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A subsurface Indian Ocean Dipole response to tropical volcanic eruptions

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Key Points:

- Tropical volcanic eruptions induce a subsurface ocean response typical of a negative Indian Ocean Dipole (IOD).
- The subsurface response results from equatorial westerlies driven by the large land cooling over Africa.
- The surface signature differs from that of an IOD, with a cooling larger in the northwest than southeast Indian Ocean as a result of the climatological cloud cover distribution.

Keywords: Tropical volcanic eruption, Indian Ocean, Indian Ocean Dipole (IOD), ocean Primary Productivity (PP), Coupled Model Intercomparison Project (CMIP).

29 **Abstract.** The impacts of explosive volcanism on the densely-populated Indian
30 Ocean (IO) region remain elusive. Dedicated sensitivity experiments indicate that
31 tropical volcanic eruptions induce a stronger surface cooling over Africa than of ocean,
32 promoting westerlies in the equatorial IO. These westerlies drive a subsurface response
33 reminiscent to that of a negative IO Dipole (IOD) during autumn in the year of eruption.
34 The eruption also drives an enhanced cooling over the northwestern IO as a direct
35 response to climatological cloud cover distribution. The resulting anomalous zonal sea
36 surface temperature gradient contributes to enhance equatorial westerly anomalies in
37 summer. The response is sensitive to the IO preconditioning, being larger when the
38 system is favourable to a positive IOD development. Volcanic eruptions also induce a
39 subsurface IOD-like response in the multi-model database from the Coupled Model
40 Intercomparison Project Phase 5 as well as a primary productivity decrease in the
41 eastern IO.

42

43 **Plain text summary.** Explosive volcanism induces a global surface cooling but its
44 regional impact on the densely-populated Indian Ocean sector is poorly documented.
45 Here we show that the volcanic eruptions induce a subsurface ocean response in
46 autumn of the eruption's year, reminiscent to that of a negative Indian Ocean Dipole, an
47 intrinsic climate mode in the Indian Ocean equivalent to El Niño in the Pacific. This
48 subsurface response is driven by equatorial westerly anomalies resulting from the
49 stronger surface cooling over the African landmass than over the ocean. The response is
50 sensitive to the Indian Ocean preconditioning, being larger when the system is
51 preconditioned towards a positive Indian Ocean Dipole development.

52

53 1) Introduction

54

55 The densely populated region surrounding the Indian Ocean (IO) is socio-
56 economically very sensitive to climate variability. As Sea Surface Temperatures (SSTs) in
57 the IO often exceeds the 28°C threshold for deep atmospheric convection (e.g. Graham
58 and Barnett, 1987), small SST perturbations can force a significant atmospheric
59 response. This sensitivity promotes active air-sea interactions, which give rise to the
60 Indian Ocean Dipole (IOD; Reverdin et al. 1986, Saji et al. 1999, Webster et al. 1999,
61 Murtugudde et al. 2000), an intrinsic mode of climate variability in the IO equivalent of
62 El Niño/La Niña Southern Oscillation (ENSO) in the Pacific. A negative IOD (nIOD) is
63 characterized by a surface warming and an anomalously deep thermocline off the west
64 coasts of Java and Sumatra and by a cooling and anomalously shallow thermocline
65 anomalies in the west IO, driven by anomalous westerlies over the central equatorial IO.
66 A positive IOD (pIOD) is characterized by opposite anomalies. IOD events usually
67 develop in boreal summer and peak in autumn. El Niño and La Niña events respectively
68 favour positive and negative IODs, but IOD can also occur independently of ENSO (e.g.
69 Annamalai et al. 2003; Yamagata et al. 2004). The IOD has important implications at
70 regional scale on the IO climate and monsoon (e.g. Schott et al. 2009), tropical cyclones
71 (Xie et al. 2002), oceanic primary production (Curie et al. 2013) and oxygen
72 (Vallivattathillam et al. 2017), and also at global scale by influencing the Madden-Julian
73 Oscillation (e.g. Izumo et al. 2010b) and ENSO (e.g. Izumo et al. 2010a, 2014, 2016,
74 Jourdain et al. 2016).

75 Stratospheric tropical volcanic eruptions induce a temporary global radiative
76 cooling that directly impacts the earth climate by inducing a surface cooling at the global
77 scale (see Robock, 2000 for a review). This cooling is larger on land than on ocean,
78 resulting in an overall rainfall reduction over the tropical landmass (e.g. Schneider et al.
79 2009, Joseph and Zeng 2011, Iles and Hegerl 2014). The upper ocean cooling translates
80 into an Oceanic Heat Content (OHC) decrease (e.g. Church et al. 2005, Gleckler et al.
81 2006, Stenchikov et al. 2009, Mignot et al. 2011). The climatological cloud cover and
82 oceanic mixed layer depth (MLD) distributions however induce a spatially
83 inhomogeneous ocean cooling: e.g. a clear sky and thin MLD induces a larger
84 cooling (McGregor and Timmerman 2011, Ohba et al. 2013). Such land/sea thermal
85 contrast and inhomogeneous ocean cooling also results in specific regional climate

86 responses. Notably, Khodri et al. (2017, hereafter K17) demonstrated using dedicated
87 sensitivity experiments that the larger cooling over tropical Africa than over the ocean
88 force westerlies over the western Pacific, hence favouring an El Niño the year following
89 the eruption.

90 While the Pacific response to tropical volcanic eruptions has been extensively
91 documented (see K17 and the references therein), there is to date no study focusing on
92 the detailed response of the IO to these eruptions. The present study aims to fill this gap
93 by analysing the Pinatubo-forced sensitivity experiments designed in K17 as well as
94 simulations from the Coupled Model Intercomparison Project phase 5 (CMIP5; Taylor et
95 al. 2012).

96

97 **2) Models, experimental strategy and methods**

98 **2.1 The IPSL-CM5B LR model and the experimental strategy**

99 In this study, we use the same set of simulations as in K17 performed with the IPSL-
100 CM5B-LR climate model (cf. Supplementary Information – hereafter SI, its section 2, for
101 details on the modelling setup). This model couples the LMDZ5B (Hourdin et al. 2013)
102 atmospheric and NEMO3.2 (Madec 2015) oceanic components, and accurately captures
103 the main observed IOD characteristics (Jourdain et al. 2016).

104 We focus on the Pinatubo eruption (June 15th, 1991), as it is the strongest and best-
105 observed tropical stratospheric eruption during the instrumental period. The main set
106 of coupled (CPL hereafter) simulations consists of a pair of 3-year-long ensemble
107 experiments of 30 members each with (PTUBO) and without (CTL) a Pinatubo-like
108 forcing. These simulations are initialized on June 1st of a selected year from the historical
109 simulation (with neutral El Niño Southern Oscillation (ENSO) conditions in the absence
110 of volcanic forcing, avoiding any interference with external ENSO forcing).

111 The CPL response to volcanic forcing is assessed by performing the difference
112 between PTUBO and CTL experiments (CPL = PTUBO-CTL), internal climate variability
113 being reduced by the ensemble average. Significance and confidence intervals are
114 computed from usual two-tailed Student t-tests, each member being an independent
115 degree of freedom.

116 We also analyse the following dedicated atmospheric experiments performed with
117 the atmospheric component of the IPSL-CM5B model (K17). They are forced by the

118 coupled model outputs to assess the dominant mechanisms driving the IO response:

119 (1) ATM experiments assess the direct effect of volcanic radiative forcing on clouds
120 and atmospheric vapour content (Soden et al. 2002).

121 (2) OCEAN experiments assess the indirect effects of volcanic forcing on the
122 atmospheric response through the SST inhomogeneous cooling.

123 (3) LAND experiments quantify the effect of the stronger cooling over land than
124 over the ocean. An additional sensitivity experiment finally isolates the specific role of
125 the cooling over African landmass (LAND-Africa).

126 Finally, the relevance of the two-tier approach is assessed using an experiment
127 including all these three processes (ALL).

128 *2.2 Linear Continuously Stratified (LCS) and slab ocean models*

129 The equatorial IO dynamical response to windstress anomalies resulting from each
130 sensitivity experiment is then estimated by forcing a Linear Continuously Stratified
131 (LCS; McCreary, 1980) ocean model of the Indo-Pacific region (Izumo et al. 2016, 2018
132 configuration; cf. SI). This LCS realistically simulates sea surface height (SSH) variations
133 in the Indo-Pacific region (Suresh et al. 2013, 2016; Izumo et al. 2016, 2018), despite
134 being linear and simple.

135 A simple slab ocean mixed layer equation using a spatially-varying monthly
136 climatological mixed layer depth is also used, to simulate the SST thermodynamical
137 response to the eruption-induced Short-Wave (SW) flux anomalies (see SI), following
138 the approach of McGregor and Timmerman (2011).

139 **2.3 CMIP5 simulations**

140 The robustness of the response from our single model experiments is further
141 assessed by analysing historical simulations from the CMIP5 multi-model dataset over
142 the 1861-2005 period (Suppl. Table S1). As reported in many previous studies (e.g.
143 Jourdain et al. 2016), CMIP5 simulations, in general, accurately capture the IOD
144 seasonality and pattern but overestimate its amplitude. Their response to tropical
145 volcanism is estimated here by compositing the 5 largest historical tropical eruptions,
146 keeping in mind that these eruptions exhibit different amplitudes and seasonal
147 evolutions (cf. Fig. S1).

148 We use the classical Dipole Mode Index (DMI, the SST difference between 50-70°E,
149 10°N-10°S west and 90-110°E, 0-10°S east poles; Saji et al. 1999) to identify IOD events,
150 as well as the SSH averaged over the DMI eastern pole box, the SSH being an accurate
151 proxy for OHC and thermocline depth variations (Rebert et al. 1985).

152

153 **3) Results**

154

155 Figure 1ac displays the surface and ocean subsurface ensemble mean CPL response
156 to the Pinatubo eruption. As expected, there is an overall cooling over the IO region,
157 stronger over land than at the ocean surface. The SST cooling is larger over the
158 northwest than over the southeast IO, leading to a northwest-southeast anomalous SST
159 gradient (Fig. 1a), which translates into a dipolar pattern in terms of relative SST (i.e.
160 SST minus its 20°N-20°S tropical mean; cf. SI, K17; Held and Soden 2006; Vecchi and
161 Soden 2007) and precipitation (Fig. S2). The equatorial IO is also characterized by
162 significant westerly/north-westerly anomalies, larger in the eastern IO. The SSH and
163 related subsurface oceanic response (Fig. 1c) is typical of a negative IOD (cf. Fig. S3;
164 Murtugudde et al. 2000, Currie et al. 2013), with a downwelling signal along the coast of
165 Java and Sumatra and the Bay of Bengal rim and an upwelling signal in the central and
166 western IO maximum off the equator.

167 The LCS experiment forced by CPL ensemble wind stress (Fig. 1d) accurately
168 captures the major SSH signature simulated by the coupled model (Fig. 1c),
169 demonstrating that linear ocean dynamics explains most of the coupled model response.
170 There is however a positive SSH offset over the entire IO, because the LCS physics
171 cannot simulate in essence the SSH decrease related to the overall upper ocean cooling

172 driven by volcanically-induced surface heat flux anomalies (e.g. Church et al. 2005,
173 Gleckler et al. 2006, Stenchikov et al. 2009). The volcanically-induced wind stress
174 anomalies easily explain the SSH dipolar pattern: equatorial westerlies indeed induce
175 downwelling equatorial Kelvin waves deepening the thermocline in the eastern IO and
176 westward propagating equatorial Rossby waves shallowing the thermocline in the
177 central and western IO.

178 As the IO subsurface response to volcanic forcing is qualitatively similar to that of a
179 negative IOD (hereafter nIOD) but different at the surface, we will term this response as
180 a « volcanic subsurface nIOD ». This ensemble mean response is about one-third to a half
181 of the typical SSH amplitudes during a typical IOD simulated by the IPSL model (Fig. S3).
182 It is significant at the 99% level for the DMI east pole SSH. The SST pattern is different
183 from that of a typical IOD (Fig. S3), although it projects onto DMI poles (Figs 1a and S2a),
184 leading to a significant negative DMI anomaly of -0.4°C .

185 The subsurface ocean response is largely wind driven (Fig. 1cd). We hence analyse
186 the dedicated sensitivity atmospheric experiments described in Section 2 to understand
187 the mechanisms responsible for the anomalous equatorial westerlies. The similar wind
188 patterns depicted in CPL and ALL experiments confirm the relevance of this two-tier
189 approach (Fig. 1ab). Despite a 30% overestimation of the CPL response at its peak, the
190 temporal evolution of equatorial zonal wind-stress (Fig. 2a) and SSH (Fig. 2c) anomalies
191 in ALL experiment are also not statistically distinguishable as a result of the large inter-
192 members spread. LAND process explains most of the equatorial westerly anomalies and
193 related east IO downwelling (Fig. 2b,d). OCEAN process also modestly contributes to
194 these westerly and eastern IO downwelling anomalies at the end of the summer. Finally,
195 ATM influence is negligible, and will not be further discussed. Figure 2e shows the
196 corresponding surface temperature and wind-stress patterns in autumn for the LAND
197 experiment. As expected from Figure 2b, LAND experiment exhibits equatorial
198 westerlies similar to those in ALL and PTUBO CPL experiments in the equatorial region
199 (Fig. 1). LAND also displays additional strong northeasterly winds in the Arabian Sea,
200 consistent with the expected thermal wind response to the Arabic peninsula cooling (e.g.
201 Samson et al. 2016).

202 Applying the LAND cooling only over tropical Africa (LAND-Africa) results in a
203 similar equatorial zonal wind (Fig. 2b,f) and SSH (Fig. 2d) evolution as those depicted for
204 LAND, implying that these equatorial westerlies are mainly driven by the African land

205 cooling (K17). These westerly anomalies are however shifted northward in this LAND-
206 Africa experiment, most likely because of the absence of the Asian land cooling and
207 related winter monsoon-like circulation anomalies in the eastern Arabian Sea and Bay of
208 Bengal. The African cooling reduces the rainfall and related diabatic heating locally,
209 thereby inducing a Matsuno-Gill response (Gill 1980) with an equatorial atmospheric
210 Kelvin wave forcing to the east of the heating anomaly, consistent with K17 finding.
211 Milder - but significant - westerly anomalies developing in the central east equatorial IO
212 in July-August in OCEAN experiment (Fig. 2b) are arguably a direct response to the
213 northwest-southeast SST gradient seen in Figs 1a and S2. To summarize, Africa-IO land-
214 sea contrast explains the westerlies and related subsurface nIOD pattern around its
215 peak, but anomalous zonal SST gradients also play a significant role at the early stage of
216 this nIOD.

217 To identify the main driver of the northwest-southeast SST gradient, we performed
218 slab ocean sensitivity experiments forced by the PTUBO minus CTL SW anomalies. They
219 indicate that the inhomogeneous SST cooling largely arises from the climatological cloud
220 cover distribution: the larger climatological cloud cover in the eastern equatorial IO
221 indeed acts to reduce the volcanic SW anomaly at the surface, resulting in a reduced
222 volcano-induced cooling compared to the north-western IO (Fig. S4). The resulting
223 anomalous spatial SST pattern matches qualitatively well with the climatological cloud
224 cover effect

225 We then investigate the sensitivity of the CPL response to the IO preconditioning,
226 defined as the IOD condition that would happen in autumn in the absence of the
227 Pinatubo eruption (i.e. in the CTL simulation). Such sensitivity has already been
228 discussed for ENSO (Ohba et al. 2013; Pausata et al. 2016; K17), but not for the IOD. The
229 dependence of the IO response to the Indo-Pacific initial preconditioning is assessed by
230 analysing two other pairs of 30-members ensemble experiments with and without a
231 Pinatubo-like forcing characterized respectively by El Niño and La Niña conditions in the
232 Pacific during the eruption's year (see SI; K17). We gather the three CTL 30-members
233 ensembles into a large 90-members ensemble, and then classify them depending on
234 their IOD conditions in autumn of the eruption's year, gathering members showing a
235 DMI above 0.5°C (21 members), below -0.5°C (49 members), and between -0.5°C and
236 0.5°C (20 members). Classifying members based on ENSO conditions in the Pacific gives
237 qualitatively similar results (with slightly less differentiation), because of the tendency

238 of ENSO and IOD events to co-occur.

239 Figure 3 displays the evolution of the DMI (Fig. 3a-c) and eastern SSH (Fig. 3d-f)
240 anomalies in CTL and PTUBO experiments for these three classes (even if the SST
241 response is different from that of an IOD, the DMI is used here to describe the SST
242 asymmetries as it broadly captures the spatial anomalies associated with this SST
243 cooling; Fig. S5 and S6). In the absence of an IOD in CTL experiment, volcanic eruptions
244 favour a nIOD-like response both at depth (~ 2 cm SSH anomaly in the eastern IO; Fig.
245 3d) and at the surface (-0.4°C DMI anomaly; Fig. 3a). In the presence of a positive IOD in
246 CTL, volcanic eruptions induce a similar subsurface response (~ 2 cm; Fig. 3de) but a
247 larger SST response (-0.7°C ; Fig. 3ab), which acts to considerably reduce the positive
248 IOD amplitude. In the presence of a negative IOD in CTL, volcanic eruptions induce a
249 weaker subsurface response (~ 1.2 cm; Fig. 3f) and no significant SST gradient change
250 (Fig. 3c). The asymmetry to IO background conditions is hence larger for the SST
251 gradient than for the subsurface response. The asymmetry of the SST response mainly
252 arises from the IOD eastern pole (Fig. S5c), and could be explained by two processes
253 described below in the case of a negative IOD background condition. First, because of
254 the non-linearity of the Clausius-Clapeyron equation, the evaporation negative feedback
255 damps more the volcanically-induced relative warming in the eastern pole during a
256 negative IOD as SSTs are warmer in the eastern IO. Second, the deep background
257 thermocline in the eastern IO during a negative IOD makes the positive thermocline
258 depth anomalies forced by the eruption westerlies less efficient to warm the eastern IO
259 SST. While the asymmetry in the subsurface response is weaker, the presence of a
260 negative IOD in CTL experiment makes the subsurface response ~ 1.5 -2 times weaker,
261 shorter, and occurring later (only significant at the 90% level in Nov.-Dec.; not shown).
262 This suggests that the volcanically-induced SST gradient could contribute, by forcing
263 additional westerlies, to the subsurface IOD, and to its asymmetry. However, our
264 modelling setup can not precisely quantify this contribution from the ocean-atmosphere
265 coupling.

266 The robustness of the IO response to volcanic forcing derived from our single model
267 analysis is finally assessed by analysing the multi-model CMIP5 database. Composites of
268 the five historical tropical eruptions in autumn of the eruption's year in the CMIP5
269 multi-model ensemble mean (48 models/158 members, cf. SI) show patterns of surface
270 temperature, relative SST, wind stress (Fig. 4a) and sea level (Fig. 4b) that share a lot in

271 common with those obtained for our ensemble strategy using the IPSL-CM5B model for
272 the Pinatubo eruption (Fig. 1ac). This includes a larger cooling over land than over the
273 ocean, a larger SST cooling over the northwest IO, an equatorial westerly anomaly and a
274 significant subsurface oceanic response typical of a nIOD.

275 This overall consensus between our simulations and CMIP5 multi-model database
276 suggests that the results discussed in this paper are relevant for most climate models
277 used in the CMIP5 exercise. There are however significant differences between our
278 single model experiments and the CMIP5 MME multi-eruption response, generally
279 weaker (e.g. nIOD subsurface signal of only ~20% of that of a typical IOD in CMIP5 MME,
280 instead of ~30-50% in CPL). This amplitude however considerably varies among models
281 (not shown). This weaker response in the MME likely partly arises from our set-up,
282 specifically designed for the Pinatubo eruption (Fig. S1). The spread of the inter-model
283 response could also be related to the different models physics and to internal variability
284 (and random preconditioning) of each historical run.

285

286 4) Discussion

287

288 Analyses of dedicated coupled experiments with the IPSL-CM5B model forced by a
289 Pinatubo-like eruption and CMIP5 models historical runs demonstrate that tropical
290 volcanic eruptions force a subsurface response typical of nIOD. The surface response
291 exhibits an enhanced cooling over the northwestern IO, resulting in a
292 northwest/southeast sea surface temperature gradient. Although this inhomogeneous
293 SST pattern projects onto the classical DMI, it differs from the classical IOD SST pattern.
294 We thus term this response a « volcanic subsurface nIOD ». A hierarchy of models shows
295 that this nIOD-like subsurface signature is caused by equatorial westerly anomalies, that
296 are themselves largely an atmospheric Kelvin wave response to the fast land cooling
297 over tropical Africa. These westerlies seem to be also enhanced by the heterogeneous
298 ocean surface cooling, attributable to the heterogeneous climatological cloud cover.
299 Finally we show that the SST and subsurface responses depend on the IO
300 preconditioning, with a larger response when the system is favourable to a positive IOD
301 development, especially for the SST.

302 This volcanic subsurface nIOD is highly statistically significant but its amplitude is

303 still modest, ~30-50% of a typical IOD in IPSL CM5B. For neutral background conditions,
304 the amplitude of the DMI SST response reaches ~0.4°C, a value comparable to the ENSO
305 response in the same model (0.5°C Niño3.4 relative SST anomalies; K17). Given that the
306 Pinatubo eruption is relatively modest compared to other eruptions over longer
307 geological timescales, we expect the volcanic nIOD to be much stronger for very strong
308 eruptions, e.g. during the last millennium. Volcanic nIOD amplitude also considerably
309 varies among CMIP models, and will certainly depend on the model, the volcanic forcing,
310 and the eruption characteristics (strength, season, duration, latitude; Stevenson et al.
311 2016). The OCEAN process depends likely also on the volcanically-induced El Niño
312 response, which could interfere destructively with the IO local SST response by forcing
313 IO easterlies. Alleviating the dry bias in the eastern IO present in the current CMIP
314 models (e.g. Li et al. 2015) could also increase the SST gradient response by increasing
315 the climatological cloud cover effect (cf. Fig. S4), and therefore the OCEAN effect. All
316 these questions highlight the need of similar experiments in other models, as proposed
317 in the coming CMIP6 VolMIP exercise (Zanchettin et al. 2016).

318 Here we have demonstrated that a *negative* subsurface IOD develops in autumn of
319 the eruption's year. Maher et al. (2016) has interestingly briefly mentioned the tendency
320 for a *positive* IOD-like SSH pattern to develop in autumn of the eruption's *following* year
321 in CMIP models and attributed this to an indirect response to the concurrent El Niño
322 signal peaking in the Pacific. Indeed, CMIP5 MME IOD temporal evolution indicates both
323 the direct *negative* IOD response in the eruption's year and the indirect *positive* IOD the
324 following year (Fig. S6). Removing ENSO influence through linear regression confirms
325 that the positive IOD occurring the year after the eruption is mainly an indirect response
326 to the co-occurring El Niño in the Pacific, while the negative IOD response during the
327 year of the eruption is not. The direct response in CMIP MME (+0.8 cm SSH in SON of
328 eruption's year for eastern DMI pole) is of the same order of magnitude as the El Niño-
329 induced positive IOD in following year's SON (-1.2 cm; not shown).

330 The « volcanic subsurface negative IOD » does not exhibit a typical IOD signature at
331 the surface, resulting in different climate impacts. The rainfall pattern (and thus
332 atmospheric teleconnections) is indeed different (Figs S2b and S7b) from a typical IOD
333 (Fig. S3): e.g. the strong land cooling over the maritime continent favours a drying, while
334 the negative IOD-like relative SST warming (Figs 1a, S2a and S7a) favours rainfall over
335 the east IO.

336 In contrast, subsurface impacts are similar to that of a typical nIOD. Primary
337 production (PP) response in the CMIP5 models (Fig. 4c) is a mirror of SSH, notably in the
338 east IO, Bay of Bengal and Arabian Sea. This PP response fits qualitatively well with the
339 PP patterns expected from a typical IOD (e.g. Murtugudde et al. 1999, Levy et al. 2007,
340 Curie et al. 2013): a ~20% PP decrease along Sumatra (with 16 out of 18 models
341 agreeing on the sign) and a more widespread but weaker PP increase in the west IO,
342 notably along India west coast (with likely a reduction in coastal anoxia and modulation
343 of the oxygen minimum zone; Vallivattathillam et al. 2017). Concerning the IO OHC, IO
344 equatorial westerlies/negative IOD tend to discharge the tropical IO, as shown by
345 idealised switched-on westerlies LCS experiments as in Izumo et al. (2018) study (not
346 shown). El Niño also favours an IO OHC discharge via the Indonesian throughflow
347 (Sprintall and Revelard 2014; Lee et al. 2015). Hence, a tropical eruption, by favouring
348 this nIOD the eruption's year, as well as a following El Niño (Fig. S8; K17), forces a
349 significant IO OHC discharge starting in the eruption's year, the negative OHC anomaly
350 lasting then for several years in the IPSL model (not shown).

351 Understanding the sensitivity of this volcanic nIOD response to the amplitude,
352 latitude and seasonality of the eruption is the next step forward for improving post-
353 eruption climate forecasts.

354

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365 downloaded from the Institut Pierre Simon Laplace (IPSL) CICALAD repository. A more detailed discussion
366 can be found in the supporting information [Sobel et al. 2001 ; Saji et al. 2006; Joshi et al. 2008 ; Gao et al.
367 2008].

368

369

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512

513 **Figures captions**

514 **Figure 1. Indian Ocean response in the autumn following a Pinatubo-like eruption.**

515 (a) 2m air temperature (T2m, in °C, shading) and surface wind stress (N m⁻², vectors)
516 difference between PTUBO and CTL ensemble coupled experiments (CPL) in autumn
517 (September-November), a few months after the eruption in June. (b) as in (a), but for the
518 ALL AGCM (°C). (c) as (a), but for the SSH (cm) response. (d) SSH response in an LCS
519 model experiment forced by PTUBO-CTL wind stress anomalies (cm). 90% significance
520 level in black contours (dashed when negative values), except in (a) in which red/green
521 contours show positive/negative relative SST (°C). Windstress overlaid in black when
522 90% significant, grey otherwise.

523

524 **Figure 2. Main mechanisms driving the IO response.**

525 Equatorial IO zonal wind stress (upper row, N.m⁻²) and eastern equatorial IO SSH response in the LCS (middle row, cm)
526 for **(left)** CPL (black) and ALL (purple) experiments and for **(right)** LAND (cyan),
527 LAND_Africa (green), OCEAN (dark blue) and ATM (red) sensitivity experiments. Circles
528 indicate when anomalies are significant at the 90% level. Light color shading indicate
529 the 90% confidence interval for selected experiments. Zonal wind stress is averaged
530 over [60°E-90°E ; 2.5°N-2.5°S] region and SSH over the [90°-110°E ; 0°-10°S] region.

531

532 **Fig. 3. Influence of background IO conditions on the IPSL CM5B response to the**

533 **eruption.** (a) SST DMI anomaly (with regards to historical run climatology, in °C) for the
534 control ensemble members (black) and the Pinatubo simulations (purple) for members
535 showing neutral IOD conditions in SON 1991 in the control simulations (49 members).
536 (b) Same for positive IOD conditions (21 members). (c) Same for negative IOD
537 conditions (20 members). d, e, f: same for DMI east pole SSH. 90% intervals of
538 confidence in light shading.

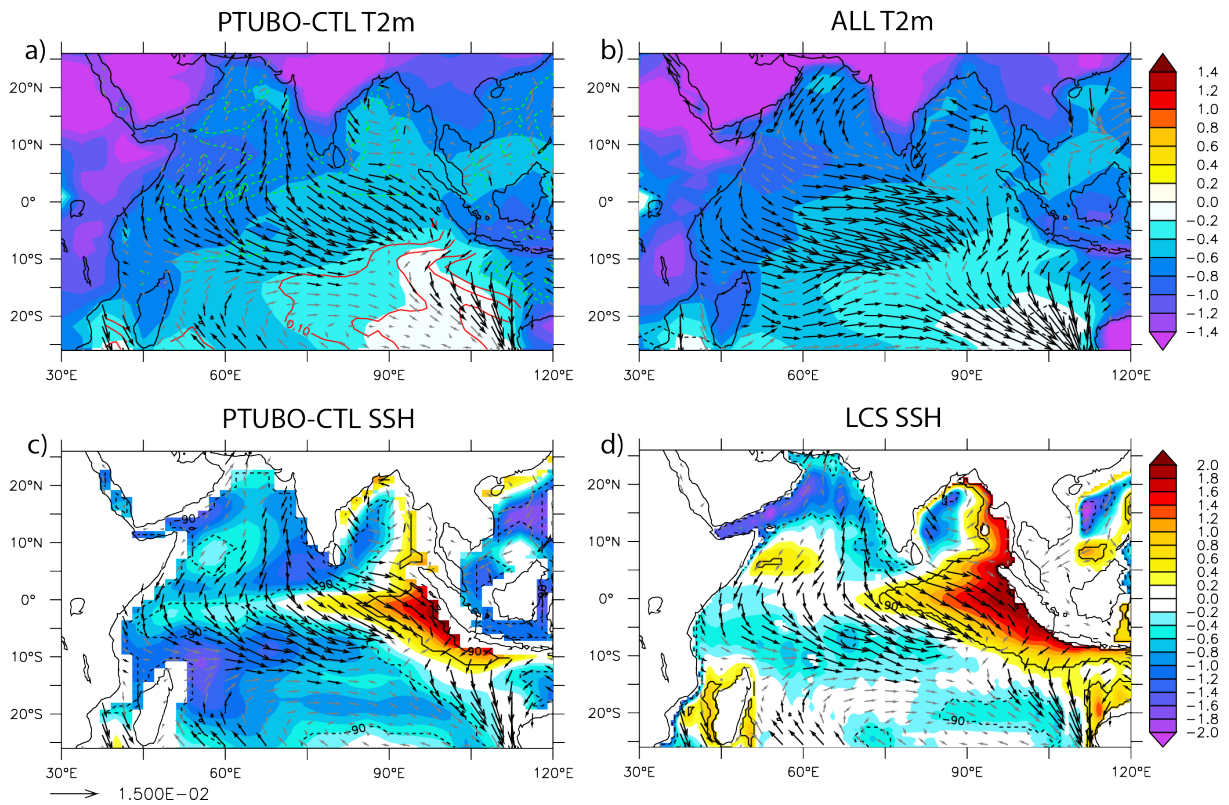
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540 **Fig. 4. IO response in the autumn following tropical volcanic eruptions in CMIP**

541 **models.** CMIP response to the last five large tropical eruptions in SON of eruption's year
542 for (a) 2m air temperature (°C, positive/negative relative SST in red/green contours),
543 (b) SSH (cm) and (c) Primary production (PP, unit : 10⁻⁸ mol m⁻² s⁻¹). Wind stress are
544 overlaid in black when 90% significant, grey otherwise. 90% significance level in black
545 contours (dashed when negative values), except for PP, whis is available for less models,

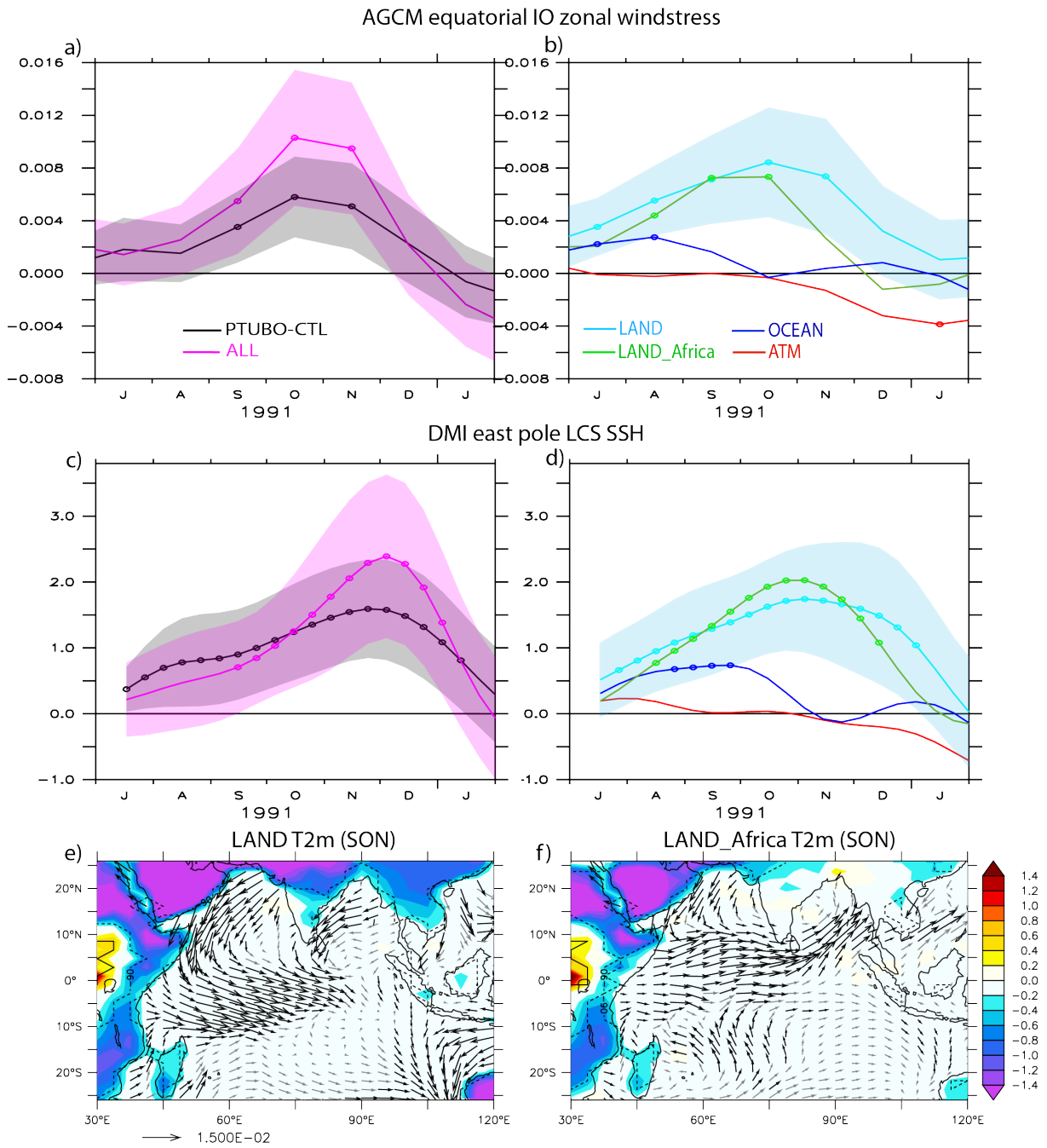
546 for which significancy is defined when there are two times more models of the adequate
547 sign by solid color shading (white-stripped when not significant).

Figures 1 to 4 with captions.



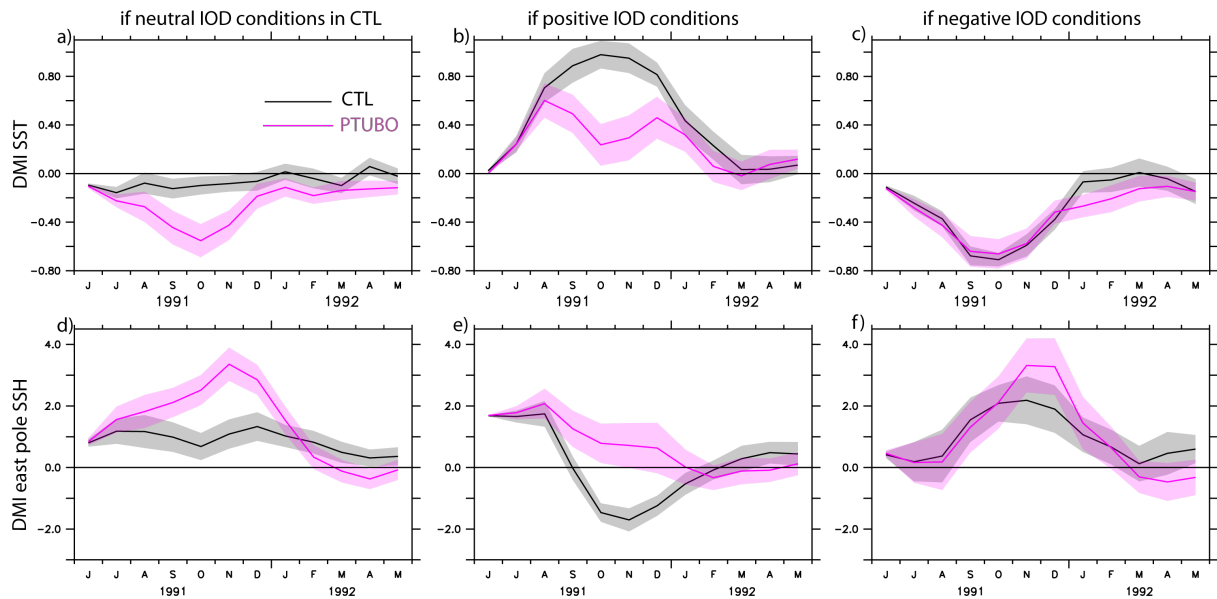
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 9 contours show positive/negative relative SST ($^{\circ}\text{C}$). Windstress overlaid in black when
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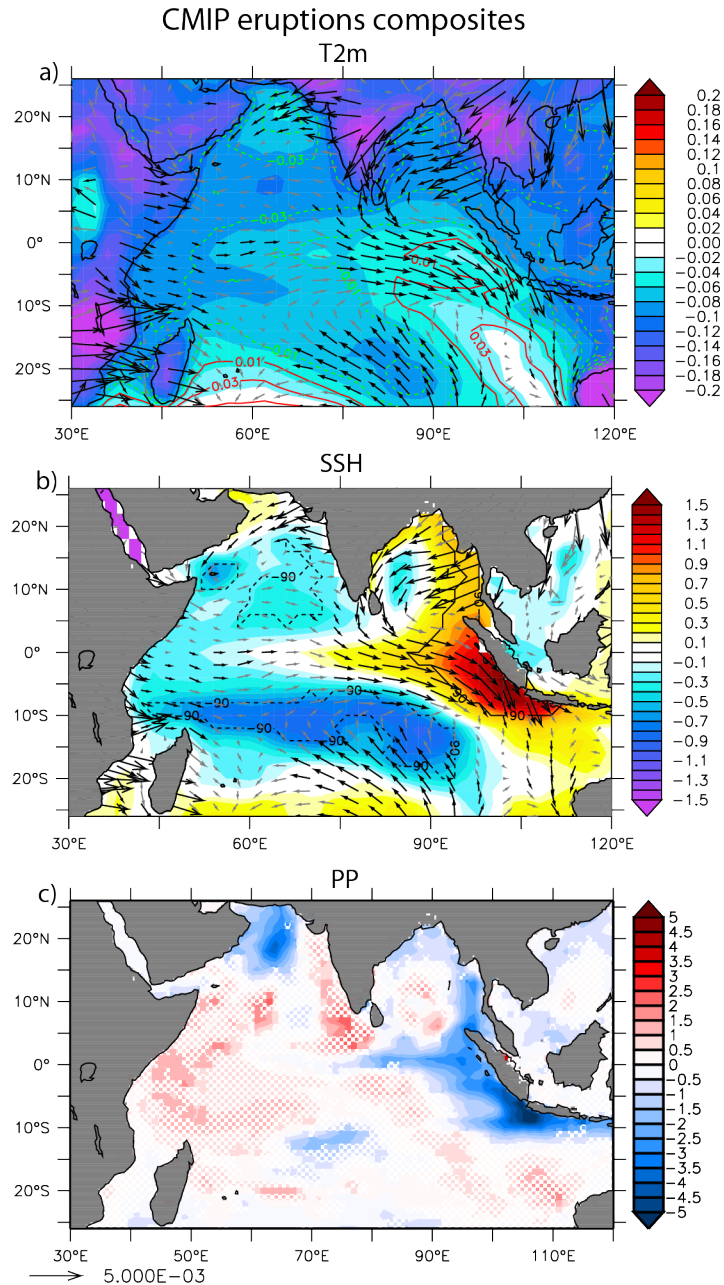
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12 **Figure 2. Main mechanisms driving the IO response.** Equatorial IO zonal wind stress
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 18 over $[60^{\circ}E-90^{\circ}E ; 2.5^{\circ}N-2.5^{\circ}S]$ region and SSH over the $[90^{\circ}-110^{\circ}E ; 0^{\circ}-10^{\circ}S]$ region.



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Fig. 3. Influence of background IO conditions on the IPSL CM5B response to the eruption. (a) SST DMI anomaly (with regards to historical run climatology, in °C) for the control ensemble members (black) and the Pinatubo simulations (purple) for members showing neutral IOD conditions in SON 1991 in the control simulations (49 members). (b) Same for positive IOD conditions (21 members). (c) Same for negative IOD conditions (20 members). d, e, f: same for DMI east pole SSH. 90% intervals of confidence in light shading.



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Fig. 4. IO response in the autumn following tropical volcanic eruptions in CMIP models. CMIP response to the last five large tropical eruptions in SON of eruption's year for (a) 2m air temperature ($^{\circ}\text{C}$, positive/negative relative SST in red/green contours), (b) SSH (cm) and (c) Primary production (PP, unit: $10^{-8} \text{ mol m}^{-2} \text{ s}^{-1}$). Wind stress are overlaid in black when 90% significant, grey otherwise. 90% significance level in black contours (dashed when negative values), except for PP, whis is available for less models, for which signficancy is defined when there are two times more models of the adequate sign by solid color shading (white-stripped when not significant).

Supplementary Information for “A subsurface Indian Ocean Dipole response to tropical volcanic eruptions”

Takeshi Izumo, Myriam Khodri, Matthieu Lengaigne, Iyyappan Suresh

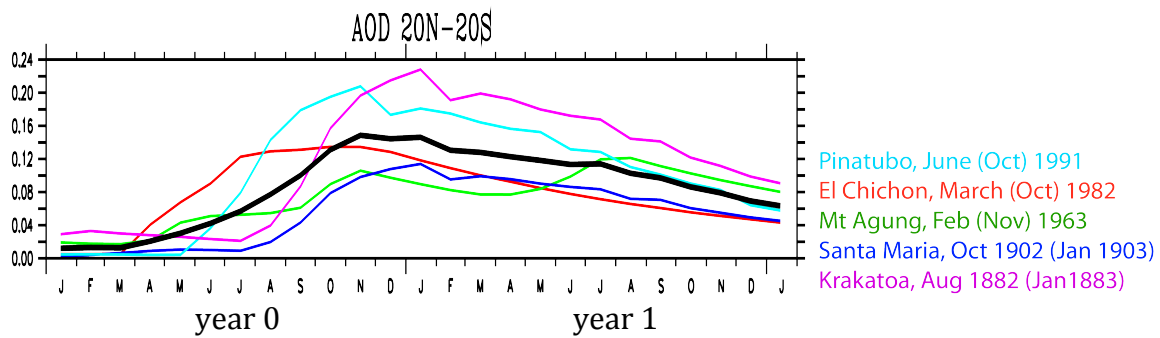


Figure S1. Tropical average (20°N-20°S) of Stratospheric Aerosol Optical Depth (AOD ; Gao et al. 2008) for the last 5 tropical stratospheric eruptions from the historical period, from year 0 (year of the eruption) to year 1 (eruption’s following year). The starting (peaking) month and year for each eruption is indicated on the right-hand side. The black line indicates the mean of the 5 eruptions. This figure illustrates the large diversity in tropical AOD amplitude, seasonality and temporal evolution, and the fact that the Pinatubo eruption had the largest radiative forcing in summer-fall, the season that matters for the development of the subsurface IOD studied here.

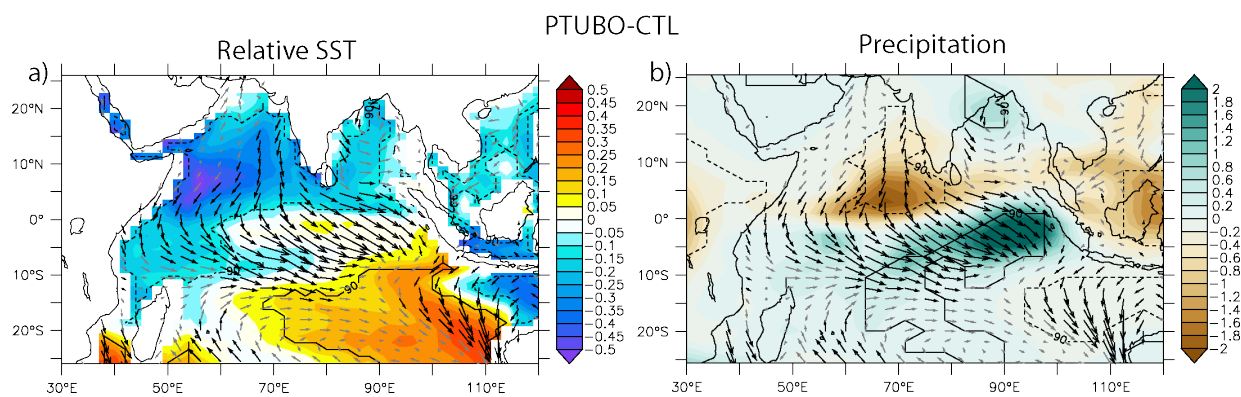


Figure S2. As Fig. 1a (difference PTUBO-CTL), but for (a) relative SST (°C) and (b) precipitation (mm/day).

Regression onto -DMI in SON, IPSL-CM5B-LR historical run

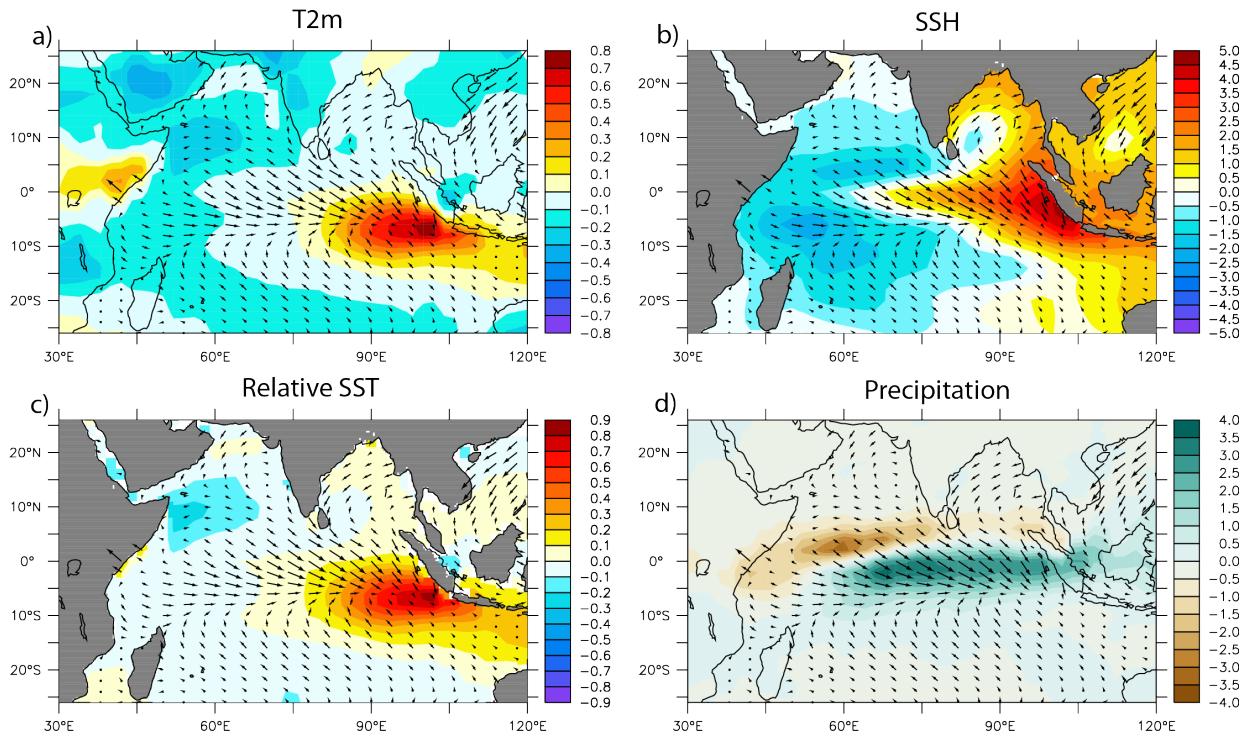


Figure S3. Typical IOD pattern in IPSL-CM5B-LR. Regression of (a) 2m air temperature ($^{\circ}\text{C}$), (b) SSH (cm), (c) relative SST ($^{\circ}\text{C}$) and (d) precipitation (mm/day) onto normalised DMI in SON (its standard deviation being 0.55°C), in IPSL-CM5B-LR CMIP5 historical run.

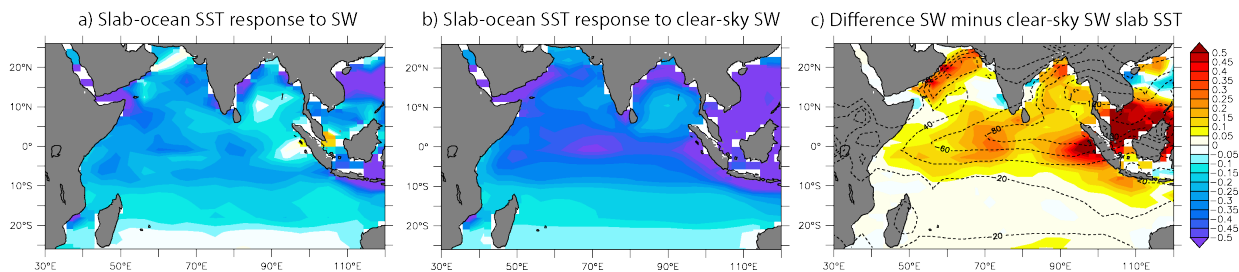


Figure S4. Slab ocean response to PTUBO-CTL SW forcing in July-September (JAS, when OCEAN has a significant effect on equatorial zonal wind stress; cf. Fig. 2). (a) Slab-ocean SST response to PTUBO-CTL SW forcing ($^{\circ}\text{C}$). (b) Slab-ocean SST response to clear-sky PTUBO-CTL SW forcing. (c) Difference between the two, SW minus clear-sky SW, i.e. mainly the effect of climatological cloud cover shown in contours (effect inferred from IPSL-CM5B-LR climatological SW minus climatological clear-sky SW; $\text{W}\cdot\text{m}^{-2}$). See details about this slab ocean using a spatially-varying MLD climatology in supplementary section 3. Note that, while the cooling and its heterogeneity is mainly due to the SW anomalies, there is a notable additional effect of latent heat flux – through windspeed anomalies – that could explain the relative SST warming seen south of 10°S in the southeast Indian Ocean, west of Australia, in Fig. S2a and Fig. 1a. The northwest anomalies there (possibly related to the anticyclonic anomalies created by the strong land cooling of Australia; not shown) oppose the trade winds, thereby decreasing local windspeed and evaporation. A similar windspeed effect, but of opposite sign, seems to have also played a role in reinforcing the Arabian Sea cooling (southwesterly anomalies reinforcing the monsoon low-level jet in JAS; not shown).

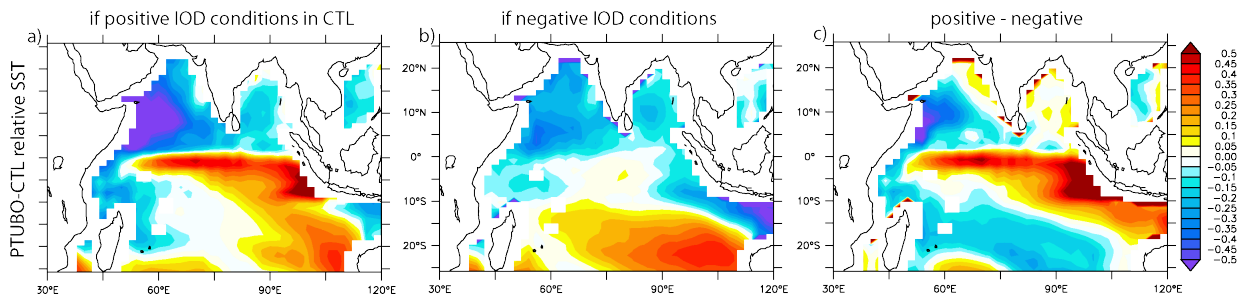


Figure S5. Influence of background IO conditions on the IPSL CM5B relative SST response. (a) As Fig. S2a, but for members showing positive IOD conditions in SON 1991 in the control simulations. (b) same for negative IOD conditions. (c) difference between panels b and a. This figure highlights the asymmetries in the SST response, with a much larger negative Northwest-Southeast SST gradient in the background positive IOD case, which projects onto the DMI poles (cf. Fig. 3a,b,c).

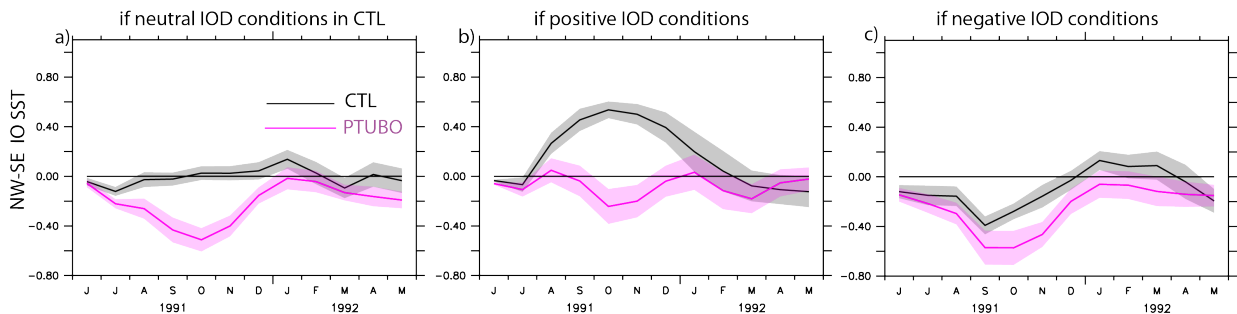


Figure S6. Influence of background IO conditions on the IPSL CM5B NW-SE SST gradient response. (a) As Fig. 3abc, but for an index especially developed to capture the Northwest – Southeast (NW-SE) SST gradient forced by the eruption, defined as the SST difference between the 50-70°E, 20°N-0°N and 90-110°E, 0-20°S regions. The PTUBO-CTL difference is even larger than for the DMI shown in Fig. 3abc, and is still significant in the negative IOD background case.

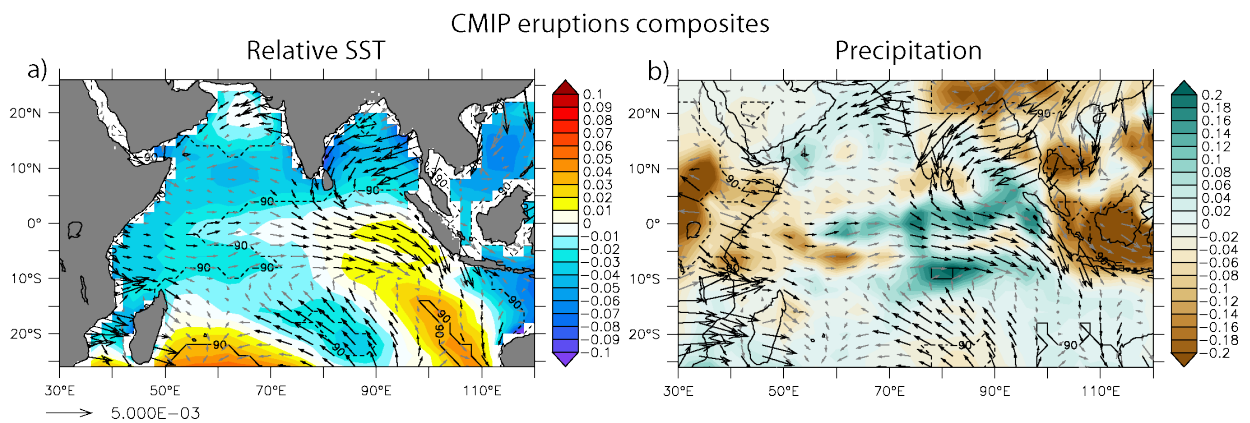


Figure S7. As Fig. 4 (CMIP eruptions composites), but for (a) relative SST ($^{\circ}\text{C}$) and (b) precipitation (mm/day).

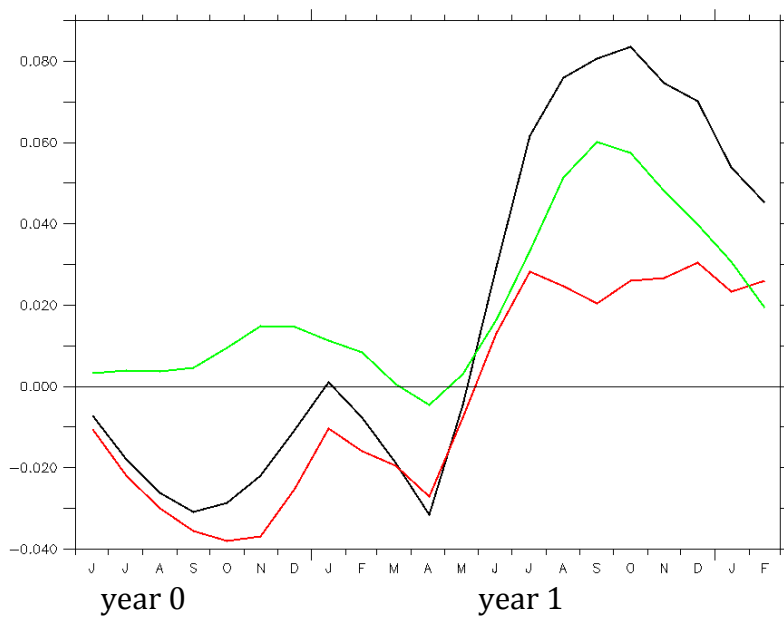


Fig. S8: DMI temporal evolution from year 0 to year 1 in CMIP MME (black), external influence of ENSO on the DMI (green; inferred from the linear regression of DMI on Nino3.4 SST computed for each CMIP model separately), and the DMI response independent of simultaneous ENSO conditions (red = black - green).

1. Inhomogeneous cooling, relative SST and rainfall responses to eruptions

To understand the rainfall response, as the eruption induces an overall cooling, especially in the tropics, it is relevant to use the relative SST, simply defined as the SST minus its tropical mean (20°N-20°S), a concept initially developed for global warming (Held and Soden 2006, Vecchi and Soden 2007), but which is also relevant here, as the volcanic aerosol optical depth radiative forcing is rather homogeneous zonally along the tropical band and evolves relatively slowly with regards to tropospheric tropical adjustment time (see also K17, which shows the relevance of using relative SST to understand the El Niño response to tropical eruptions).

Concerning the physical process behind the faster land cooling than ocean one, it is likely related to the larger lapse rate over the land, as the troposphere over the land generally is drier than that over the oceans (as suggested by global warming studies, e.g. Joshi et al., 2008), rather than to the lower heat capacity of the land (as suggested by some earlier studies for volcanism). The upper tropospheric cooling is relatively uniform in the tropics: a larger lapse rate hence yields a larger surface cooling over land, in particular in dry areas (same suggested mechanism, for a cooling instead of a warming, as in the case of global warming discussed in e.g. Sobel et al., 2001; Joshi et al. 2008).

2 The IPSL-CM5B-LR model and the AGCM experimental strategy (detailed description)

In the present study, we use the same set of simulations as in K17 including experiments with the Institut Pierre Simon Laplace (IPSL) CM5B-LR coupled climate model, atmospheric and oceanic sensitivity ensemble experiments, with and without a Pinatubo-like forcing, to identify the main mechanisms at play. We choose the Pinatubo eruption (June 15th, 1991), as it is the largest and best-documented tropical stratospheric eruption of the instrumental period. The details of these experiments are already described in K17 and we only summarize here their salient features. The atmospheric component is LMDZ5B (Hourdin et al. 2013) with 39 pressure levels (including 18 in the stratosphere) and a horizontal resolution of 1.87° in latitude and 3.75° in longitude. It uses ORCHIDEE as its land surface component. It is coupled to the NEMO3.2 oceanic component (Madec 2015) (31 vertical levels, and an horizontal resolution of about 2°, with a refinement of the latitudinal resolution to 0.5° near the equator; ORCA2 grid). This model is able to capture reasonably well the main observed IOD features (Jourdain et al. 2016).

The above IPSL-CM5B-LR climate model is forced by Pinatubo stratospheric aerosol optical depth (AOD) data provided by Gao et al. (2008). The 30-member ensemble experiment is run from 1st of June of a selected year from the historical run, chosen to be close to 1990s greenhouse gases and tropospheric aerosols concentrations, to lie more than 5 years away from any volcanic eruption and to lead to neutral El Niño Southern Oscillation (ENSO) conditions in the absence of volcanic forcing, avoiding any interference with external ENSO forcing (K17). For this initial state, 30 members were generated by adding a small white noise on the initial SST. For Fig. 3, we used two others 30-member ensemble experiments starting from the 1st June of two other selected years, these two years being characterized respectively by El Niño and La Niña conditions in the Pacific.

Corresponding control ensembles starting from the same initial conditions but without Pinatubo volcanic forcing were analysed (i.e. 3 CTL ensembles of 30 members each).

K17 have developed targeted ensemble experiments using the atmospheric component of the IPSL-CM5B model forced by the outputs of the coupled model ensemble (see K17 for details) to investigate the respective role of the following three processes in the generation of atmospheric anomalies, notably windstress :

(1) the direct effect of volcanic radiative forcing on clouds and atmospheric vapour content (Soden et al. 2002), potentially altering the atmospheric vertical structure and inducing an atmospheric dynamical response (referred to as ATM). Volcanic aerosol forcing is included in ATM, but the surface albedo is modified so that continental surfaces do not cool in response to volcanic forcing in this experiment.

(2) the indirect effects of volcanic forcing through the SST cooling and its spatial heterogeneities, and related horizontal SST gradients (OCEAN; inferred from the difference between the AGCM forced by the SST from PTUBO experiment and the unperturbed AGCM experiment, cf. K17). The OCEAN experiment includes not only the effect of SST anomalies that develop in response to volcanic forcing, i.e. the initial thermodynamical forcing and the slab ocean response, but also the positive Bjerknes feedback (i.e. the LCS dynamical response) once nIOD-like relative SST anomalies have started developing. However, the dominance of the LAND process (cf. Fig. 2) suggests that coupling processes, notably the Bjerknes feedback, should only be of secondary importance during the summer-fall of the eruption year, at least in our experimental framework.

(3) continents cool faster than the ocean in response to volcanic forcing. The third and last experiment (referred to as LAND) tests the effect of this differential surface cooling between land and ocean. In LAND, there is no prescribed aerosol forcing as in OCEAN, but the land surface albedo modification enforces a land surface cooling that is consistent with that of the IPSL-CM5B coupled model experiment.

This strategy allows land surface temperature variations in OCEAN experiments, in response to SST anomalies, as described in the context of global warming (Compo and Sadershmukh 2009, Dommenges 2009). It is hence more physically consistent than experiments that directly constrain land surface temperature (Ohba et al. 2013), which can result in an exaggerated land cooling and land-ocean gradients. An additional sensitivity experiment from K17, of interest here, explores the effect of the faster land cooling only over one specific region, Africa (LAND-Africa). An additional experiment (ALL) that includes all three LAND, OCEAN and ATM processes reproduces reasonably well the anomalies simulated by the IPSL-CM5B ensemble simulation (Fig. 1a,b and Fig. 2a,c), confirming the relevance of this two-tier approach for understanding how volcanic eruptions induce equatorial IO westerlies.

3 Linear Continuously Stratified (LCS) model and slab ocean model

The equatorial Indian Ocean dynamical oceanic response to wind-stress anomalies resulting from each process is estimated separately by forcing a linear continuously stratified (LCS; McCreary, 1980) ocean model of the Indo-Pacific region (same configuration as in Izumo et al. 2016, 2018). This LCS model, that resolves vertical baroclinic modes (with mainly the first two modes contributing; not shown), realistically simulates SSH anomalies (see Suresh et al. 2013, 2016 ; Izumo et al. 2016, 2018; K17 for further details).

A simple slab ocean mixed layer equation is also used, to simulate the SST thermodynamical response to the eruption-induced ShortWave (SW) flux anomalies (see Supplementary Information), following the approach of McGregor and Timmerman (2011). This simple slab ocean mixed layer equation is based on the following linear SST equation to realistically resolve the variations of SST forced by Pinatubo shortwave anomalies (SW'):

$$\frac{dSST'}{dt} = SW' / (\rho C_p MLD) - \alpha SST'$$

the “prime” denoting the Pinatubo-forced anomalies (i.e. Pinatubo-forced IPSL PTUBO run minus control IPSL CTL run). SW is here the SW surface net downward flux minus the downward SW at the bottom of the mixed layer. Mixed Layer Depth (MLD) is the local MLD monthly climatology, inferred from the depth at which the potential density is 0.06kg/m³ larger than at 10m depth (equivalent to a 0.2°C decrease in temperature) of the CTL IPSL run. The forcing is simply from this SW', other heat fluxes being parameterised as a constant Newtonian damping $-\alpha SST'$, with the typical damping timescale $1/\alpha = 2$ months, as in Izumo et al. (2016). The time integration (with initial start in June 1991) of the above linearized SST tendency equation gives us the SST' temporal evolution. A reference simulation is set up using the MLD climatology and the SW' anomalies induced by Pinatubo in IPSL ensemble mean. In addition to the clear-sky sensitivity experiment (in which the clear-sky SW forcing is rather zonally-homogeneous in the tropical band) shown in Figure S4, we used this very simple model to evaluate the roles of the zonal variations of SW' (mainly related to climatological cloud cover/albedo spatial variations) and of mean MLD zonal variations on the SST response to Pinatubo, and so on SST zonal gradient. While the former explains most of the actual gradient, the latter does not: the zonal variations of the MLD climatology have less impact on equatorial SST gradient (not shown) and generally oppose those from climatological cloud cover distribution (MLD, being shallower in the east, will tend to increase cooling there).

4 CMIP5 simulations

To assess the robustness of our results with our single model experiments, we also analyse historical simulations from the Climate Model Intercomparison Project phase 5 (CMIP5) dataset (Taylor et al. 2012) over the 1861-2005 period (see Suppl. Table S1). To estimate their responses to tropical volcanism, we composited the 5 largest tropical eruptions (cf. K17 for details). 49 models were available for 2m and surface temperature, 44 for windstress, 37 for SSH and 18 for Primary Productivity. We used all models available for each field. The Multi-Model-Ensemble (MME) mean is computed by first

averaging all members for each model to smooth out internal noise before averaging between models. We checked that the results are qualitatively insensitive to the choice of these methods: using the first member of each model, or reducing the number of models, gives similar results, but with slightly more internal noise.

In preprocessing, we removed the 10 previous years sliding climatology in CMIP historical runs, to filter out possible trends and interdecadal variability. 3 months running mean (for maps) or Hanning filter (for time series) is applied on data to reduce the strong intraseasonal variability of the tropics. In order to better compare models, which possibly have different typical interannual variations, we also computed « standardised » amplitudes of some relevant indices, by normalizing by the interannual standard deviation computed for Sept.-Nov. (SON) over the full historical run period and for each model.

In addition to the main analyses shown in the present paper, we have also done a CMIP inter-eruption comparison. We have compared various eruptions in CMIP in terms of impacts on the IO. At least all the eruptions had a subsurface negative IOD in SON0, except for Krakatau with neutral IOD in autumn of yr0 (eruption in August, possibly too late to trigger a negative IOD; not shown), even if there is a large inter-model diversity because of the sensitivity to initial conditions (and so to internal variability of each model/member) that lowers the signal to noise ratio.

Table S1. List of the 48 CMIP models (158 members) used in this study, and the maximum number of members in each model (mainly for surface air temperature at 2m and sea level height).

Institute name	Model name	Max. number of members
BCC	bcc-csm1-1	3
	bcc-csm1-1-m	3
BNU	BNU-ESM	1
CCCma	CanESM2	5
CMCC	CMCC-CESM	1
	CMCC-CM	1
	CMCC-CMS	1
CNRM-CERFACS	CNRM-CM5	10
	CNRM-CM5-2	1
CSIRO-BOM	ACCESS1-0	1
	ACCESS1-3	3
CSIRO-QCCCE	CSIRO-Mk3-6-0	10
FIO	FIO-ESM	3
ICHEC	EC-EARTH	9
INM	inmcm4	1
IPSL	IPSL-CM5A-LR	6
	IPSL-CM5A-MR	3
	IPSL-CM5B-LR	1
LASG-CESS	FGOALS-g2	1
LASG-IAP	FGOALS-s2	1
MIROC	MIROC5	5
	MIROC-ESM	3
	MIROC-ESM-CHEM	1
MOHC	HadCM3	10
	HadGEM2-CC	3
	HadGEM2-ES	4
MPI-M	MPI-ESM-LR	3
	MPI-ESM-MR	3
	MPI-ESM-P	2
MRI	MRI-CGCM3	3
	MRI-ESM1	3
NASA-GISS	GISS-E2-H	6
	GISS-E2-H-CC	1
	GISS-E2-R	6
	GISS-E2-R-CC	1
NCAR	CCSM4	6
NCC	NorESM1-M	3
	NorESM1-ME	1
NIMR-KMA	HadGEM2-AO	1
NOAA-GFDL	GFDL-CM2p1	10
	GFDL-CM3	5
	GFDL-ESM2G	1
	GFDL-ESM2M	1
NSF-DOE-NCAR	CESM1-BGC	1
	CESM1-CAM5	3
	CESM1-FASTCHEM	3
	CESM1-WACCM	4
Total number of models/members	48	158