arabian-Iran-
94 Thar desert located just to the west of the ISM system (Sikka, 1997) and the Sahara just to the
95 north of the West African Monsoon (WAM) (Lavaysse et al., 2009). In this Chapter, these high
96 albedo regions lying across north Africa and west Asia (i.e., the geographical region bounded by
97 20°W – 75°E , 15° – 40°N ; Fig. 15.1b) are hereafter collectively referred to as “subtropical deserts”
98 (or simply “arid regions”).

99 As per the traditional theory dating back to the 1970s, the subtropical desert climates are
100 associated with the descending branch of the Hadley circulation over the subtropics of the NH
101 (Warner, 2004). However, NH Hadley circulation is weakest during boreal summer, which is
102 completely out of synchronization with some of the observed summer climate features along the
103 NH subtropics (e.g., the co-existence of both moist monsoon and desert climates at the same
104 latitude). Consequently, this traditional view is now not well accepted in the literature (Yang et
105 al., 1992; Rodwell and Hoskins, 1996; Wang, 2006; Wu et al., 2009). In a pioneering study,
106 Charney (1975) made an attempt to explain the enhancement of subsidence over subtropical
107 desert of the NH during boreal summer, by invoking a mechanism called biosphere-albedo
108 feedback. As per their study, the pronounced reduction in vegetation cover (i.e., over-grazing)
109 over subtropical deserts regions increases surface albedo, thus causing enhanced radiative
110 cooling (Fig. 15.1c). This radiative loss is balanced by enhanced descent (Fig. 15.2), which in
111 turn results in reduced rainfall, thus leading to further decrease in vegetation cover (i.e.,
112 desertification amplification). However, their mechanism ignored the possible influence of

113 horizontal heat advection and hence found to be inappropriate over the NH subtropical deserts
114 (e.g., Hoskins, 1986). Later, Yang et al. (1992) and Webster (1994) proposed another concept
115 involving a closed “Walker type” circulation linking convection over south Asia to the
116 subsidence over arid regions to its west. Nevertheless, subsequent studies (i.e., the seminal work
117 of Rodwell and Hoskins, 1996, 2001; Hoskins, 1986) dispelled this notion, as their studies
118 showed no signature of a closed overturning circulation. Based on the scale analysis of the
119 thermodynamic equation, Rodwell and Hoskins (1996) also argued for the importance of
120 horizontal heat advection for the existence of these subtropical desert regions. This currently
121 recognized paradigm is often referred in the literature as the “monsoon-desert mechanism”, and
122 is further described in Section 15.3.

123



124

125 **Figure 15.2:** Vertical cross-section of atmospheric circulation in terms of vertical component of
126 velocity ($10^{-2} \text{ Pa s}^{-1}$, shaded) and horizontal wind divergence (10^{-6} s^{-1} , contours) during July,
127 along a pressure-longitude plane averaged over two latitude bands (a) 20°N and (b) 40°N . The
128 circulation fields are taken from the ERA-Interim reanalysis (Dee et al., 2011) and the
129 climatology is computed for the 1986-2014 period. The negative (dashed) and positive
130 (continuous) contours correspond, respectively, to absolute magnitudes of 1, 2, 3 and 4 units.
131 The zero contours are highlighted in thick black color. Negative (positive) shading implies
132 ascending (descending), while negative (positive) contours implies convergence (divergence).
133 The presentation using July climatology follows the observational conjecture that it corresponds
134 to the peak of monsoon activity (e.g., Tyrlis et al., 2013).

135

136 In contrast to the subtropical desert, the vegetated land surface over south Asian landmass
137 shows a reduced surface albedo and the ISM region is radiatively surplus highlighting the role of
138 clouds and moisture on the radiation balance (Fig. 15.1b-c; Sooraj et al., 2019). Furthermore, the
139 ISM land region and the Bay of Bengal are characterized by strong ascending motion extending
140 throughout the troposphere (see Fig. 15.2a), thus exemplifying the monsoon induced
141 stratification due to the ISM rainfall and its associated diabatic heating (Fig. 15.1a). The
142 corresponding low-level convergence is consistent with the thermo-dynamical response due to
143 this diabatic heating (Fig. 15.2a; e.g., Neelin and Held, 1987).

144 The transition zone between this ISM convection center and the hot subtropical deserts to
145 its west (i.e., extending from Mediterranean to west Asian landmass) is characterized by the
146 existence of several regional heat lows with lower-level convergence and ascending motions
147 (i.e., below 600-hPa) capped aloft by upper-level subsidence and divergence (i.e., mostly

148 confined between 200 and 600-hPa) as displayed in Figure 15.2a. Figure 15.2b shows the
149 intensification of these descending motions (extending throughout the troposphere) over west
150 Asia and eastern Mediterranean regions at the northern boundary of the domain, thus
151 highlighting the pronounced subsidence over these regions. Figure 15.2a-b highlights that the
152 southern and northern regions of the hot subtropical deserts show totally different vertical
153 structures of the local atmospheric circulation (Sooraj et al., 2019).

154 **15.3 Monsoon influence over hot subtropical deserts**

155 The idea that the existence of subtropical deserts is related to the ISM is dated back to
156 mid-90s. Rodwell and Hoskins (1996) suggested a monsoon-desert mechanism whereby, in a
157 linear modelling framework, remote diabatic heating in the Asian monsoon region induces a
158 Rossby-wave pattern to the west, interacting with the southern flank of the mid-latitude
159 westerlies and causing descent over eastern Sahara and the Mediterranean (Figs. 15.2 and 15.3).
160 The associated atmospheric response has the form of an anticyclone at upper levels and a
161 cyclone at lower levels, with a deepening of the isentropic surfaces due to the mid-tropospheric
162 warming (Fig. 15.3a). When this far poleward thermal structure interacts with the southern flank
163 of the mid-latitude westerlies, the air moves down the isentropes on their western side (Wang et
164 al., 2012), as evident from Figure 15.3b where the northerly component of the westerly flow
165 crosses the isentropic surfaces. Further, according to Rodwell and Hoskins (1996), the subsiding
166 air is also of mid-latitude origin as revealed by their trajectory analysis. Regions of adiabatic
167 descent are strengthened by “local diabatic enhancement” and longitudinal mountain chains
168 induce a blocking of the westerly flow (Rodwell and Hoskins, 2001). The idea of the monsoon-
169 desert mechanism linking the ISM to the Mediterranean has been then easily extended to other
170 regions with Mediterranean-type of climates, like California and Chile, remotely forced by

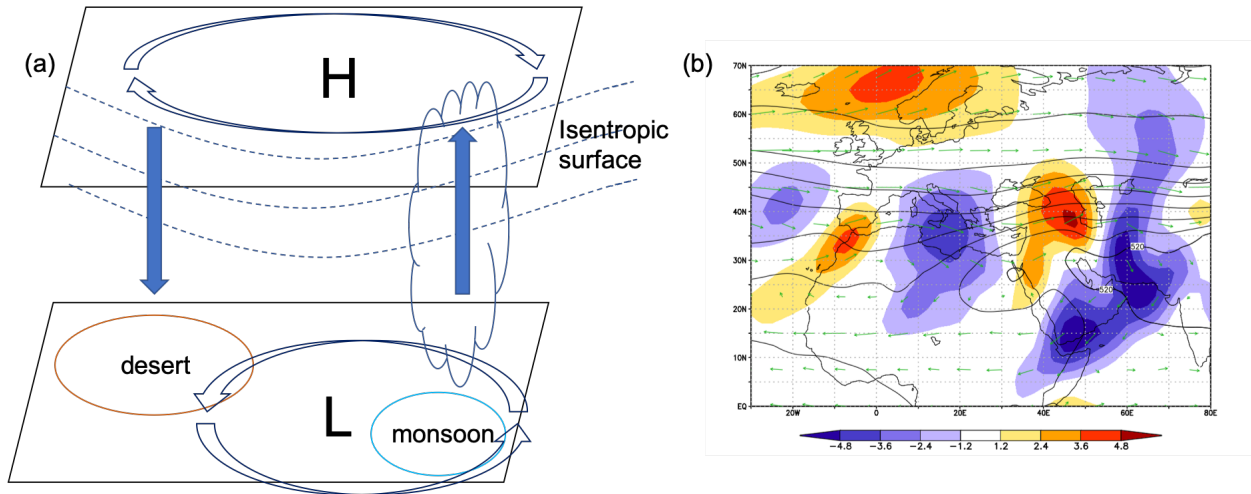
171 monsoon regions to the east (Rodwell and Hoskins, 2001). Overall, the theoretical framework
172 based on these pioneering works is well established in terms of the origin of subtropical highs in
173 the summer hemispheres (Cherchi et al., 2018).

174 In agreement with idealized model simulations, results focused on ERA40 reanalysis
175 (Uppala et al., 2005) confirmed strong mid and upper-level warming spreading westward and
176 causing significant depression of the isentropes, which are linked to Rossby wave activity away
177 from the diabatic heating sources (Tyrlis et al., 2013). Over the eastern Mediterranean and
178 Middle East, the subsidence and northerly flow (Etesian winds) are recognized as manifestations
179 of the Rossby wave structure triggered by the ISM convection (Tyrlis et al., 2013; Rizou et al.,
180 2018). In a similar framework, processes at work in the monsoon-desert mechanism have been
181 analyzed in 20th century simulations from coupled GCMs participating to the 5th Coupled Model
182 Intercomparison Project (CMIP5; Taylor et al., 2012), showing how few of them are able to
183 simulate the mechanism for the correct reasons (Cherchi et al., 2014). CMIP5 coupled GCMs
184 tend to underestimate (overestimate) ISM related diabatic heating at upper (lower) levels,
185 resulting in a weaker forced response (Cherchi et al., 2014). Moreover, CMIP5 coupled GCMs
186 with a severe dry bias over the ISM region depict the weakest mid-latitude winds and minimum
187 descent over the Mediterranean, as well as little vertical variation in diabatic heating (Cherchi et
188 al., 2014). On the other hand, when simulated precipitation over India improves in the coupled
189 GCMs, the extratropical teleconnection with the Mediterranean region improves as well (Jin et
190 al., 2019). In other words, the uncertainty associated with monsoon simulations needs to be
191 considered in future climate projections even outside the monsoon domain. Interestingly, air-sea
192 coupling seems also to be important to correctly represent the monsoon-desert teleconnections,
193 as in coupled model simulations the divergence field associated with the ISM is more favorable

194 for inducing westward Rossby wave propagation than in uncoupled simulations (Osso' et al.,
195 2019).

196

197



198

199 **Figure 15.3:** (a) Schematic diagram of the monsoon desert-mechanism with descent over the
200 desert region (orange) induced by the heating over the monsoon convective region (cyan)
201 interacting with local westerlies. [Adapted by permission from Wang et al., 2012]. (b) JJA mean
202 wind (ms^{-1} , green vectors), meridional wind velocity (ms^{-1} , shaded), and pressure levels (hPa,
203 contours with 50-hPa contour interval) on isentropic surfaces at 325K using NCEP/NCAR
204 reanalysis dataset (Kalnay et al., 1996). The climatology is computed for the period 1958-2019.

205

206 Within boreal summer, descent enhancement over the Mediterranean and east Sahara has
207 been observed during the onset of the ISM (Rodwell and Hoskins, 1996). In early July, when
208 convection migrates over northern continental India, the Rossby wave structure amplifies
209 impacting the circulation over the eastern Mediterranean and Middle East (Tyrlis et al., 2013). In
210 the peak of the monsoon season, the combined diabatic heating pattern over the Arabian Sea and

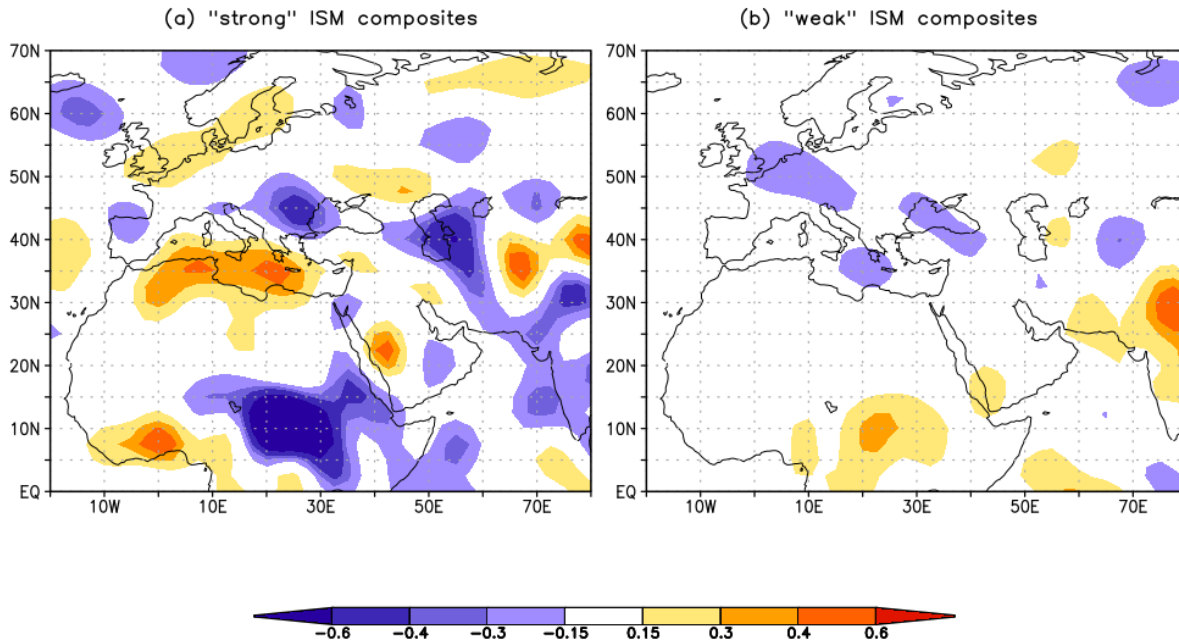
211 Bay of Bengal regions exerts the largest descent over the eastern Mediterranean (Cherchi et al.,
212 2014). As the monsoon activity is associated with active and break spells, it could be argued that
213 successive Rossby wave pulses are released, inducing a cumulative effect over the region until
214 the peak of the monsoon activity is reached in July (Tyrlis et al., 2013).

215 The processes recognized as part of the monsoon-desert mechanism vary at interannual
216 timescales as well, with imprints of severe weak and strong ISMs noticeable over the
217 Mediterranean (Cherchi et al., 2014). Interestingly, the monsoon forcing is more significant
218 during strong ISMs, with enhanced subsidence over the Mediterranean region, than during weak
219 ISMs (see Fig. 15.4). During a strong monsoon, upper tropospheric heating expands westward
220 reaching a maximum over northern Arabian Peninsula modulating thermodynamic and dynamic
221 patterns there and likely governing the occurrence of extremes over the region (Attada et al.,
222 2019). ISM rainfall is significantly and negatively correlated with precipitation over Black
223 Sea/Balkans region (Osso' et al., 2019). This kind of relationships suggests the possibility of
224 using the ISM characteristics as potential predictors for the summer conditions over the
225 Mediterranean region. At interannual to inter-decadal timescales, these dynamics are part of a
226 larger picture involving North Africa (Liu et al., 2001; Wu et al., 2009; He et al., 2017) and
227 portions of Eurasia via the so-called “silk road pattern” and “circumglobal teleconnection” (Lu et
228 al., 2002; Enomoto et al., 2003; Ding and Wang, 2005; Saeed et al., 2014; Wang et al., 2017;
229 Stephan et al., 2019, also see Chapter 13 and 14).

230 The monsoon-desert mechanism as first established using a linear modelling framework,
231 is also supported by relating prehistoric lake-levels to Milankovitch-monsoon forcing (Rodwell
232 and Hoskins, 1996) and extrapolated to the existence of “mega-deserts” and “mega-monsoon”
233 areas in specific periods (Wang et al., 2014). Evidences of moist events over east Sahara have

234 been observed (Gaven et al., 1981; Kowalski et al., 1989), consistent with shutting off of the
235 monsoon-desert mechanism (Quade et al., 2018). Some of those wet events have been dated to
236 correspond to minimum NH summertime insolation, linked with minimum intensity of ISM
237 (Clemens et al., 1991). Some of these arguments have been confirmed with GCM studies
238 (Claussen, 1994; Perez-Sanz et al., 2014).

239 As the ISM rainfall is projected to increase, the monsoon-desert teleconnection is
240 consistent with the changes in 21st century projections over the Mediterranean region, at least for
241 its central part corresponding to the largest increase in subsidence (Cherchi et al., 2016).
242 However, other factors at play, i.e. like the warming and moistening of the troposphere under
243 climate change, which enhance significantly the greenhouse effect over the deserts (Zhou, 2016;
244 Wei et al., 2017), the North Atlantic Oscillation (Blade et al., 2012; Kalimeris and Kolios, 2019)
245 or the changes in frequency/intensity of blocking systems (Masato et al., 2013; Tyrlis et al.,
246 2015), may influence and shape the response in future climate projections. In future projections,
247 the local atmospheric dynamics contribute in maintaining local temperature and precipitation
248 balance over eastern and southern Mediterranean regions, suggesting a stronger influence of land
249 surface warming on local atmospheric circulation and progressing desertification (Zhou, 2016;
250 Wei et al., 2017; Barcikowska et al., 2020). In a climate change perspective, the influence from
251 ISM on the Mediterranean region may have consequences also on the sea surface characteristics
252 in terms of temperature and ecosystems (Kim et al., 2019).



253
 254 **Figure 15.4:** Composite of omega at 500 hPa (units hPa hr⁻¹) for (a) strong and (b) weak
 255 monsoon years for the boreal summer monsoon period (June to September), using NCEP/NCAR
 256 reanalysis dataset (Kalnay et al., 1996) for the period 1958-2019. Strong years are 1959, 1961,
 257 1970, 1975, 1983, 1988, 1994 and weak years are 1965, 1966, 1968, 1972, 1974, 1979, 1982,
 258 1985, 1986, 1987, 2002, 2004, 2009, 2014, 2015 (based on normalized All India Rainfall (AIR)
 259 index exceeding 1 standard deviation). The AIR index is taken from
 260 <https://www.tropmet.res.in/~kolli/MOL/Monsoon/Historical/air.html>.

261 **15.4 Potential role of deserts in modulating ISM**

262 The “monsoon-desert mechanism” described in the previous section, perceives the
 263 subtropical deserts of the NH to be a rather “passive” recipient in the relationship. However,
 264 from a different perspective, it is also argued that heat lows, changes of surface heating over the
 265 subtropical deserts and dry air intrusions (whether natural or anthropogenic) from arid regions
 266 into the monsoon domain can exert a significant control on the WAM and ISM systems, at both

267 the intra-seasonal and seasonal time scales (Charney, 1975; Charney et al., 1977; Shukla and
268 Mintz, 1982; Sud and Fennessy, 1982; Sud et al., 1988; Claussen, 1997; Bonfils et al., 2000;
269 Douville et al., 2001; Xue et al., 2004; Yasunari et al., 2006; Xue et al., 2010; Krishnamurti et
270 al., 2010; Bollasina and Nigam, 2011a, 2011b; Bollasina and Ming, 2013; Parker et al., 2016;
271 Terray et al., 2018; Sooraj et al., 2019). For example, Sahelian Heat Low plays a key-role in the
272 WAM evolution through its low-level cyclonic circulation and induced moisture convergence,
273 and this is well established in observations, reanalysis and GCM simulations (Haarsma et al.,
274 2005; Biasutti et al., 2009; Lavaysse et al., 2009; Shekhar and Boos, 2017). On similar lines,
275 ISM, especially its onset and early part, is known to be influenced by the atmospheric variability
276 over the subtropical desert regions (i.e., northwest part of India and west central Asia; Ramage,
277 1966; Smith, 1986a, 1986b; Mooley and Paolino 1988; Parthasarathy et al., 1992; Bollasina and
278 Nigam, 2011b; Bollasina and Ming, 2013; Vinoj et al., 2014; Rai et al., 2015; Sooraj et al.,
279 2019). During boreal summer, the monsoon depressions formed over the Bay of Bengal move
280 northwestward across the Indian landmass, and eventually merge and dissipate in the heat-low
281 region (Wang, 2006; Krishnamurti et al., 2013). The relationship between this seasonal heat low
282 and ISM has been examined in perpetual boreal spring atmospheric GCM (i.e., AGCM)
283 experiments by Bollasina and Ming, (2013). Remarkably, they found that the surface heating
284 (i.e., with no seasonal variation of insolation in their perpetual experiments) over northwestern
285 semi-arid areas is able to uniquely control the northwestward migration of ISM rainfall at
286 seasonal time scale and of the monsoon depressions at intra-seasonal time scale.

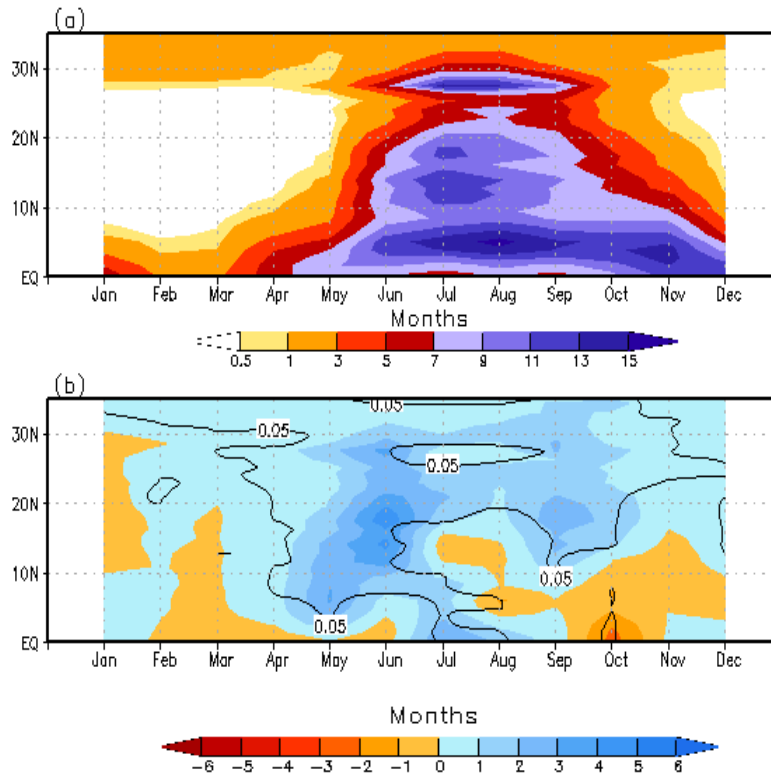
287 Furthermore, recent modeling studies show increasing evidence of the impact of dust
288 aerosols on the ISM at various time scales (i.e., from weekly to decadal) through aerosol
289 radiative effects (Ramanathan et al., 2005; Lau et al., 2006; Meehl et al., 2008; Wang et al.,

290 2009; Nigam and Bollasina, 2010; Bollasina et al., 2011; Vinoj et al., 2014; Jin et al., 2014,
291 2016a, 2016b; Solmon et al., 2015). For example, Vinoj et al. (2014) using AGCM sensitivity
292 experiments argue that the presence of dust loading over North Africa and West Asian arid
293 regions induced atmospheric heating over there, thus promoting abundant moisture transport into
294 the ISM region, with associated enhanced rainfall (e.g., over Monsoon Core Region, MCZ).
295 Thus, their study basically showed that the arid regions can also act as a dust induced
296 atmospheric heat source (at least in a relative sense) during boreal summer and modulate ISM
297 rainfall at the synoptic time scale. This is consistent with earlier studies (Smith 1986a, 1986b;
298 Mohalifi et al., 1998), and also in line with other recent studies focusing on the impact of Middle
299 East dust aerosols on ISM rainfall (Jin et al., 2014, 2015, 2016a; Solmon et al., 2015). However,
300 the rainfall responses in each of these studies display heterogeneous distributions both in space
301 and time (Solmon et al., 2015; Jin et al., 2016b; Sanap and Pandithurai, 2015). For example,
302 Solmon et al. (2015) conducted a study (using a regional climate model) to examine the
303 interaction between Middle East dust and ISM. Similar to Vinoj et al. (2014), they found an
304 increase in rainfall restricted to the southern part of India with other parts, and central and
305 northern India showing a significant decrease. Recently, Kumar and Arora, (2019) also claimed
306 that enhanced dust forcing and associated warming over the Arabian Sea are unlikely to create a
307 positive feedback on ISM rainfall, because of its limited spatial extent. To sum up, it is not
308 altogether clear whether this part of the theory focusing on the role of Middle East dust on ISM
309 is in definitive form.

310 This framework based on the relative heat source concept has also been extended to other
311 time scales and to the ISM onset (Rai et al., 2015; Chakraborty and Agrawal, 2017; Samson et
312 al., 2017; Terray et al., 2018; Sooraj et al., 2019). In particular, Samson et al. (2017) and Terray

313 et al. (2018) demonstrated the sensitivity of NH monsoon regions (in particular African-Asian
314 monsoon) to the land surface thermal forcing over these arid regions. Using regional and global
315 coupled GCMs, respectively, they modified the land surface albedo using up-to-date satellite
316 estimates. According to them, the persistent tropical rainfall errors in current coupled GCMs are
317 partly associated with insufficient surface thermal forcing and incorrect representation of the
318 surface albedo over the NH continents. Improving the parameterization of the land albedo in a
319 regional coupled model (Samson et al., 2017) and two global coupled GCMs (i.e., SINTEX;
320 Masson et al., 2012 and CFSv2; Saha et al., 2014; Terray et al., 2018), leads to a significant
321 reduction of the model systematic dry bias over land. They further showed that African-Asian
322 monsoon circulation is, partly, a response to the large-scale pressure gradient between the hot
323 NH subtropical deserts and the relatively cooler oceans to the south. A concept, which may have
324 implications for the ISM response in the context of climate change as desertification has been
325 identified as a robust feature of global warming (Zhou, 2016; Wei et al., 2017).

326 In a companion modeling study, Sooraj et al. (2019) found that ISM evolution and
327 intensity are significantly affected with opposite polarity to prescribed negative and positive
328 surface albedo perturbations over the whole hot subtropical desert lying to the west of ISM,
329 including the remote Sahara Desert. This is consistent with the hypothesis that the arid regions
330 can also act as a relative heat source and modulate the ISM, but also the whole tropical climate
331 during boreal summer (Terray et al., 2018). The darkening of the deserts (negative albedo
332 perturbations) leads to advancement of the ISM onset by one month, with a rapid northward
333 propagation of the rainfall band over the Indian domain (see Fig. 15.5). The brightening of the
334 deserts (i.e., positive albedo perturbation) shows non-linear response in ISM rainfall and
335 circulation with significantly larger amplitude (figure not shown).



336

337

338 **Figure 15.5:** (a) Time-latitude evolution of rainfall climatology (mm day^{-1}) averaged along 70°-
 339 90°E (over both ocean and land points) from a control simulation performed with the SINTEX-
 340 F2 coupled model (Masson et al., 2012). (b) time-latitude evolution of anomalous rainfall
 341 response averaged along 70°-90°E from “Desert_m20” sensitivity experiment (see below). In
 342 (b), the responses that are above the 95% confidence level according to a permutation procedure
 343 with 9999 shuffles are encircled. “Desert_m20” is a sensitivity experiment performed with the
 344 SITEX-F2 coupled model in which the background land albedo has been artificially decreased
 345 by -20% over the whole hot subtropical desert extending up to the Sahara in the west (see
 346 highlighted box in Figure 15.1). [Adpated by permission from Sooraj et al., 2019]

347 Whilst the processes highlighted above (i.e., surface thermal forcing over arid regions)
 348 are mostly demonstrated at seasonal time scale, a recent observational study shows the evidence
 349 at the daily timescale as well (Sooraj et al., 2020). While examining the ISM rainfall extremes

350 over the MCZ region at daily time scale, these authors found that a similar surface thermal
351 signature over Indo-Pakistan Arid Region (IPAR) is among the best precursors of the ISM wet
352 daily extremes at longer lead times (e.g., one or two weeks in advance), outperforming Sea
353 Surface Temperature precursors based on the Monsoon Intra-Seasonal Oscillation (e.g., Wang et
354 al., 2005, Roxy et al., 2017). With the additional role of moist processes, the study revealed the
355 surprising existence of a strong water vapour positive feedback over the drier IPAR, at
356 intraseasonal and daily time scales, in line with the strong longwave greenhouse effect as found
357 over arid regions in the context of global warming (Zhou, 2016; Wei et al., 2017). Further,
358 according to their study, the enhanced surface warming and amplified water vapor feedbacks
359 may be another significant contributor to the recent increasing trend in rainfall wet extremes over
360 Indian landmass. In the backdrop of global warming, the results from Sooraj et al. (2020) may
361 have larger implications concerning the changes in ISM rainfall extremes as well as their
362 potential predictability.

363 **15.5 Summary and future perspectives**

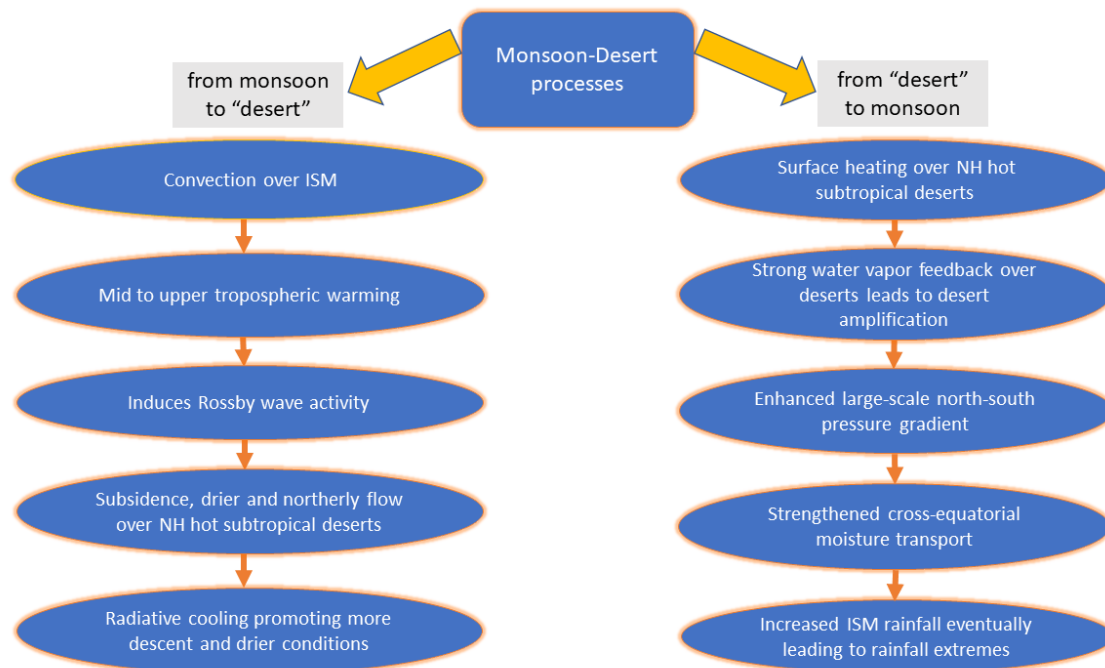
364 This chapter presents a comprehensive review on the monsoon-desert system focusing on
365 the mutual relationships between these two contrasting climates, which, surprisingly, coexist at
366 the same latitude of the NH, and highlights the underlying mechanisms. Such a review assumes
367 significance, as recently there is a renewed interest to understand the close relationships between
368 the two systems since both of them are expected to be severely affected by anthropogenic
369 climate change in the forthcoming decades.

370 Firstly, the literature focusing on the monsoon influence on subtropical deserts is
371 reviewed by highlighting the monsoon-desert mechanism whereby convection over ISM (and
372 also the neighboring Bay of Bengal) induces strong descending motions over the NH subtropical

373 deserts (i.e., to the west of ISM region), with the local subsidence and northerly flow (Etesian
374 winds) being manifestations of the Rossby wave structure associated with ISM convection (Fig.
375 15.3). These different processes are further summarized in Figure 15.6. Hence, the theoretical
376 framework, as suggested by earlier pioneering works (Rodwell and Hoskins, 1996, 2001), is now
377 well established from more recent observational and modeling studies (e.g., Uppala et al., 2005;
378 Tyrlis et al., 2013; Cherchi et al., 2014; Cherchi et al., 2018; Jin et al., 2019). Additional
379 evidences are also presented to show that the monsoon-desert mechanism and the underlying
380 processes operate at interannual timescales as well (Cherchi et al., 2014). However, most of the
381 state-of-the-art coupled GCMs are unable to simulate this monsoon-desert relationship, partly
382 due to the systematic monsoon dry bias affecting many current GCMs. For example, CMIP5
383 coupled GCMs with a severe dry bias over the ISM region depict the minimum descent over the
384 subtropical deserts (Cherchi et al., 2014). In future projections, the monsoon-desert relationship
385 remains elusive with ISM affecting mostly the central part of the Mediterranean region (Cherchi
386 et al., 2016), despite that desertification and ISM rainfall are both projected to increase during
387 the 21th century (Zhou, 2016; Wei et al., 2017; Wang et al., 2020). But, assessing the robustness
388 of this counter-intuitive signal is challenging for several reasons. For example, it is not known if
389 this paradox is related to the large uncertainty affecting ISM simulations (i.e., large biases in
390 ISM mean characteristics as mentioned above) in current coupled GCMs or if it is the sign of the
391 increasing role of radiative or surface land processes in shaping the climate of the deserts under
392 global warming. Furthermore, the subtropical deserts are also affected by large temperature and
393 radiation biases in current coupled GCMs (e.g., Samson et al., 2016; Terray et al., 2018). To sum
394 up, both significant modeling developments, and an exhaustive analysis of all the external

395 forcings at play (Blade et al., 2012; Kalimeris and Kolios, 2019; Masato et al., 2013; Tyrlis et al.,
396 2015) are required to assess the fate of the monsoon-desert paradigm in our future world.

397 Secondly, we offered a different perspective by emphasizing the potential role of deserts
398 in modulating ISM either through the changes of surface heating over the subtropical deserts or
399 related to dry air intrusions from arid regions into the ISM domain (e.g., Krishnamurti et al.,
400 2010; Bollasina and Nigam, 2011a, 2011b; Bollasina and Ming, 2013; Vиноj et al., 2014; Jin et
401 al., 2014, 2016a; Solmon et al., 2015; Parker et al., 2016; Terray et al., 2018; Sooraj et al., 2019).
402 Most of these studies, based on numerical modeling frameworks, show that the surface land
403 warming paves the way for enhanced ISM rainfall through increased southwesterly monsoon
404 flow and hence moist transport into the Indian continent. The associated processes are further
405 summarized in Figure 15.6. For example, recent studies using fully coupled GCMs found that by
406 darkening/brightening the subtropical deserts, the seasonality of ISM could be significantly
407 modified (Fig. 15.5; e.g., Samson et al., 2017; Terray et al., 2018; Sooraj et al., 2019).
408 Furthermore, a recent observational study provided new insights of the role of subtropical deserts
409 (e.g., surface thermal forcing over IPAR) on the subseasonal modulations of ISM, and the
410 occurrence of ISM daily rainfall extremes (Sooraj et al., 2020). These new studies challenge both
411 the passive role of the desert as described in the monsoon-desert paradigm and the view that the
412 lower tropospheric thermal contrast has only a minor role in the ISM evolution at different time
413 scales from days to decades (Dai et al., 2013).



414

415 **Figure 15.6:** Schematic diagram summarizing the various interactive process and mechanisms in
 416 the monsoon-desert system over the African-Asian continent (i.e., from the monsoon to the
 417 “deserts” on the left and from the “deserts” to the monsoon on the right).

418 The subtropical deserts are particularly vulnerable to climate change (i.e., desert
 419 amplification with enhanced warming signal especially over the Sahara, the Arabian Peninsula
 420 and Middle-East) as suggested by several recent studies (e.g., Cook and Vizy, 2015; Zhou,
 421 2016). The warming and moistening of the troposphere under climate change may in turn
 422 influence the local monsoon atmospheric circulation and, thus, eventually affect the monsoon-
 423 desert teleconnection in future climate. Further, the enhanced surface warming and amplified
 424 water vapor feedbacks over the arid regions adjacent to ISM (Sooraj et al., 2020) may have also
 425 significant implications for changes in frequency and intensity of ISM rainfall daily extremes
 426 under global warming scenario. However, as for the monsoon-desert mechanism, elucidating in a
 427 more robust way the potential role of the subtropical deserts on ISM will require both additional

428 analysis of observations and improved GCMs with reduced rainfall biases, more detailed
429 parameterizations for aerosols and improvements of the radiation codes embedded in these
430 models.

431

432

433 **Conflict of interest statement**

434 The corresponding author states no conflict of interest on behalf of all authors.

435

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737

738 **Figure captions:**

739 **Figure 15.1:** Climatological map of (a) rainfall (mm day^{-1}), (b) land surface albedo (%) and (c)
740 net radiation budget at TOA (Wm^{-2}), for boreal summer period (June to September). In (a)
741 rainfall climatology is computed for the 1986-2014 period from Global Precipitation
742 Climatology Project (GPCP version 2.1; Huffman et al., 2009). In (b) and (c), albedo and
743 radiation climatology is computed for the 2000-2018 period from the Clouds and the Earth's
744 Radiant Energy System Energy Balanced and Filled (CERES-EBAF edition 4.0; Kato et al.,
745 2018). In (b) and (c), the region highlighted in the black rectangle refers to “subtropical deserts”.

746 **Figure 15.2:** Vertical cross-section of atmospheric circulation in terms of vertical component of
747 velocity ($10^{-2} \text{ Pa s}^{-1}$, shaded) and horizontal wind divergence (10^{-6} s^{-1} , contours) during July,
748 along a pressure-longitude plane averaged over two latitude bands (a) 20°N and (b) 40°N . The
749 circulation fields are taken from the ERA-Interim reanalysis (Dee et al., 2011) and the
750 climatology is computed for the 1986-2014 period. The negative (dashed) and positive
751 (continuous) contours correspond, respectively, to absolute magnitudes of 1, 2, 3 and 4 units.
752 The zero contours are highlighted in thick black color. Negative (positive) shading implies
753 ascending (descending), while negative (positive) contours implies convergence (divergence).
754 The presentation using July climatology follows the observational conjecture that it corresponds
755 to the peak of monsoon activity (e.g., Tyrlis et al., 2013).

756 **Figure 15.3:** (a) Schematic diagram of the monsoon desert-mechanism with descent over the
757 desert region (orange) induced by the heating over the monsoon convective region (cyan)
758 interacting with local westerlies. [Adapted by permission from Wang et al., 2012]. (b) JJA mean
759 wind (ms^{-1} , green vectors), meridional wind velocity (ms^{-1} , shaded), and pressure levels (hPa,

760 contours with 50-hPa contour interval) on isentropic surfaces at 325K using NCEP/NCAR
761 reanalysis dataset (Kalnay et al., 1996). The climatology is computed for the period 1958-2019.

762 **Figure 15.4:** Composite of omega at 500 hPa (units hPa hr⁻¹) for (a) strong and (b) weak
763 monsoon years for the boreal summer monsoon period (June to September), using NCEP/NCAR
764 reanalysis dataset (Kalnay et al., 1996) for the period 1958-2019. Strong years are 1959, 1961,
765 1970, 1975, 1983, 1988, 1994 and weak years are 1965, 1966, 1968, 1972, 1974, 1979, 1982,
766 1985, 1986, 1987, 2002, 2004, 2009, 2014, 2015 (based on normalized All India Rainfall (AIR)
767 index exceeding 1 standard deviation). The AIR index is taken from

768 <https://www.tropmet.res.in/~kolli/MOL/Monsoon/Historical/air.html>.

769 **Figure 15.5:** (a) Time-latitude evolution of rainfall climatology (mm day⁻¹) averaged along 70°-
770 90°E (over both ocean and land points) from a control simulation performed with the SINTEX-
771 F2 coupled model (Masson et al., 2012). (b) time-latitude evolution of anomalous rainfall
772 response averaged along 70°-90°E from “Desert_m20” sensitivity experiment (see below). In
773 (b), the responses that are above the 95% confidence level according to a permutation procedure
774 with 9999 shuffles are encircled. “Desert_m20” is a sensitivity experiment performed with the
775 SITEX-F2 coupled model in which the background land albedo has been artificially decreased
776 by -20% over the whole hot subtropical desert extending up to the Sahara in the west (see
777 highlighted box in Figure 15.1). [Adpated by permission from Sooraj et al., 2019]

778 **Figure 15.6:** Schematic diagram summarizing the various interactive process and mechanisms in
779 the monsoon-desert system over the African-Asian continent (i.e., from the monsoon to the
780 “deserts” on the left and from the “deserts” to the monsoon on the right).