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2 Roles of land surface albedo and horizontal resolution
3 on the Indian Summer Monsoon biases
4 in a coupled ocean-atmosphere tropical-channel model
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32 **Abstract**

33 The Indian Summer Monsoon (ISM) simulated over the 1989-2009 period
34 with a new 0.75° ocean-atmosphere coupled tropical-channel model extending from
35 45°S to 45°N is presented. The model biases are comparable to those commonly
36 found in coupled global climate models (CGCMs): the Findlater jet is too weak,
37 precipitations are underestimated over India while they are overestimated over the
38 southwestern Indian Ocean, South-East Asia and the Maritime Continent. The ISM
39 onset is delayed by several weeks, an error which is also very common in current
40 CGCMs.

41 We show that land surface temperature errors are a major source of the ISM
42 low-level circulation and rainfall biases in our model: a cold bias over the Middle-
43 East (ME) region weakens the Findlater jet while a warm bias over India strengthens
44 the monsoon circulation over the southern Bay of Bengal. A surface radiative heat
45 budget analysis reveals that the cold bias is due to an overestimated albedo in this
46 desertic ME region. Two new simulations using a satellite-observed land albedo show
47 a significant and robust improvement in terms of ISM circulation and precipitation.
48 Furthermore, the ISM onset is shifted back by one month and becomes in phase with
49 observations. Finally, a supplementary set of simulations at 0.25°-resolution confirms
50 the robustness of our results and shows an additional reduction of the warm and dry
51 bias over India. These findings highlight the strong sensitivity of the simulated ISM
52 rainfall and its onset timing to the surface land heating pattern and amplitude,
53 especially in the ME region. It also illustrates the key-role of land surface processes
54 and horizontal resolution for improving the ISM representation, and more generally
55 the monsoons, in current CGCMs.

56

57

58 **Keywords**

59 Indian Summer Monsoon; land surface albedo; horizontal resolution;

60 precipitation biases; monsoon onset; CGCM

61 **1. Introduction**

62

63 The Indian Summer Monsoon (ISM; see Table 1 for acronyms) brings
64 substantial rainfall from June to September to some of the world most populated
65 regions, whose economy relies mainly on agriculture and water resources. But despite
66 recent progress in our understanding of mechanisms driving ISM precipitation,
67 Coupled General Circulation Models (CGCMs) are still not able to correctly represent
68 its main spatial and temporal characteristics (Sperber et al. 2013) and the skill of
69 seasonal ISM predictions by dynamical or statistical models remains currently very
70 low, contrary to what is observed in other tropical regions (Wang et al. 2015).

71 While some improvements have been achieved with the last generation of
72 CGCMs, especially in terms of intraseasonal variability (Abhik et al. 2014, Sabeerali
73 et al. 2013, Goswami et al. 2014), some basic features of the ISM, such as the onset or
74 the rainfall spatial distribution, are still poorly captured with a persisting (wet) dry
75 bias over (ocean) land (see Fig. 2 of Sperber et al. 2013).

76 The limited horizontal resolution of CGCMs is frequently listed as a major
77 caveat because current coarse atmospheric models cannot properly resolve orography
78 (Wu et al. 2002, Chakraborty et al. 2002, Cherchi and Navarra 2007, Boos and Hurley
79 2013), intraseasonal oscillations (Saha et al. 2013), tropical disturbances (Sabin et al.
80 2013) or convection (Pattnaik et al. 2013, Ganai et al. 2015), which all significantly
81 contribute to the total ISM rainfall, especially in the monsoon trough region.

82 Regional Climate Models (RCMs) allow simulating the ISM at higher
83 resolutions than global CGCMs, but with a strong control of the lateral boundaries
84 imposed to the RCMs. This allows to distinguish the effects of local versus remote
85 forcings on the ISM (Seo et al. 2009, Samala et al. 2013), to test the sensitivity of the
86 simulated ISM to different physical parameterizations (Mukhopadhyay et al. 2010,
87 Srinivas et al. 2013, Samson et al. 2014) or to prescribe the orography in a more
88 realistic way (Ma et al. 2014). But despite those specificities, significant biases still
89 exist in terms of precipitation and surface temperature (Lucas-Picher et al. 2011),
90 which suggest that high resolution is not the unique missing ingredient in order to
91 improve ISM rainfall in current CGCMs and RCMs.

92 Local and remote Sea Surface Temperature (SST) errors, amplified by ocean-
93 atmosphere coupling, can also adversely affect the coupled model performance in
94 simulating the ISM rainfall or its onset timing, and has gained a lot of attention in
95 recent years (Bollasina and Nigam 2009, Bollasina and Ming 2013, Levin and Turner
96 2012, Joseph et al. 2012, Prodhomme et al. 2014, 2015; among others). Common SST
97 biases have been clearly identified in many CGCMs and their consequences on the
98 ISM have been addressed in these studies. However, the specific origins of these SST
99 errors are not well understood, they may vary from one CGCM to another and they
100 cannot account alone for the ISM rainfall errors in current CGCMs (Prodhomme et al.
101 2014, 2015, Li et al. 2015).

102 Less attention has been paid to large-scale long-standing biases such as land
103 temperature errors, which can also influence the ISM simulation (Christensen et al.
104 2007, Lucas-Picher et al. 2011, Boos and Hurley 2013). The ISM onset timing
105 primarily depends on the meridional land-sea thermal contrast between the Indian
106 subcontinent and the tropical Indian Ocean (IO) (Li and Yanai 1996, He et al. 2003,
107 Xavier et al. 2007, Prodhomme et al. 2015). Consequently, models errors on Land
108 Surface Temperature (LST) can directly influence the ISM onset characteristics
109 (Prodhomme et al. 2015). Aside from the onset timing, the ISM structure and intensity
110 also depends on the meridional Tropospheric Temperature (TT) gradient, which relies
111 on both surface local and remote heat sources during boreal summer (Wu et al. 2009,
112 Bollasina and Nigam 2011, Dai et al. 2013). Hence, LST and TT biases can also
113 influence the ISM representation in the models. However, many studies suggest that
114 the orographic effects, mountains and TT errors are stronger than the direct impact of
115 the land surface heating on the ISM (He et al. 2003, Bollasina and Nigam 2011,
116 Molnar et al. 2010, Boos 2015). Especially, Boos and Kuang (2010, 2013) showed by
117 changing the Tibetan plateau albedo that this region is not a dominant heating source
118 for the atmosphere during the ISM, but rather a good insulator preventing mixing
119 between tropical warm and humid air with extra-tropical cold and dry air.

120 LST biases and their influence on the ISM have been relatively poorly studied
121 compared to TT biases and errors due to orography (Kumar et al. 2014). The
122 pioneering work of Charney et al. (1977) addressed the sensitivity of summer
123 monsoon regions to land surface heating by modifying the land albedo in their model.
124 They proposed a mechanism linking the land albedo with the monsoon strength. This

125 mechanism was subsequently summarized by Meehl (1994): an increase in land
126 albedo creates a decrease of the solar flux absorption leading to a colder land surface
127 and thus to a decreased land-sea thermal gradient. This decrease of the land-sea
128 contrast weakens the monsoon flow and the associated precipitation. This mechanism
129 has been further explored and confirmed by Meehl (1994) with various atmospheric
130 General Circulation Models (GCMs). Zhaohui and Qingcun (1997) showed some
131 improvements in the East Asian monsoon and associated rainfall in their GCM when
132 using an observed climatological albedo. Using an idealized configuration, Chou
133 (2003) also showed that changing the land surface albedo can strengthen or weaken
134 the meridional TT gradient and, consequently, the ISM migration and intensity.
135 Furthermore, this method has been successfully used by Kelly and Mapes (2013) to
136 control the strength of the ISM in dedicated sensitivity experiments. Finally, Flaounas
137 et al. (2012) showed that lowering land albedo modifies the Inter-Tropical
138 Convergence Zone (ITCZ) position during the West African monsoon. In a nutshell,
139 the specific effect of land surface heating processes, versus orographic effects, on the
140 monsoon is still an open problem (Kelly and Mapes 2010, Rajagopalan and Molnar
141 2013, Wu et al. 2012, Ma et al. 2014).

142 The present work aims at revisiting the relationship between land surface
143 albedo, surface heating and the ISM biases in a state-of-the-art Coupled Tropical
144 Channel Model (CTCM). Due to the significant ocean-atmosphere feedbacks involved
145 in ISM variability and seasonal cycle (Wang et al. 2005), our work is based on a
146 coupled model, rather than a forced atmospheric model as done in many previous
147 studies. We demonstrate that constraining land surface heating by using an observed
148 albedo climatology leads to significant improvements in ISM simulation in our
149 CTCM, especially in terms of the ISM rainfall onset and climatology. Moreover, we
150 illustrate that our results are valid in both coupled and forced frameworks with several
151 dedicated experiments, highlighting the robustness of our findings. The paper is
152 organized as follows. Section 2 provides a detailed description of the CTCM, the
153 experimental setup and observed datasets used in our analysis. Section 3 describes the
154 ISM mean characteristics and biases simulated with the CTCM. The sensitivity of the
155 ISM to the land surface albedo and horizontal resolution is further analyzed in Section
156 4 with the help of sensitivity experiments. Finally, Section 5 provides a summary of
157 our findings.

158

159 **2. Model description and experimental setup**

160

161 2.1. Model description

162 The CTCM is composed of the WRF-ARW v3.3.1 atmospheric model
163 (Skamarock and Klemp 2008) and the NEMO v3.4 ocean model (Madec 2008)
164 coupled through the OASISv3-MCT coupler (Valcke et al. 2013).

165 The oceanic and atmospheric components share an identical horizontal grid
166 discretization (Arakawa-C grid), projection (Mercator) and resolution (0.75° or 0.25°).
167 The standard horizontal resolution used here is 0.75° , but some of our sensitivity
168 experiments (described later in Section 2.b) use a 0.25° horizontal resolution in order
169 to assess the robustness of our results with respect to the model resolution. The CTCM
170 domain extends from 45°S to 45°N , covering about 70% of the earth surface.
171 Consequently, the extratropics (poleward of 45° of latitude) can exert an influence on
172 the CTCM through the lateral atmospheric and oceanic forcings, but the model is also
173 able to generate its own tropical internal variability at all the timescales, as seen from
174 the simulated El Niño events, which timing does not match the observed El Niño
175 events during the 1989-2009 period (not shown). This original model configuration
176 presents several advantages compared to the classical GCMs and RCMs approaches.
177 Compared to RCMs, the tropical-channel configuration is not subject to issues related
178 to domain size, which can influence the realism of the model solution (Leduc and
179 Laprise 2009, Dash et al. 2015). Because of the absence of meridional boundaries, the
180 model is also able to simulate zonally-propagating atmospheric and oceanic waves in
181 a coherent way, as well as zonal teleconnections and remote tropical forcings.
182 Consequently, this model avoids an important caveat observed with RCMs, which can
183 generate spurious circulations and precipitations along their meridional boundaries in
184 zonal flows (Hagos et al. 2013). Such tropical-channel configuration has
185 demonstrated its usefulness to study tropical waves activity, such as the Madden-
186 Julian Oscillation (Ray et al. 2011, Ulate et al. 2015) and inertia-gravity waves (Evan
187 et al. 2012). Compared to GCMs, the model does not include the extratropics, which
188 limits the inclusion of additional errors and reduces the simulation computational cost.

189 The ocean vertical grid has 75 z-levels, with 25 levels above 100 m and a
190 resolution ranging from 1 m at the surface to 200 m at the bottom. Partial filling of the
191 deepest cells is allowed. The atmospheric grid has 60 eta-levels with a top of the
192 atmosphere located at 50 hPa. The WRF default vertical resolution has been
193 multiplied by three below 800 hPa. Thus, the first 33 levels are located below 500 hPa
194 with a vertical resolution of 2 hPa near the surface. The vertical resolution then
195 decreases to ~50 hPa around 800 hPa and increases again when approaching the top
196 of the model with ~6 hPa for the top level.

197 The WRF model can be configured with an important choice of physical
198 schemes. In this study, the model physical setup is the same as in Samson et al.
199 (2014), who showed that a NEMO-OASIS-WRF (NOW) regional coupled model is
200 able to realistically simulate the tropical IO climate, including the ISM main
201 characteristics. This physical package is listed here: the longwave Rapid Radiative
202 Transfer Model (RRTM; Mlawer et al. 1997), the “Goddard” Short Wave (SW)
203 radiation scheme (Chou and Suarez 1999), the “WSM6” microphysics scheme (Hong
204 and Lim 2006), the Betts-Miller-Janjic (BMJ) convection scheme (Betts and Miller
205 1986; Janjic 1994), Yonsei University (YSU) planetary boundary layer scheme (Hong
206 et al. 2006), the unified NOAH Land Surface Model (LSM) with the surface layer
207 scheme from MM5 (Chen and Dudhia 2001). Mukhopadhyay et al. (2010) and
208 Samson et al. (2014) shown that the BMJ convection scheme produces a reasonable
209 ISM climatology in both forced and coupled WRF configurations, respectively.
210 Supplementary sensitivity tests have been performed with different sets of physical
211 parameterizations, but with no clear improvement when compared to the selected set.
212 A brief description of these sensitivity tests is given below and a more complete
213 analysis of the sensitivity of the simulated tropical mean state to various model
214 parameters can be found in Crétat et al. (2016). Thus, this study uses a well-tested
215 suite of parameterization schemes for the WRF model.

216 The oceanic component is based on NEMO (Nucleus for European Modeling
217 of the Ocean numerical framework) version 3.4 (Madec 2008). The set of physical
218 parameters employed here is similar to the set used for the default global
219 configuration at 1°-resolution (Voltaire et al. 2013). The lateral diffusion scheme for
220 tracers is an iso-neutral Laplacian with a constant coefficient of 1000 m²/s. Tracer
221 advection is treated with a total variance dissipation scheme (Lévy et al. 2001) with

222 an additional term coming from the eddy-induced velocity parameterization (Gent and
223 McWilliams 1990) with a space and time variable coefficient (Tréguier et al. 1997).
224 The lateral diffusion of momentum is a horizontal Laplacian with an eddy viscosity of
225 $10\,000\text{ m}^2/\text{s}$, which is reduced to $1000\text{ m}^2/\text{s}$ in the 2.5°S - 2.5°N equatorial band, out of
226 the western boundaries regions. The vertical mixing is parameterized using an
227 improved version of turbulent kinetic energy closure scheme (Blanke and Delecluse
228 1993) with a Langmuir cell (Axell 2002) and a surface wave breaking
229 parameterization (Mellor and Blumberg 2004).

230 The OASIS coupler exchanges the surface fields between the models every 2
231 hours without any spatial interpolation as the models are using the same horizontal
232 grid (see Samson et al. 2014 for details). Such a high coupling frequency is crucial in
233 the tropics to correctly represent the solar diurnal cycle effect on the ocean. It has
234 been shown that high-frequency coupling is instrumental in representing
235 realistically the monsoon dynamics (Terray et al. 2012) as well as air-sea scale
236 interactions from small scales to large scales and up to ENSO variability (Masson
237 et al. 2012). There is no restoring of any kind in atmosphere and ocean. Initial state
238 and boundary conditions come from the ERA-Interim reanalysis (Dee et al. 2011) for
239 the atmospheric component and from the Drakkar 0.25° -resolution global ocean
240 model (Barnier et al. 2007) for the oceanic component over the 1989-2009 period.

241

242 2.2. Experimental Setup

243 The reference simulation (ALB1 hereafter) described in the previous
244 paragraph is compared with three different sets of simulations. Table 2 summarizes all
245 the model simulations used in this study.

246 A first set of four 10-years atmospheric simulations forced with observed SST
247 (sensitivity set hereafter) is used to assess the sensitivity of the simulated ISM biases
248 to the SST errors (due to the coupling with the ocean model), to the model resolution
249 and to the atmospheric SW and convective schemes. In this sensitivity set of forced
250 simulations, the FORC simulation is similar to the ALB1 simulation, except that the
251 atmospheric model is forced with observed SSTs from version 2 of the 0.25° daily
252 optimum interpolation SST analysis from the NOAA (Reynolds et al. 2007). The
253 HIRES simulation differs from FORC by its horizontal resolution of 0.25° instead of

254 0.75°. The CONV simulation is the same as FORC, but with a different convection
255 scheme (Kain-Fritsch instead of the BMJ scheme; Kain 2004). Finally, the RAD
256 simulation is similar to FORC, but with a different short-wave radiation scheme
257 (Dudhia scheme instead of the Goddard scheme; Dudhia 1989). These forced
258 atmospheric simulations will be analyzed in Section 3.

259 The second set of simulations (albedo set hereafter) is composed of two
260 additional 20-years fully coupled CTCM simulations identical to the ALB1
261 simulation, except for the land surface albedo used in the CTCM. In this second set of
262 coupled simulations, two different background land surface albedo fields are
263 employed as explained below. These simulations will be analyzed in Section 4.

264 The NOAA LSM version available with WRF uses a simplified and direct
265 method to compute the SW fluxes at the surface. This LSM only considers the
266 broadband SW wavelength, which means that no distinction is made between visible
267 and infrared wavelength albedos. The albedo dependence to the solar zenith angle is
268 also neglected. Moreover, no distinction can be made between the diffuse and the
269 direct components of solar radiation, as they are not available in the WRF version
270 used in this study. In this simplified context, albedo associated with the diffuse SW
271 component (i.e. white-sky) is neglected and the total incoming solar flux is considered
272 as purely direct. Consequently, the NOAA LSM uses a snow-free direct (e.g. black-
273 sky at local noon) background SW broadband albedo climatology to compute the SW
274 fluxes at the surface during the simulation. Two methods are available in WRF to
275 prescribe this land surface albedo monthly climatology.

276 In the CTCM reference simulation (ALB1), albedo extreme values (annual
277 minimum and maximum) are associated with a dominant land-use category. A
278 weighted average is then computed between the two albedo extreme values depending
279 on the corresponding climatological monthly green fraction. Consequently, the albedo
280 is equal to its minimum (maximum) annual value when the green fraction is
281 maximum (minimum). The land-use dataset used by the NOAA LSM is the
282 MODERate resolution Imaging Spectroradiometer (MODIS) land-cover classification
283 of the International Geosphere-Biosphere Program (IGBP; Friedl et al. 2002) and
284 modified for the NOAA LSM (lakes detection and 3 new categories (18-19-20) have
285 been added). The dataset has been updated with MODIS data up to March 2011 (see
286 [WRF FAQ link](#)) and the annual climatology of this dataset is displayed in Fig. 1a. The

287 vegetation fraction used by the NOAH LSM is the NESDIS/NOAA 0.144° monthly
288 annual cycle of the vegetation greenness fraction dataset (Gutman and Ignatov 1998).
289 This dataset is a 5-year (1985-1990) climatology of the Advanced Very High-
290 Resolution Radiometer (AVHRR) vegetation index.

291 The second method available with WRF consists of directly prescribing a
292 snow-free black-sky SW broadband albedo climatology. The albedo dataset provided
293 with WRF for the NOAH LSM is the NESDIS/NOAA 0.144° monthly 5-year
294 climatology surface albedo derived from the AVHRR satellite (henceforth AVHRR
295 product; Csiszar and Gutman 1999). The error analysis performed in Csiszar and
296 Gutman (1999) suggests that the AVHRR surface albedo is retrieved with 10 to 15%
297 relative accuracy. The simulation using this prescribed albedo is referred as ALB2
298 hereafter.

299 In order to test the robustness of our results, a third CTCM simulation using an
300 up-to-date snow-free black-sky SW broadband albedo climatology estimated from
301 MODIS data (henceforth MODIS product; Schaaf et al. 2011) has also been
302 performed (ALB3 in Table 2). This MODIS albedo dataset is described in the next
303 paragraph and its annual climatology is presented in Fig. 1b. This simulation gave
304 results very similar to ALB2 despite the fact that the MODIS albedo is slightly higher
305 than the NESDIS/NOAA albedo used in ALB2 (Fig. 1d). This confirms the
306 robustness of the results obtained with ALB2. Results from ALB3 are consequently
307 not shown in this study for conciseness, but demonstrate that our results are
308 independent of the observed albedo product used in the simulations.

309 Finally, two more coupled simulations similar to ALB1 and ALB2
310 respectively have been performed but with a horizontal resolution of 0.25° instead of
311 0.75° in order to demonstrate the robustness of our results with respect to the
312 resolution used in the CTCM ("High-Resolution Set" in Table 2). These two high-
313 resolution coupled simulations (ALB1HR and ALB2HR, respectively) are also
314 analyzed in order to determine the cumulative (positive) effects of both an increased
315 spatial resolution and a change of the albedo on the simulated ISM characteristics by
316 the CTCM. The results from ALB1HR and ALB2HR are discussed in Section 4.

317

318 2.3. Observational Datasets

319 Several datasets are used in this study. First, the two NOAH snow-free albedo
320 fields (ALB1 and ALB2) are compared with the MODIS snow-free gap-filled black-
321 sky SW broadband albedo product MCD43GF-v5 (Schaaf et al. 2011). This product is
322 generated by merging data from the Terra and Aqua platforms produced every 8 days,
323 with 16-days acquisition and available on a 0.05° global grid. It is important to note
324 that some regions are systematically masked by clouds during monsoon months,
325 which makes direct albedo measurements difficult, or even impossible (Rechid et al.
326 2009). A temporal interpolation is applied to fill these missing values. The MODIS
327 snow-free climatology is computed over the 2003-2013 period and is presented in Fig.
328 1b. The accuracy of this product is about 2% when compared to ground observations
329 (Jin et al. 2003, Wang et al. 2004). The differences between NOAH LSM and MODIS
330 snow-free albedo annual climatologies, as well as between the AVHRR and MODIS,
331 are presented in Figs. 1c-d, respectively. We also use the MODIS MCD43C3-v5
332 product to validate output model surface albedo, which also includes the snow cover
333 effect. This dataset is identical to the MCD43GF-v5 product, but surface data
334 including snow covered areas are included in the processing.

335 Model precipitation is compared with the monthly 0.25° Tropical Rainfall
336 Measuring Mission (TRMM) 3B43-v7 rainfall product (Huffman et al. 2010)
337 averaged over the 1998-2014 period. This dataset combines the 3-hourly merged
338 high-quality/infrared estimates with the monthly-accumulated Global Precipitation
339 Climatology Centre (GPCC) rain gauge analysis. Monthly climatological fields such
340 as surface temperature, wind and Sea-Level Pressure (SLP) are derived from the 0.75°
341 ERA-Interim reanalysis (Dee et al. 2001) averaged over the 1989-2009 period.
342 Finally, surface radiative heat budget is computed from the monthly 1° CERES-
343 EBAF v2.8 product (Kato et al. 2013) over the 2001-2012 period.

344

345 **3. ISM biases in ALB1 simulation and the sensitivity set**

346

347 Modeling systems must be evaluated for their basic performance in terms of
348 their capability to correctly reproduce the main features of the climate system. More
349 specifically, the simulation of a realistic boreal summer precipitation climatology is a
350 primary requirement that a model should possess for monsoon studies, but it remains

351 a difficult task for current state of the art CGCMs (Sperber et al. 2013, Prodhomme et
352 al. 2014, Annamalai et al. 2015). As a first step, we thus examine in this section the
353 systematic errors that characterize the climatologies of rainfall, low-level winds,
354 surface temperature and SLP simulated in the reference run (ALB1) of the CTCM
355 during boreal summer (JJAS). The possible origins of these systematic errors are then
356 investigated with the help of several dedicated experiments performed with the forced
357 atmospheric component of the CTCM (see Section 2.b for details about these
358 experiments). Annual cycles of simulated ISM rainfall indices are also discussed.

359

360 3.1. ISM description in observations and ALB1 control simulation

361 The climate of South Asia is dominated by the monsoon. During boreal
362 summer, a strong inter-hemispheric SLP gradient is observed over the Indian Ocean
363 area, with a deep low centered over Pakistan and northwestern India (Fig. 2a). The
364 close correspondence between SLP and surface temperature over northwest India
365 suggests that the intense solar heating over the northern hemisphere during spring and
366 summer favors the development of this low. This explains why this deep low is often
367 referred as a “heat” low in the literature (Flohn 1968). However, orography and
368 diabatic heating over the Bay of Bengal (BoB) also exert a dominant control on the
369 deepening of this low during the rainy season through remotely forced subsidence
370 over Iran–Turkmenistan–Afghanistan and the Rodwell-Hoskins’ monsoon–desert
371 mechanism (Yanai et al. 1992, Rodwell and Hoskins 1996, Bollasina and Nigam
372 2011). This “heat” low is connected to a tilted band of low SLP extending from the
373 northern BoB to northwest India over the Indo-Gangetic plain, which is usually
374 referred to as the monsoon trough. The monsoon trough is the signature of transient
375 Low-Pressure Systems (LPSs) propagating inland from the BoB during the summer
376 monsoon (Krishnamurthy and Ajayamohan 2011).

377 As expected, the large inter-hemispheric SLP gradient over the Indian domain
378 generates vigorous cross-equatorial southerly monsoonal winds over the western
379 IO/east African highlands (Fig. 2d). Due to the Coriolis effect, this monsoon low-
380 level flow gradually becomes westerly over the Arabian Sea (AS), resulting in a
381 strong moisture flux toward the Asian landmass and bringing abundant rainfall over
382 South Asia during boreal summer.

383 Precipitation increases sharply from April to June, which corresponds to the
384 monsoon onset and the sudden “jump” of the ITCZ from its oceanic to continental
385 position during boreal summer (Fig. 3; see Wang 2006). The orography provides
386 anchor points where monsoon rainfall maxima are located, especially along the
387 Western Ghats, the Burmese coast and the Philippines (Fig. 2d). Abundant rainfall is
388 also observed over the Gangetic plain and the foothills of Himalaya associated with
389 the LPSs propagating from the BoB into northwest India during the summer monsoon
390 (Krishnamurthy and Ajayamohan 2011). During this season, SST maximum is
391 observed in the eastern equatorial IO, while the western AS is characterized by colder
392 SSTs as a result of coastal upwelling and strong evaporation in response to the strong
393 southwesterly alongshore winds (Figs. 2a and d) (de Boyer Montégut et al. 2007).
394 This low-level jet (the so-called Findlater jet) and the associated cold SSTs prevent
395 atmospheric deep convection to occur in the western part of the basin (Gadgil et al.
396 1984).

397 The spatial pattern of the JJAS precipitation bias in ALB1 (Fig. 2f) exhibits
398 many similarities with the systematic errors commonly observed in CMIP5 models
399 (see Sperber et al. 2013; Sooraj et al. 2015). In particular, a dry bias is present over
400 the Indian subcontinent with two maxima along the Ghats and over the foothills of the
401 Himalaya. A relationship exists between precipitation biases and 850 hPa wind biases
402 in regions where orographic forcing is important. A rainfall dry (wet) bias is usually
403 associated with an underestimation (overestimation) of the low-level wind in these
404 regions. Deficient rainfall is also simulated over the monsoon core region (65° - 100° E
405 / 5° - 30° N) suggesting that the whole ISM is too weak in ALB1. The simulated ISM
406 rainfall annual cycle over the continent is very poor, to say the best, with a monthly
407 maximum hardly reaching 6 mm/day in August (Fig. 3a). Moreover, ISM onset is
408 delayed by almost 2 months in ALB1 (Figs. 3a-b). Consequently, the dry bias
409 observed over India during boreal summer (Figs. 2e-f) is due to underestimated
410 precipitation intensity, but also to a significant underestimation of the duration of the
411 rainy season. Consistently, the Findlater jet is significantly underestimated, too much
412 zonal, and its northward extension is limited to 15° N in the CTCM instead of 20 - 25° N
413 in ERA-Interim (Figs. 2d-f).

414 Excessive rainfall is present over the south-eastern AS, reflecting again this
415 limited northward propagation of the ITCZ during boreal summer in ALB1 (Fig. 2f).

416 This wet bias is usually associated with warmer-than-observed local SST as a
417 consequence of a too weak monsoon flow, reduced latent heat loss and under-
418 representation of the upwelling along the Somali and Omani coasts in our CTCM and
419 in CMIP5 models (Fig. 2c; Prodhomme et al. 2014, Li et al. 2015). East Asia and
420 South China Sea also exhibit excessive rainfall associated with overestimated
421 westerly low-level winds over eastern equatorial IO and South China Sea in the
422 CTCM (Fig. 2f).

423 SST biases are moderate with a warm bias slightly exceeding 2°C in the
424 western tropical IO (Fig. 2c), as discussed above. The largest surface temperature
425 biases are found over land with a maximum warm bias slightly exceeding 10°C over
426 most of central and northern India. A warm bias of $\sim 3^{\circ}\text{C}$ is also observed over South-
427 East Asia and over the Maritime Continent despite the excessive rainfall simulated in
428 these regions. On the contrary, cold surface temperature biases are found (i) over the
429 western part of the Tibetan plateau, suggesting an indirect effect of overestimated
430 snow and precipitation over this elevated area during boreal winter (not shown) and
431 (ii) in the desertic region extending from Pakistan to Afghanistan, Iran and over the
432 Arabic Peninsula. This region will be referred as the “Middle-East” (ME) region
433 hereafter for simplicity.

434 Simulated SLP also exhibits significant biases with lower-than-observed SLP
435 over most of the domain, except in the ME region where a positive SLP bias of
436 several hPa is found. The low-pressure bias is maximum over the core monsoon
437 region and along the Himalayan foothills. This SLP bias is also commonly observed
438 in CMIP5 models as shown by Sooraj et al. (2015; see their Fig. 3d). As a
439 consequence, the SLP minimum located over the ME region in ERA-Interim is shifted
440 eastward over the monsoon trough region in ALB1 (Figs. 2a-b). The large dry bias
441 over the monsoon trough region in ALB1 suggests that LPSs and clouds are less than
442 observed or even absent in this region during boreal summer and, hence, that the low-
443 pressure bias is not related to excessive LPSs, but rather to the strong warm surface
444 temperature bias. In turn, this excessive land-surface heating may result from reduced
445 rainfall and clouds associated with the absence of these LPSs over this region. Indeed,
446 the ITCZ is locked over the ocean in ALB1, southward of its observed position (Figs.
447 2e-f). Alternatively, the warm bias may also be related to deficient land processes in
448 the CTCM, as we will see in the next section.

449

450 3.2. ISM biases in sensitivity experiments

451 Various mechanisms have been proposed to explain the dry bias and the
452 delayed monsoon onset over the Indian landmass simulated by current CGCMs (see
453 Introduction). To explore these various potential sources of errors in our modeling
454 framework, the “sensitivity” set of simulations is analyzed in this section (see Section
455 2 and Table 2 for further details). All configurations, excepted FORC and HIRES,
456 underestimate the total amount of rainfall during the monsoon season. In most cases,
457 this is related to dry conditions over land (Fig. 3a), but also to an ISM onset delayed
458 by almost 2 months over land (except HIRES) as in ALB1. Figure 4 further illustrates
459 model sensitivity to changes in the ocean-atmosphere coupling, the physics and the
460 resolution used:

461 • Ocean-atmosphere coupling and SST (FORC)

462 A first source of model errors is the coupling with an ocean model and the
463 resulting SST errors. Compared to ALB1, the onset delay is attenuated and the total
464 precipitation is increased in FORC (Figs. 3 and 4a, first line). Nonetheless, the
465 monsoon peak time and withdrawal time remain delayed, especially when considering
466 land areas (Fig. 3b). There is also a strong spatial compensation of rainfall error
467 patterns between north India and the BoB in FORC, as in ALB1, and the warm (cold)
468 bias is still present in northern and east India (Pakistan), but is attenuated over
469 southern India (Fig. 4a).

470 • Resolution (HIRES)

471 Higher horizontal resolution induces a better simulation of orographic
472 precipitation. The improved Himalayan orography also prevents mixing between cold
473 and dry air from mid-latitudes with warm and moist air from the tropics, allowing a
474 stronger TT gradient and hence a more intense and realistic ISM (Boos et al. 2010;
475 2013). HIRES significantly improves the precipitation seasonal cycle with a
476 maximum reached in July as in observations (Fig. 3). However, the dry bias persists
477 over India, especially along the Western Ghats and in the northern and eastern BoB,
478 although it is well attenuated compared to FORC (Fig. 4b). The same holds for the
479 warm temperature bias, which is still present, but attenuated over northern India in
480 HIRES. Nevertheless, it is noteworthy that the spatial patterns of rainfall, low-level

481 wind and temperature errors remain basically the same as in the FORC experiment
482 (Figs. 4a-b).

483 • Physics (RAD and CONV)

484 CONV and RAD experiments suffer from the same deficiencies, with a dry
485 bias - even more pronounced - over India and a more southward and oceanic position
486 of the ITCZ (Figs. 4c and d). Pronounced wet biases are also found over the Maritime
487 Continent, the eastern equatorial IO and China in CONV, and along a line extending
488 from the equatorial IO to the South China Sea in RAD. The warm bias over India is
489 also present in these two simulations, even if it is well attenuated in RAD. On the
490 other hand, the use of the Dudhia (1989) radiation scheme leads to an enhancement of
491 the ME cold bias and to an erroneous zonal surface temperature gradient between this
492 region and South Asia, suggesting that these surface temperature variations do affect
493 the latitudinal position of the ITCZ during boreal summer (Fig. 4d).

494 In a nutshell, the ISM and associated precipitation patterns are very sensitive
495 to the model configuration settings. Our SST-forced and high-resolution simulations
496 show significant improvements in terms of precipitation amount and seasonal cycle,
497 even if dry and warm biases persist over North India. On the contrary, our convective
498 and radiative sensitivity tests show a clear deterioration of the simulated ISM with a
499 further increased dry bias over India and an even more southward and oceanic
500 position of the ITCZ compared to the other simulations (e.g. ALB1, FORC and
501 HIRES). Finally, the improvements or degradations in the simulated rainfall concern
502 mainly the amplitude of the rainfall biases over land and ocean, not the spatial pattern
503 of these systematic errors: in all these sensitivity experiments, as in ALB1, we
504 observe excess rain over the ocean compared to observations, especially in the
505 southern part of the BoB (FORC, HIRES and RAD) and the southeastern AS (FORC
506 and RAD), and dry conditions over the land, especially along the Western Ghats and
507 over the monsoon core region.

508

509 3.3. Surface temperature and SLP biases origins

510 All sensitivity simulations systematically present a high-pressure bias over the
511 ME region and a low-pressure bias over India and southeast Asia (Fig. 4, right
512 column). More intriguingly, all the configurations, including HIRES, exhibit similar

513 spatial pattern of skin temperature errors during the monsoon season with warmer-
514 than-observed surface temperature over the core monsoon region and the foothills of
515 the Himalaya, and cooler-than-observed surface temperature over the ME region. This
516 seems to induce significant errors in the SLP field due to erroneous surface heating
517 forcing over the land. It is noteworthy, that these surface temperature errors are also
518 present in HIREs, despite reduction of the rainfall dry bias over the monsoon trough
519 region in this simulation. This suggests that at least part of these temperature errors
520 are not due to reduced cloudiness and evaporation over the Indo-Gangetic plain, but to
521 other reasons related to land processes.

522 Figure 5 shows the observed surface temperature and SLP climatologies
523 during the spring season (March-April-May) preceding the ISM onset and the
524 corresponding ALB1 biases. In observations (Fig. 5a), the surface temperature and
525 SLP patterns are strikingly different from the JJAS period, especially in the ME
526 region where the surface heating remains small compared to what is observed during
527 the monsoon season. On the contrary, India and South Asia are much warmer during
528 spring than during JJAS since the incident solar radiation is not balanced by clouds
529 and precipitation cooling as during the monsoon. Consequently, the land temperature
530 warming is very homogenous during spring. This is also true for the tropical IO, as no
531 significant SST gradient is present during this season. The model captures quite well
532 this homogenous spatial pattern in terms of surface temperature (pattern correlation =
533 0.95) and SLP (Fig. 5b). But surface temperature biases, previously described during
534 JJAS, are already present during the pre-monsoon hot and dry season with a strong
535 warm (cold) bias over India and South-East Asia (Tibetan plateau) and a relatively
536 smaller cold bias in the ME. This suggests that surface temperature biases observed
537 during the monsoon already exist before the monsoon onset and are thus not solely
538 related to the dry bias and improper ITCZ position simulated during JJAS in ALB1.
539 This is further explored in the next sections with more detailed diagnostics and the
540 “albedo” set of coupled experiments.

541

542 **4. Effect of changing the land surface albedo on the ISM biases**

543

544 As discussed in the previous section, the warm bias over India cannot be
545 entirely related to the dry bias during ISM and this dry bias is not due to SST errors
546 since it persists in the forced-atmospheric experiments. The same stands for the cold
547 bias over the ME region, which already exists before the monsoon season (Fig. 5c).
548 Consequently, it appears that the model biases are at least partly related to the land
549 surface properties. In this section, we focus on the effect of changing the land surface
550 albedo on the ISM biases by comparing ALB1 and ALB2 coupled simulations.

551

552 4.1. ALB1-ALB2 albedo comparison

553 As seen in Figure 1c, ALB1 snow-free albedo annual climatology is affected
554 by significant biases when compared to MODIS. ALB1 albedo is globally higher than
555 MODIS, even if some significant underestimations are seen in the North African
556 desert and some other local areas. The positive errors can reach ~20% in some regions
557 such as the Andes mountains, the Tibetan plateau and the Iran–Turkmenistan–
558 Afghanistan region. These errors are mainly due to the fact that the number of land-
559 use categories is too limited to correctly represent the diversity of land surfaces at a
560 regional scale. For example, the same albedo value (0.38) is used in all the desertic
561 regions, while their albedo can vary significantly according to their surface
562 composition (e.g. black rocks vs white sand). Consequently, this simplified approach
563 can lead to important differences when compared to in situ or satellite-based observed
564 albedo, especially in arid regions (Fig. 1c). On the contrary, ALB2 snow-free albedo
565 climatology, derived from AVHRR albedo product, is relatively close to the MODIS
566 snow-free product with an overall underestimation of about 5%, except in some
567 regions such as India and South-East Asia where the albedo is slightly overestimated
568 (Fig. 1d).

569 Figures 6a-b show the JJAS total albedo (e.g. including snow effect)
570 differences between ALB2 and ALB1 simulations and ALB2 biases compared to the
571 corresponding MODIS product (also including snow effects). ALB2 albedo is almost
572 everywhere lower than ALB1 with maximum differences located in the ME region,
573 on the western Tibetan plateau and along the Himalaya mountains. The differences in
574 high-elevated areas are mainly due to differences in the snow cover, with less snow in
575 ALB2 compared to ALB1 (not shown). But despite a smaller snow cover, ALB2

576 albedo is still overestimated in these mountainous regions when compared to MODIS
577 because snow albedo is much higher than bare soil albedo (Fig. 6b). Various reasons
578 can explain this bias: too much snow during boreal winter, unrealistic snow melting
579 (e.g. too slow) due to improper LSM physics or snow albedo parameterization, which
580 prevent the spring snow melt. Except in these snowy and elevated regions, ALB2
581 biases do not exceed 5% in the considered domain. Maximum albedo differences
582 reaching locally ~30% between ALB1 and ALB2 are located in the ME region (Fig.
583 6a). This area is also affected by a cold bias in the various forced simulations
584 analyzed in Section 3. To investigate the relationship between this cold bias and the
585 land surface albedo and to quantify the sensitivity of the LST to albedo, a surface
586 radiation budget analysis is performed over a box covering the ME region (40°-70°E /
587 15°-37°N).

588

589 4.2. Albedo radiative effect over the Middle-East region

590 The various terms of the land surface radiative heat budget in the simulations
591 are compared and validated against the CERES-EBAF dataset in Figure 7. In both
592 simulations, the land surface receives too much downward SW flux ($\sim 40 \text{ W.m}^{-2}$)
593 when compared to CERES-EBAF observations (Fig. 7a). This bias is related to the
594 "Goddard" SW scheme, which tends to overestimate the SW downward flux at the
595 surface (Crétat et al. 2016). However, the fraction of SW downward flux reflected by
596 the surface varies according to the background albedo used in the simulations (Fig.
597 7b). Consequently, the lower albedo in ALB2 simulation efficiently decreases the
598 upward SW flux (by about 40 W.m^{-2}) and the land surface receives a higher net SW
599 flux (30 to 35 W.m^{-2}) compared to ALB1 and CERES-EBAF. This additional SW
600 flux induces a higher LST in ALB2 (Fig. 6c). In turn, this higher LST induces an
601 increased upward longwave (LW) flux emitted by the surface ($\sim 20 \text{ W.m}^{-2}$), which
602 results in a higher net LW heat loss ($\sim 10 \text{ W.m}^{-2}$) in ALB2 than in ALB1 (Fig. 7b).
603 Finally, the net radiative flux is underestimated in ALB1 by $\sim 10 \text{ W.m}^{-2}$ while it is
604 overestimated by $\sim 15 \text{ W.m}^{-2}$ in ALB2 compared to CERES-EBAF. It corresponds to a
605 difference between ALB1 and ALB2 of $\sim 25 \text{ W.m}^{-2}$. These differences are
606 significantly greater than the CERES land surface LW and SW root-mean-square
607 errors given by Kato et al. (2013), which both amount to about 8 W.m^{-2} , respectively.

608

609 4.3. Albedo effect on the ISM

610 As we said, this modification of the surface radiative budget in ALB2
611 compared to ALB1 induces a strong warming over the ME region ranging from 2 to
612 5°C (Fig. 6c). A robust ($r^2=0.6$) and negative (-0.2°C when albedo increases by 1%)
613 relation is found when we compare the Middle-East JJAS mean climatological surface
614 temperature difference between ALB2 and ALB1 with the albedo difference between
615 ALB2 and ALB1. As a consequence, the cold bias observed in ALB1 turns into a
616 warm bias in ALB2 (Fig. 6d). A warming of the Tibetan plateau locally reaching
617 10°C is also simulated in ALB2 compared to ALB1 (Fig. 6c). It can be explained by
618 the lower snow-free albedo in ALB2 compared to ALB1, as for the ME region.
619 Consequently, the cold bias is also reduced in this area (compare Figs. 2c and 6d).

620 On the contrary, a surface cooling is observed in ALB2 over the eastern part
621 of the domain (Fig. 6c). The colder area extends from southern India through northern
622 China. Consequently, the significant warm bias present over southern India and
623 South-East Asia in ALB1 is slightly reduced when compared to the ERA-Interim LST
624 (Fig. 6d). However, this surface cooling is not directly related to the local albedo
625 because it is slightly lower in ALB2 than in ALB1, which would contribute to warm
626 the surface in these regions. On the other hand, no significant LST change is observed
627 over the foothills of Himalaya. Finally, the SST is not significantly affected by the
628 albedo change, except in the upwelling region along the Omani coast, which is about
629 ~1.5°C cooler in ALB2 than in ALB1 (Fig. 6c).

630 The land surface warming difference between ALB2 and ALB1 is associated
631 with important SLP changes. Globally, SLP is lower in ALB2 compared to ALB1
632 (Fig. 6c). The decrease is relatively weak over ocean (~1-2 hPa), but it is superior to 4
633 hPa throughout the ME region (with a maximum of 6 hPa at 30°N, 55°E). Over this
634 area, the similarity between SLP and surface temperature differences (Fig. 6c)
635 suggests a direct relation between surface warming and SLP decrease through air
636 density adjustment following the ideal gas law for dry air. This is confirmed by a
637 significant ($r^2=0.6$) and negative (-0.5 hPa/°C) linear regression between the ME JJAS
638 mean climatological SLP (ALB2-ALB1) difference and the surface temperature
639 (ALB2-ALB1) difference (not shown). Over the rest of the domain, such

640 correspondence is less obvious. Over the ME region, SLP bias compared to ERA-
641 Interim turns from positive with ALB1 to negative with ALB2 (Fig. 6d), in agreement
642 with the corresponding surface temperature biases and our radiative budget analysis.
643 Overall, the inter-hemispheric SLP gradient and SLP land-sea contrast, which drive
644 the monsoon, are both enhanced in ALB2 compared to ALB1.

645 In agreement with this improved SLP pattern over the ME region,
646 precipitation over the Indian subcontinent is significantly increased between ALB2
647 and ALB1 with maxima located along the Western Ghats and the Himalayan foothills
648 (Fig. 6e). The dry bias is also well attenuated over India southward of 25°N (Fig. 6f).
649 On the contrary, precipitation is decreased in the equatorial Indian Ocean, over South-
650 East Asia and, especially, over South China Sea, even though a wet bias persists in
651 this region. This is the signature of a more northward and continental position of the
652 ITCZ in ALB2 than in ALB1. As a consequence, rainfall pattern and intensity are
653 globally improved in ALB2 (even if significant biases persist). The spatial matching
654 between the increased precipitation over land (Fig. 6e) and the land surface cooling
655 (Fig. 6c) suggests that the warm bias reduction is a consequence of the enhanced
656 rainfall in those regions.

657 The low-level wind pattern is also clearly improved in ALB2 compared to
658 ALB1 with both a strengthening and a more poleward extension of the Findlater jet
659 and a zonal wind decrease in the eastern part of the BoB and South China Sea. More
660 moisture is advected from the BoB into Bangladesh and the plains of northern India in
661 ALB2, and the rainfall is enhanced over these areas in ALB2 compared to ALB1 (Fig.
662 6e). These patterns of differences are again in agreement with a more northward
663 propagation of the monsoon in ALB2 compared to ALB1.

664

665 4.4. LST-SLP-wind relationship

666 In order to understand the differences between ALB2 and ALB1, we can
667 assume that the 850hPa wind is approximately in geostrophic equilibrium with the
668 SLP outside the equatorial or elevated regions and above the boundary layer where
669 frictional effects are important. This relationship between low-level wind and SLP is
670 well illustrated in Figures 2a and d, in which the Findlater jet closely follows SLP
671 contours and its speed is maximum where the SLP gradient formed between the

672 western equatorial IO and the ME region is also maximum. A similar relationship can
673 be observed in the BoB with the SLP gradient formed between northern India (e.g. the
674 monsoon trough) and the eastern equatorial IO, and the low-level wind pattern over
675 the eastern IO (north of the equator).

676 An important implication is that SLP biases can be a major source of errors for
677 the simulated 850hPa wind pattern over the IO, which brings the moisture over India
678 during monsoon. This is clearly the case in ALB1 simulation: a positive SLP bias
679 over the ME region weakens the SLP gradient over the AS and the Findlater jet, while
680 a negative SLP bias over the monsoon trough region enhances the SLP gradient over
681 the southern BoB and, hence, the 850hPa zonal wind in this same region, carrying
682 away the moisture from the BoB further eastward (Figs 2c-f).

683 In addition, SLP biases in our model are directly related to surface temperature
684 biases over land, which, in turn, are related to albedo errors as demonstrated above.
685 Following the ideal gas law, an air temperature increase (decrease) is associated with
686 an air density decrease (increase), which reduces (rises) the SLP. Consequently, LST
687 biases drive errors in the pattern of SLP gradient between land and ocean, which have
688 a direct consequence on the simulated low-level circulation over the ocean. This link
689 between LST, SLP and 850hPa wind explains most of the differences in the monsoon
690 flow pattern over the ocean between ALB2 and ALB1 (Fig. 8). Over the AS, where
691 the SLP gradient is stronger in ALB2 than in ALB1, stronger and shifted (northward)
692 850hPa wind are also found. Conversely, over the eastern IO and the South China
693 Sea, the weaker SLP gradient in ALB2 compared to ALB1 induces a weaker and less
694 zonal monsoon flow over these regions. The positive SLP gradient and 850hPa wind
695 differences observed over the northern AS, northern BoB and China Sea are the
696 signature of a greater northward extension of the monsoon flow in ALB2 than in
697 ALB1. As it turns more northward, the monsoon flow reaches the Himalaya foothills
698 where it brings more orographic precipitation (Fig. 6e). On the contrary, precipitation
699 is decreased over South-East Asia and South China Sea where the wind is reduced.

700

701 4.5. Albedo effect on the ISM seasonal evolution

702 The temporal ISM evolution is also modified by the albedo change from
703 ALB1 to ALB2. Figure 9 shows the annual cycle of SLP gradient between the ME

704 region and a western equatorial IO box (60° - 80° E / 5° S- 5° N), the 850hPa wind speed
705 annual cycle over the AS (40° - 75° E / 0 - 26° N) and precipitation over the monsoon
706 core region (65° - 100° E / 5° - 30° N).

707 The SLP gradient is positive during winter (from November to February), then
708 turns negative during summer corresponding to the monsoon onset and the seasonal
709 reversal of the Findlater jet (Fig. 9a). Interestingly, there is almost no difference
710 between ALB1 and ALB2 during boreal winter. Subsequently, the SLP gradient
711 grows faster in ALB2 than in ALB1 consistent with the seasonal increase of solar
712 radiation over the northern part of the domain from March to June. Consequently, a 1-
713 month time lag progressively builds up between the simulated SLP gradient in the two
714 simulations. The difference reaches its maximum in July during the monsoon peak.
715 The SLP gradient in ALB2 is also in much better agreement with the corresponding
716 estimates from ERA-interim.

717 Furthermore, the seasonal variability of SLP gradient is mainly driven by the
718 low SLP over land because the SLP over the equatorial IO remains relatively steady
719 along the year (not shown; see also Li and Yanai 1996). So the ALB2-ALB1 SLP
720 gradient differences originate mainly from the SLP differences over the ME region,
721 and ultimately, from the LST differences.

722 This time lag between ALB1 and ALB2 directly impacts the wind reversal
723 timing over the AS and the strengthening of the Findlater jet during the monsoon (Fig.
724 9b). The monsoon flow begins about one month earlier in the ALB2 simulation
725 compared to ALB1 and its time evolution is in better agreement with ERA-Interim
726 reanalysis. The peak wind speed is also stronger in ALB2 than in ALB1, by about 3
727 $\text{m}\cdot\text{s}^{-1}$. The maximum wind intensity reached during July in ALB2 is even greater than
728 in ERA-Interim due to a positive wind bias between the equator and 10° N (Fig. 6f).
729 The earlier and stronger monsoon onset in the AS directly influences precipitation
730 over India and the BoB (Figs. 9c-d). Precipitation increases more rapidly in ALB2
731 and its seasonal cycle is consistent with TRMM observations over land: whereas the
732 rainfall maximum is delayed by about two months in ALB1, its timing and magnitude
733 is much better captured in ALB2 simulation. Finally, the continental dry bias is also
734 well attenuated in ALB2, throughout the monsoon season (Fig. 9d).

735

4.6. Discussion on the relative influence of resolution and albedo on the ISM

In the previous sections, we have shown that land surface properties and resolution appear as two major sources of improvement in our model. However, several questions arise from these results. Is the reduction of monsoon biases observed at higher resolution in the forced HIRES simulation robust in a coupled ocean-atmosphere simulation? Is the albedo influence on the ISM the same at 0.75° and 0.25° resolutions? And, finally, are the high-resolution and albedo positive effects on the ISM simulation additive? To address these questions, we carried out two 0.25° -resolution 20-years coupled simulations using, respectively, ALB1 and ALB2 albedo (ALB1HR and ALB2HR, respectively; see Section 2 and Table 2 for details).

The benefit of increasing the horizontal resolution can be assessed by comparing ALB1 and ALB1HR simulations (Figs 10a-b). ALB1HR surface temperature is globally colder than ALB1, except in the western Tibetan plateau where a strong warming is observed (5 to 10°C). This warming is again related to the snow cover, which has a reduced spatial extension in ALB1HR compared to ALB1 (not shown). This change in the snow cover is directly related to the better representation of the orography at 0.25° resolution, which allows to represent separately the Himalayan mountain range and the Tibetan plateau. At 0.75° resolution, such distinction is not possible, which induces important errors in the snow cover and, consequently, in the surface temperature. A wide region extending from central India to north of the BoB and the Himalayan foothills is colder in ALB1HR than in ALB1. This surface cooling ranging from 2 to 6°C is directly related to the increased precipitation in the same regions (Fig. 10b). A significant rainfall increase is also observed at 0.25° resolution in regions of strong orographic forcing. On the contrary, precipitation is decreased over South China Sea and over the Maritime Continent region. Interestingly, no significant change is observed in the large-scale monsoon circulation (Fig. 10b), which suggests that the rainfall differences between ALB1HR and ALB1 are mainly related to local changes and not to large-scale environment modifications. A similar statement can be made by comparing the FORC and HIRES simulations described in Section 3.2.

The sensitivity of the simulated ISM to the land surface albedo is very similar at 0.75° resolution (ALB2-ALB1) and 0.25° resolution (ALB2HR-ALB1HR) as shown in Figs. 6c-e and 10c-d, respectively. A large land surface warming and SLP

769 decrease, directly related to the albedo change, are observed over the ME region and
770 the western part of the Tibetan plateau at both resolutions. The warming is roughly
771 the same at both resolutions, except in some localized places of the Tibetan plateau,
772 where the warming is greater at 0.75° resolution. On the other hand, the surface
773 cooling observed at 0.75° resolution over southern India, Bangladesh and China is
774 well attenuated at 0.25° resolution, where it only reaches 1°C locally. Concerning the
775 precipitation over land, the change due to albedo shows a similar impact at both
776 resolutions, even if the rainfall increase is more concentrated along the western Ghats
777 and the Himalaya foothills at 0.25° resolution (Fig. 10d). Contrarily to the resolution
778 increase, albedo change induces a large-scale strengthening of the simulated ISM
779 flow, which brings more humidity, and hence more precipitation, over land. This
780 mechanism appears to be robust at the two different resolutions we considered (e.g.
781 0.75 and 0.25°).

782 Finally, Figures 10e-f show that the benefits from high-resolution and
783 modified land surface albedo are clearly cumulative in terms of surface temperature
784 and precipitation biases. The net and significant result is a warming of the ME region
785 and the western Tibetan plateau and a cooling over continental India and Bangladesh.
786 Precipitation is significantly increased along the Ghats, the Himalayan foothills, the
787 Myanmar mountains and -though to a lesser extent- over continental India. The low-
788 level circulation strengthening and northward migration of the ITCZ are almost
789 entirely related to the albedo change as increasing the horizontal resolution does not
790 significantly modify the 850hPa wind pattern (Fig. 10b). The combination of
791 modified albedo with high resolution significantly reduce ISM biases (see Fig. 10g-h),
792 but a significant (limited) warm (dry) bias persists over India. A wet bias also persists
793 over South-East Asia and the South China Sea, which is related to a too strong low-
794 level wind circulation over the same region and the BoB. Those biases are directly
795 related to the warm temperature and low SLP biases over India.

796

797 **5. Conclusion and Perspectives**

798

799 5.1. Summary

800 The present study revisits the mechanism originally presented by Charney et
801 al. (1977) linking land surface albedo, surface heating and the ISM characteristics in a
802 state-of-the-art general circulation model extending between 45°S and 45°N. More
803 precisely, we demonstrate that constraining land surface heating by using observed
804 albedo climatology leads to significant improvements in ISM simulation with our
805 model, especially in terms of the ISM rainfall onset and climatology. Moreover, we
806 illustrate that our results are valid in both coupled and forced frameworks, at two
807 spatial resolutions (0.75 and 0.25°) and with two albedo datasets (AVHRR and
808 MODIS), hereby highlighting the robustness of our findings.

809 These results emphasize the important role of the non-elevated land surface
810 heating pattern on the ISM: the Middle-East area appears as a key region, which
811 exerts a strong control on the meridional migration on the ITCZ through its warming
812 pattern and amplitude. This is consistent with results from Boos and Kuang (2013),
813 suggesting that the monsoon responds significantly to surface heat fluxes associated
814 with temperature maxima. The mechanism proposed in our study to explain the ISM
815 biases is different from the Tibetan plateau theory described by Li and Yanai (1996),
816 in which the sensible heat flux from this high-elevated surface directly contributes to
817 the reversal of the meridional temperature gradient. Here, the land surface heating
818 locally lowers the surface pressure, which modifies the large-scale pressure gradient
819 between the ME region and the western equatorial IO. The low-level circulation
820 adjusts to these changes in the SLP gradient and directly affects the humidity
821 transport necessary for improving continental precipitation in our simulations.
822 Concretely, the JJAS Indian land dry bias, which is about 46% (-3.6 mm/day) in
823 ALB1 compared to TRMM (7.9 mm/day), is reduced to 18% (-1.4 mm/day) in ALB2.
824 The ISM duration in ALB2 is also extended by 1 month in agreement with TRMM
825 observations. This suggests that surface heating may play an important role in
826 modulating the ISM biases, even though the deep low over the ME region cannot be
827 purely considered as a “heat” low, as demonstrated by Bollasina and Nigam (2011).

828 Another important implication of this result is that any significant LST bias
829 over the northern plains of India can generate errors in the representation of the
830 monsoon trough, through the mechanism discussed above. This is clearly the case
831 with the warm temperature and low SLP biases over India, which strengthen the
832 pressure gradient between India and the eastern equatorial IO. The associated zonal

833 wind intensification brings too much rainfall to South-East Asia and South China Sea,
834 instead of feeding the monsoon trough region.

835 Horizontal resolution also appears as a key parameter to improve the ISM
836 representation in both forced and coupled configurations of our model. Precisely, the
837 JJAS Indian land dry bias is reduced by 31% (+1.3 mm/day) between ALB1 (4.3
838 mm/day) and ALB1HR (5.6 mm/day) and by 11% (+0.7 mm/day) between ALB2 (6.5
839 mm/day) and ALB2HR (7.2 mm/day). Increasing the horizontal resolution also
840 improves the rainfall pattern correlation over the Indian region from 0.5 to 0.7 with
841 both ALB1 and ALB2 albedos. The absence of modification in the low-level
842 circulation between ALB1 and ALB1HR also suggests that a 0.75° resolution is fine
843 enough to resolve the main orographic features necessary to prevent the ventilation
844 mechanism with cold and dry air from high latitudes described by Chakraborty et al.
845 (2002, 2006) and by Boos and Kuang (2010). On the contrary, the absence of large-
846 scale atmospheric response to the strong warming observed in the western Tibetan
847 plateau supports the idea that the Tibetan plateau is not a dominant source of heating
848 for the ISM. Nonetheless, supplementary experiments following Boos and Kuang
849 (2010, 2013) and Ma et al. (2014) methodology would be necessary to precisely
850 assess the respective roles of the Himalayan mountains and Tibetan plateau heating
851 effects in our model.

852

853 5.2. Perspectives

854 Understanding the development of the Indian warm LST bias over the Indo-
855 Gangetic plains during boreal spring and its maintenance during the monsoon season
856 is of critical importance for future ISM studies. Furthermore, an important number of
857 CGCMs suffer from the same caveats as recently illustrated by the CMIP5 ensemble
858 mean (Sooraj et al. 2015). Consequently, these models could also benefit from
859 substantial improvements in terms of monsoon representation if the Indian warm LST
860 bias was successfully understood and corrected.

861 Various promising directions can be followed to improve LST and rainfall
862 over continental India in current state-of-the-art climate models. Concerning
863 specifically the WRF-NOAH LSM model, a necessary step would be the
864 implementation of a complete land surface albedo parameterization, such as in

865 NCEP/GFS (Hou et al. 2002), NCAR/CAM (Bonan et al. 2002) and ECHAM6
866 (Brovkin et al. 2013) models. This would allow a more realistic computation of the
867 surface SW fluxes, and consequently an additional LST bias reduction. Other domains
868 of improvement concern the representation of soil characteristics and irrigation in the
869 land surface models (Saeed et al. 2009; Kumar et al. 2014), convection
870 parameterization (Ganai et al. 2015) or further horizontal grid refinement (Sabin et al.
871 2013) to correctly capture all the important processes, which contribute to ISM
872 rainfall. Furthermore, the impact of SST biases on ISM in remote regions and not only
873 in the IO must be also properly evaluated in a coupled framework (Prodhomme et al.
874 2015). Concerning RCMs, our study emphasizes the importance of including the ME
875 region in the model domain when simulating the ISM in order to correctly represent
876 the large-scale land-sea pressure gradient which drives the low-level monsoon flow.

877 Finally, due to the large diversity of the albedo estimation in current CGCMs
878 and RCMs (Wang et al. 2007), similar experiments with other models are clearly
879 needed to demonstrate that the results presented here are robust and may lead to
880 improvements in our capability to predict the monsoon at different time scales or to
881 assess the future of the monsoon in a global warming context. Such improvements of
882 monsoon simulations are of utmost importance for the society and the livelihood of
883 the population in South Asia (Annamalai et al. 2015; Sabeerali et al. 2015; Wang et
884 al. 2015).

885

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897

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1202

1203 **Table Captions**

1204 **Table 1:** List of the acronyms used.

1205 **Table 2:** Summary of tropical-channel simulations. Differences between the
1206 simulations configurations are given in the “Setup” column.

1207

1208 **Figure Captions**

1209 **Figure 1.** (a) MODIS-IGBP dominant land-use categories. (b) Annual snow-free
1210 black-sky broadband SW land albedo from MODIS product (%). Note that in all
1211 figures, ocean albedo is not displayed for clarity. (c) Time-average difference between
1212 ALB1 albedo and MODIS snow-free product (%). (d) Same as (c) for ALB2 (%).

1213 **Figure 2.** (a) Summer monsoon (JJAS) mean climatological ERA-Interim surface
1214 temperature ($^{\circ}\text{C}$, shaded) and SLP (hPa, contours; contours greater than 1020 hPa are
1215 not drawn for clarity). (b) Same as (a) for ALB1. (c) ALB1 surface temperature ($^{\circ}\text{C}$,
1216 shaded) and SLP (hPa, contours) biases compared to ERA-Interim (contours bias
1217 greater than ± 8 hPa are not drawn for clarity). (d) TRMM precipitation (mm/day,
1218 shaded) and ERA-Interim 850hPa wind (m/s, vectors). (e) Same as (d) for ALB1. (f)
1219 Biases of ALB1 precipitation and wind computed as the difference between (b) and
1220 (a).

1221 **Figure 3.** (a) Rainfall monthly seasonal climatology (mm/day) averaged over land
1222 only in the $65^{\circ}\text{-}100^{\circ}\text{E} / 5^{\circ}\text{-}30^{\circ}\text{N}$ box (see inset map for box limits) for TRMM
1223 observations (black) and for the various numerical simulations (colors); see Table 2
1224 for the description of the experiments. (b) Same as (a), averaged over land and ocean.
1225 The dashed lines show the annual long-term mean of the various climatologies.

1226 **Figure 4.** (left column) JJAS rainfall (mm/d, shaded) and 850hPa wind (m/s, vectors)
1227 biases of the various sensitivity experiments, compared to TRMM and ERA-Interim
1228 datasets, respectively: (a) FRC, (b) HIRES, (c) CONV and (d) RAD; see Table 2 for
1229 the description of these experiments. (right column) JJAS surface temperature ($^{\circ}\text{C}$,
1230 shaded) and SLP (hPa, contours) biases compared to ERA-Interim dataset : (a) FRC,
1231 (b) HIRES, (c) CONV and (d) RAD.

1232 **Figure 5.** (a) Pre-monsoon (MAM) mean climatological ERA-Interim surface
1233 temperature ($^{\circ}\text{C}$, shaded) and SLP (hPa, contours). (b) Biases of ALB1 surface
1234 temperature and SLP compared to ERA-Interim.

1235 **Figure 6.** (a) JJAS mean climatological albedo difference between ALB2 and ALB1
1236 (unit %). (b) ALB2 albedo bias compared to MODIS (unit %). (c) Surface
1237 temperature ($^{\circ}\text{C}$, shaded) and SLP (hPa, contours) differences between ALB2 and
1238 ALB1. (d) ALB2 surface temperature ($^{\circ}\text{C}$, shaded) and SLP (hPa, contours) biases
1239 compared to ERA-Interim. (e) Precipitation (mm/day, shaded) and 850hPa wind (m/s,

1240 vectors) differences between ALB2 and ALB1. (f) ALB2 precipitation (mm/day,
1241 shaded) and 850hPa wind (m/s, vectors) biases compared to ERA-Interim.

1242 **Figure 7.** (a) JJAS mean climatological surface radiative heat fluxes averaged over
1243 the “Middle East” region (see inset map for box limits, only the land points in the box
1244 are considered). Black, blue and red bars show CERES-EBAF, ALB1 and ALB2
1245 estimates, respectively. The bars from left to right are for downward shortwave
1246 (SW_DN), upward shortwave (SW_UP), net shortwave (SW_NET), upward
1247 longwave (LW_UP), downward longwave (LW_DN), net longwave (LW_NET) and
1248 total radiative heat fluxes (SW+LW_NET) at land surface (in W/m^2). (b) same as (a),
1249 for ALB1 and ALB2 errors compared to CERES-EBAF dataset.

1250 **Figure 8.** JJAS mean climatological differences of SLP gradient (Pa/km, shaded) and
1251 850 hPa wind (m/s, vectors) differences between ALB2 and ALB1 experiments. Land
1252 surfaces are masked for clarity.

1253 **Figure 9.** (a) SLP (hPa) monthly climatological seasonal cycle difference between the
1254 Middle-East ("ME") and the Western Equatorial IO ("WIO") regions in ERA-Interim
1255 (black), ALB1 (blue) and ALB2 (red). The boxes limits are featured on the inset map.
1256 (b) 850hPa wind (m/s) 5-days climatological seasonal cycle averaged over the
1257 Arabian Sea ("AS", see inset map for box limits). (c) Rainfall (mm/day) 5-days
1258 climatological seasonal cycle averaged over an extended Indian domain (65° - 100° E /
1259 5° - 30° N, see inset map for box limits). (d) Same as (c), for the land area of the box
1260 only.

1261 **Figure 10.** (a) JJAS mean climatological differences of surface temperature ($^{\circ}$ C,
1262 shaded) and SLP (hPa, contours) between ALB1 and ALB1HR experiments. (b) JJAS
1263 mean climatological differences of precipitation (mm/day, shaded) and 850hPa wind
1264 (m/s, vectors) differences between ALB1 and ALB1HR. (c) Same as (a), but between
1265 ALB2HR and ALB1HR. (d) Same as (b), but between ALB2HR and ALB1HR. (e)
1266 Same as (a), but between ALB2HR and ALB1. (f) Same as (b), but between ALB2HR
1267 and ALB1. (g) JJAS mean climatological biases of surface temperature ($^{\circ}$ C, shaded)
1268 and SLP (hPa, contours) of ALB2HR compared to ERA-Interim. (h) JJAS mean
1269 climatological biases of precipitation (mm/day, shaded) and 850hPa wind (m/s,
1270 vectors) ALB2HR biases compared to TRMM and ERA-Interim, respectively.