

The effect of soil moisture perturbations on Indian Monsoon Depressions in a numerical weather prediction model

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1 **The Effect of Soil Moisture Perturbations on Indian Monsoon Depressions**
2 **in a Numerical Weather Prediction Model**

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ABSTRACT

4 Indian monsoon depressions (MDs) are synoptic-scale cyclonic systems
5 that propagate across peninsular India three or four times per monsoon sea-
6 son. They are responsible for the majority of rainfall in agrarian north India,
7 thus constraining precipitation estimates is of high importance. Here, we use a
8 case study from August 2014 to explore the relationship between varying soil
9 moisture and the resulting track and structure of an incident MD using the Met
10 Office Unified Model. We use this case study with the view to increasing un-
11 derstanding of the general impact of soil moisture perturbations on monsoon
12 depressions. It is found that increasing soil moisture in the monsoon trough
13 region results in deeper inland penetration and a more developed structure
14 – e.g. a warmer core in the mid-troposphere and a stronger bimodal potential
15 vorticity core in the middle/lower troposphere – with more precipitation, and a
16 structure that in general more closely resembles that found in depressions over
17 the ocean, indicating that soil moisture may enhance the convective mecha-
18 nism that drives depressions over land. This experiment also shows that these
19 changes are most significant when the depression is deep, and negligible when
20 it is weakening. Increasing soil moisture in the sub-Himalayan *arable zone*,
21 a region with large irrigation coverage, also caused deeper inland penetration
22 and some feature enhancement in the upper troposphere but no significant
23 changes were found in the track heading or lower-tropospheric structure.

4 **1. Introduction**

5 Indian monsoon depressions (MDs) are synoptic scale systems that usually originate in the Bay
6 of Bengal and propagate northwestward across the Indian peninsula, with a mean duration of 4-6
7 days, and an average frequency of between two and four per summer (Boos et al. 2015; Hunt et al.
8 2016a). Their spin-up mechanism remains uncertain (Cohen and Boos 2016), although it appears
9 likely that convective instability of the second kind (CISK; Charney and Eliassen 1964) plays at
10 least some role (Shukla 1978); however, their primary propagation mechanism has been well de-
11 scribed, albeit fairly recently (Boos et al. 2015; Hunt and Parker 2016), as a coupling of horizontal
12 nonlinear advection of the mid-tropospheric potential vorticity maximum and an image vortex in-
13 teraction of the lower-tropospheric PV maximum with the no-normal flow condition imposed by
14 the Himalayas.

15 It also remains unclear what synoptic variables, if any, control the duration and ultimate dis-
16 sipation of MDs; there is some evidence that a contemporaneous monsoon flood year or active
17 spell tends to extend the duration of depressions in the north of the peninsula (Krishnamurthy and
18 Shukla 2007; Krishnamurthy and Ajayamohan 2010), although this has not yet been disentangled
19 into a primarily synoptic or mesoscale (troposphere or land surface conditions, respectively,
20 favourable for longer duration) theoretical framework. Nevertheless, recent work has shown that
21 favourable conditions (e.g. higher vorticity, more moisture) at both scales is correlated with in-
22 creased MD activity, duration, or intensity: e.g. for soil moisture by Chang et al. (2009); Kishtawal
23 et al. (2013), and for the active phase of the monsoon by Hunt et al. (2016a).

24 Eltahir (1998) was the first to provide a solid theoretical pathway to accompany the long-held
25 assertion that an increase in large-scale soil moisture induces enhanced precipitation. He proposed
26 that the drops in surface albedo and Bowen ratio caused by wetting soil work to increase the near-

27 surface specific moist static energy and boundary layer moist static energy gradient, which results
28 in more favourable conditions for precipitation. If, however, this is to be an important process in
29 MDs, it is likely to be indirect (it must also overcome a negative feedback at the MD centre – the
30 associated lower-tropospheric cold core (Godbole 1977; Hunt et al. 2016a) acts to cool the surface
31 and increase stability there): the area of maximum precipitation is found to the southwest of the
32 centre (e.g. Ramanathan and Ramakrishnan 1933) where the (adiabatic) quasigeostrophic omega
33 equation (e.g. Holton and Hakim 2012) predicts the greatest ascent associated with the balanced
34 MD vortex will be (Boos et al. 2015); in contrast the Bowen ratio tends to reach a minimum just
35 ahead (northwest) of the centre (Hunt et al. 2016a). To elucidate this, following Hunt et al. (2016a),
36 Fig. 1 shows the mean Bowen ratio (ERA-Interim; Dee et al. 2011) and precipitation (TRMM;
37 Kummerow et al. 1998; Huffman et al. 2010) for a 34-depression composite in which location and
38 orientation are normalised such that the centre lies at the origin and the heading is up the page;
39 land-only data were used. As asserted, there is not much spatial similarity between the extrema
40 of precipitation and Bowen ratio - indicating that if we are to believe previous work suggesting a
41 link between MD behaviour and underlying soil moisture, it may be a more subtle feedback, or
42 work on a finer spatial scale, than that suggested by Eltahir (1998). The caveat here is that surface
43 fluxes are an entirely modelled product in ERA-I, and so have substantial uncertainty; however
44 this is at least partially addressed by the similarity of composite MD precipitation between ERA-I
45 and TRMM, and the fact that most rainfall near the centre of a depression is stratiform in nature
46 (Hunt et al. 2016b). To date, a number of studies have shown that assimilation of soil moisture,
47 or better initial representation of it, improves the forecast of monsoon depressions in mesoscale
48 models (Chandrasekar et al. 2007; Vinod Kumar et al. 2007; Chandrasekar et al. 2008; Rajesh
49 and Pattnaik 2016). Further, it has been shown that inland soil moisture is capable not only of

50 extending the duration of tropical cyclones (Andersen and Shepherd 2017), but in some cases of
51 allowing them to re-intensify (Kellner et al. 2012).

52 Soil moisture is one of the meteorological variables subject to greatest change with respect to the
53 progression of the Indian monsoon, largely due to its correlation with accumulated precipitation.
54 The NOAA CPC reanalysis soil moisture climatology (Van den Dool et al. 2003) and the ESA CCI
55 satellite-derived soil moisture climatology (Liu et al. 2011, 2012; Wagner et al. 2012) for India for
56 April, June, August, and September are given in Fig. 2(a) and Fig. 2(b) respectively and show
57 a clear northwestward advance through most of the season: some areas in the monsoon trough
58 have September soil moisture more than double that of June. Naïvely, then, we might expect MD
59 tracks to penetrate deeper inland later into the monsoon season, given the expected influence of
60 antecedent soil moisture on the development of MDs. Fig. 3 shows the mean MD track for each
61 month (1979-2015) from the track datasets of Hunt et al. (2016a) and Hurley and Boos (2015)
62 respectively; note that the MD tracks have been extended to include parts where the depression is
63 strictly in a monsoon low regime (that is to say, the surface winds are below 8.5 m s^{-1}). There
64 is some weak evidence here to suggest that not only do MDs tend to progress further inland later
65 in the season, they also seem more likely to have over-land genesis. This should be taken with
66 the caveat that large-scale conditions over the subcontinent also clearly play some part, given that
67 there is evidence that the September tracks start to recede, despite high levels of soil moisture
68 remaining.

69 So, if soil moisture has some effect on the duration of MDs, which seems at least plausible,
70 we are then faced with with the secondary question of whether antecedent soil moisture patterns
71 could affect the heading of existing MDs. Chen et al. (2005) showed that, in theory, the off-
72 centre latent heat released by the asymmetric rainfall distribution would interact with the local
73 circulation to create a negative velocity potential southwest of the MD centre, and therefore there

74 would be some tendency for the MD to move in that direction. However, this mechanism is
75 unlikely to be the primary one, since depressions typically move towards the northwest, rather
76 than the southwest. Furthermore, Baisya et al. (2017) recently showed using a mesoscale model
77 that precipitation intensity in MDs is strongly coupled with antecedent soil moisture. Two simple
78 experiments are therefore proposed: firstly a uniform change in soil moisture across the monsoon
79 trough region to determine the sensitivity of MD duration to antecedent land surface conditions;
80 secondly a uniform change in soil moisture in the highly farmed region across the Himalayan
81 foothills (typically several hundred kilometres north of MD tracks; Roy et al. 2015) to determine
82 to what extent MDs can be steered by soil moisture. These questions are presented in the context
83 of an initial case study, but we hope that the results are sufficiently thought-provoking that further
84 research on this topic will be motivated.

85 We will discuss the experimental setup and outline the methodology in section 2, then outline
86 and interrogate the results, looking at contrasts in track and structure in section 3 before concluding
87 in section 4.

88 **2. The Met Office Unified Model and Experimental Setup**

89 *a. Overview and Case Study Selection*

90 The version of the Met Office Unified Model (hereafter, the UM) used for this study runs the
91 Global Atmosphere 6.0 scheme (GA6.0; Walters et al. 2015) at N768 resolution (~ 26 km) with 85
92 vertical levels over a global domain; the numerical scheme is semi-implicit and semi-Lagrangian
93 (Davies et al. 2005), and due to the resolution a number of subgrid processes are parameterised,
94 including convection (e.g. Gregory and Rowntree 1990, with additions).

95 In choosing an appropriate case study to use in this experiment, we were subject to two criteria:
96 firstly, and more importantly, that the MD happened within the last few years - this means that
97 higher resolution, better quality analyses are available for initialisation; secondly, that the MD had
98 a track resembling the average for MDs (see Fig. 3) that it could be seen as a fair representative of
99 the spectrum of MDs incident on the east coast of the peninsula. The most suitable such event was
100 the MD of early August 2014, which featured depression-status wind speeds from 200 km south
101 of Kolkata until it was downgraded to a monsoon low 400 km due south of Delhi. All experiments
102 were initialised at 00Z on August 3rd, the day this event was declared a monsoon depression.

103 *b. The Land Surface Scheme and Parameterisation*

104 The operational land surface model in the Met Office UM is the Joint UK Land Environment
105 Simulator (JULES; Best et al. 2011). This employs the Met Office Surface Exchanges Scheme
106 (MOSES; Cox et al. 1999; Essery et al. 2003) to handle hydrological processes both subterranean
107 and in the boundary layer. A brief description of the governing equations in the soil hydrology
108 subroutine, which is taken from the relevant part of the MOSES documentation, is given in the
109 Appendix. The interaction between clouds and shortwave/longwave radiation is also handled ex-
110 plicitly by the prognostic cloud scheme in the UM (PC2; Wilson et al. 2008) following Edwards
111 and Slingo (1996).

112 *c. Ensemble Generation*

113 There are two types of stochastic perturbation that can be employed to generate a spread of
114 forecasts in a numerical weather prediction model: uncertainties in the analysis can be represented
115 by perturbing the initial conditions, whereas uncertainties in the model can be represented by using
116 any number of physics perturbations (e.g. time-varying parameterisations). Operationally, the Met

117 Office use The Met Office Global and Regional Ensemble Prediction System (MOGREPS; Bowler
118 et al. 2008) to generate ensemble NWP runs; given that this was designed specifically for the UM,
119 we aim to make our ensemble generation as similar as possible. MOGREPS uses two distinct
120 stochastic physics schemes: random parameters (RP) and stochastic kinetic energy backscatter
121 (SKEB). The former uses the premise that many parameters in the various parameterisations in
122 the UM are tuned to empirical values that appear to give the best representation of the relevant
123 process, these can be periodically varied at differing frequencies between physically reasonable
124 values to produce a spread of forecasts; the latter reintroduces kinetic energy lost through poor
125 representation of the mechanisms by which small-scale processes cascade energy to larger scales
126 (Shutts 2005). Initial tests suggested that using SKEB perturbations tended to artificially weaken
127 MDs and cause them to have much shorter tracks. Thus in our study we used a stochastic perturbed
128 tendencies (SPT) scheme which simply randomly perturbs the summation of tendencies from all
129 parameterisations in the model (Buizza et al. 1999).

130 In our ensemble, we must also attempt to represent uncertainties in the analyses that are used
131 to initialise the model. In MOGREPS this is typically done by applying an ensemble transform
132 Kalman filter (ETKF; Bishop et al. 2001) to a previous ensemble run, assimilating observations to
133 assess where perturbations will have the largest impact. As operational ensemble analyses were
134 not readily available for our case study, we opted to simulate the uncertainty by adding white
135 noise of amplitude 0.5 K to boundary layer potential temperature. Sensitivity tests determined
136 that this gave a realistic spread of MD tracks from a short initialisation without suppressing the
137 development and progression of the depression. For each sub-experiment, which are differentiated
138 by varying soil moisture in the same region, a ten-member ensemble was used; for each ensemble
139 member, a random seed was used such that across each experiment each ensemble was generated
140 via the same set of pseudorandom parameters to allow intercomparability.

141 *d. Soil Moisture Ancillaries*

142 As discussed in the Introduction, two case study experiments are proposed to explore the sen-
143 sitivity of duration and heading respectively to underlying soil moisture. Fig. 4 shows the masks
144 used to set up the soil moisture ancillary files: the red polygon covers much of South Asia, the
145 green polygon covers the typical monsoon trough region, and the orange covers the sub-Himalayan
146 arable land that is becoming increasingly intensively irrigated and farmed. In each instance, the
147 soil moisture control (perturbations to which will be used in the experiments) is the August clima-
148 tology as computed from a fully coupled high-resolution climate simulation in the UM. This was
149 chosen to reduce spin-up/resolution issues that could be introduced by using a climatology from,
150 e.g., either of the datasets in Fig. 2. This is the current method used for soil moisture initialisation
151 of the MetUM in operational NWP mode.

152 For the first experiment (hereafter: *trough zone*), soil moisture in the monsoon trough region
153 (the green polygon in Fig. 4) - in which MD tracks are typically entirely embedded - was altered
154 to 1%, 80%, 100% (control), 120%, and 500% of its August climatological value. The 500%
155 value unsurprisingly gives significant oversaturation across much of the region, where this was
156 the case, soil moisture values at these locations were set to their saturation values; in reality, this
157 scaling is achievable only over the dry northwest, and the average saturation value over the trough
158 region is approximately 167%. Conversely, for the second experiment (hereafter: *arable zone*),
159 soil moisture over South Asia (the red polygon in Fig. 4) is set to 1% of its August climatological
160 value, except for inside the arable sub-Himalayan area (orange polygon) where the values were
161 set to 1%, 50%, 100%, and 500% of the climatology. This region was traced to resemble, as much
162 as possible, the belt of sub-Himalayan arable grassland where irrigation is becoming rapidly and
163 increasingly prevalent (Roy et al. 2015) - the area where anthropogenic changes to the surface

164 are likely to have the biggest impact. Values of soil moisture approaching 1% of the August
165 climatology could be found in an *extremely* dry pre-monsoon period, but we remind the reader
166 that the purpose of this experiment is to test the effect of soil moisture contrast in the region, not
167 necessarily to replicate a physical event.

168 *e. Tracking*

169 The tracking algorithm used to determine the trajectories of MDs in output data is an updated
170 and extended version of that described in Hunt et al. (2016a). Data at individual timesteps in
171 the output are filtered subject to the IMD criteria for MDs (minimum 8.5 m s^{-1} surface wind
172 speed and two closed surface isobars at even hPa values) as well as some transient-filtering criteria
173 (lower-tropospheric vorticity above $3 \times 10^{-5} \text{ s}^{-1}$, smoothed MSLP must be local minimum), and
174 single-timepoint candidates are linked together using a simple nearest-neighbour algorithm.

175 **3. Results**

176 *a. Tracks*

177 Tracking results from the *trough zone* experiment are shown in Fig. 5(a). The average tracks for
178 each sub-experiment (thick, coloured lines) were computed using normalised track durations for
179 each of the 10 ensemble members; that is to say points were grouped and averaged by total MD
180 lifetime fraction rather than absolute time since genesis, with termination points for all ensem-
181 ble members across the experiment given by crosses of the relevant colour. The pale green area
182 underneath is a concave hull of all points of all ensemble tracks from the control sub-experiment
183 (i.e. underlying soil moisture set at 100% of the August climatology). The official IMD track for
184 the event is also given in black for illustration.

185 A first inspection of the average tracks seems to suggest that an increase in underlying antecedent
186 soil moisture results in deeper penetration of MDs through the monsoon trough region - this is vis-
187 ible both in the average termination points and the individual ones. Further inspection indicates
188 that both the 500% and 120%, and 100% and 80% average tracks are closely matched pairs, both
189 along track and at termination. The former couple is a result of the August soil climatology already
190 being fairly close to saturation in this region, so the difference between 20% extra moisture and
191 saturation is fairly small. Performing Hotelling's t^2 -test (Hotelling 1992) – the multidimensional
192 generalisation of the standard student's t-test for determining whether data are significantly differ-
193 ent from each other (we have also applied Welch's generalisation to allow for unequal variance in
194 the two comparison populations (Welch 1947)) – to assess whether the sub-experiment ensemble
195 terminations are distinct from each other, we find that all pairs apart from the aforementioned two
196 are significantly different from each other at the 95% confidence level. This leads us to conclude
197 there is a likely causal relationship between large-scale antecedent soil moisture in the monsoon
198 trough region, and the duration/distance travelled by incident monsoon depressions. So, is this
199 deeper penetration due to faster inland propagation or a longer duration? Using the ensembles, we
200 can compute the mean speeds and durations for the 1%, 80%, 100%, 120%, and 500% ensembles,
201 the mean propagation speeds are: 3.7, 3.7, 3.7, 3.9 and 3.9 m s^{-1} respectively, with corresponding
202 mean durations of 3.7, 4.3, 4.4, 4.2, and 4.3 days. Applying a significance test, we find that the
203 mean ensemble speeds for the two wettest cases (500% and 120%) are significantly different from
204 the drier ones, and that the mean duration for the driest case (1%) is significantly different from
205 the four wetter ones.

206 The *arable zone* experiment was set up to determine to what extent moisture changes in relatively
207 distant soil could affect the steering of a contemporaneous MD. Recall that for this experiment,
208 the soil moisture over South Asia was set to 1% of the climatology, and to the value specified (1%,

209 50%, 100%, or 500%) of the climatology in the sub-Himalayan belt. The results from this exper-
210 iment are presented in Fig. 5(b) in an identical fashion to those from the *trough zone* experiment.
211 In the absence of a control run, the concave hull given is for the “100%” ensemble plume. While it
212 may seem contrived to have such extremely dry soil over almost the entire peninsula for the sake of
213 establishing a strong contrast for our experiment, these desiccated conditions are not particularly
214 uncommon in the pre-onset conditions of late May (Fan and van den Dool 2004) where extreme
215 surface temperatures and scarce precipitation are usual, and depressions can still form in the Bay
216 of Bengal (Rao and Jayaraman 1958; Mooley 1980).

217 An initial overview of Fig. 5(b) suggests two broad characteristics: firstly, that the spread of en-
218 semble mean terminations is smaller than in the *trough zone* experiment - this is almost certainly
219 attributable to the altered soil area both having a smaller area and being further away, and thus
220 being less influential; secondly, that all the average tracks are shorter than in the previous exper-
221 iment - plausibly due to a larger area of desiccation than in the 1% *trough zone* sub-experiment
222 resulting in even less water being available over the peninsula, bearing in mind that MDs draw
223 moisture in from distances of up to 1000 km (Hunt et al. 2016a). We also note that whilst there is
224 a perfect rank correlation between soil moisture fractional change and mean termination latitude,
225 the mean track for the 100% sub-experiment is longer than that for the 500% ensemble. Repeating
226 the termination point significance analysis carried out for the *trough zone* experiment, we find that
227 the three wettest sub-experiments have mean track termination points significantly different from
228 the driest (1%), but not from each other, at a 95% confidence level.

229 *b. Structure and evolution*

230 Having established that soil moisture changes, both local and distant, are capable of significantly
231 altering the track of a passing MD, we will now examine the differing synoptic structure that these

232 changes cause and attempt to bring the discussion to its conclusion. The largest contrast was seen
233 in the *trough zone* experiment, so we shall start the discussion there. Fig. 6 shows longitude-height
234 cross-sections through 500%-minus-1% composite variables from the *trough zone* experiment. We
235 will briefly note here that structural changes of similar shape are found by comparing composites
236 arising from smaller changes in soil moisture, but with varying losses in magnitude, and hence,
237 significance. The centre of the MD (assuming one existed) at each timepoint across all ensemble
238 members for the relevant sub-experiment is centered at the origin; but unlike Fig. 1, we do not
239 rotate these composites since the soil moisture changes introduced were anisotropic. We note
240 that these differences are consistent across the other, non-extreme, experiments (not shown) albeit
241 with reduced areas of significance (typically more confined to the upper troposphere) and smaller
242 magnitudes.

243 We see that the composite MD for the wettest soil moisture case (in contrast to the driest) is more
244 intense, as the mid-tropospheric thermal high (Godbole 1977; Hurley and Boos 2015; Hunt et al.
245 2016a) is markedly stronger, with accompanied strengthening of both the 700 hPa and 500 hPa PV
246 maxima; secondarily there is evidence of an anomalous west-east circulation with enhanced ascent
247 ahead of the MD centre (i.e. to the west) with enhanced relative humidity there, and decreased
248 humidity and PV in the upper troposphere behind the centre; and, further, there is evidence of
249 increased westward axial tilt with height. We would expect these effects to be associated with
250 increased precipitation west of the centre, and we see in Fig. 7(a) that this is indeed the case.
251 Fig. 7 gives the 500%-minus-1% horizontal composite surface precipitation and 850 hPa wind
252 for both experiments. In the case of the *trough zone* experiment, we see, as expected from the
253 previous discussion, a substantial increase (beyond 40 mm day^{-1}) in precipitation downshear
254 (i.e. to the west) of the MD, with some slight reduction towards the east of the centre; however it is
255 not clear whether the increase in soil moisture enhances precipitation via the Eltahir mechanism,

256 or simply whether it allows more moisture to be inserted into the MD that then grows by other
257 means. The 850 hPa composite difference winds are also given in this figure; they indicate the
258 increased soil moisture sets up a large-scale, weak anomalous anti-cyclone that is split roughly
259 in half, noticeably intensifying the zonal components of the MD circulation near the centre, thus
260 making the core more cyclonic. This localised feature enhancement of the MD is very similar
261 to the behaviour over ocean (Hunt et al. 2016a) where features (particularly wind) tend to have
262 greater magnitude but smaller radial extent.

263 For comparison, the equivalent figure to Fig. 7(a) for the *arable zone* experiment is Fig. 7(b).
264 Here, the consequence of increased soil moisture is largely confined to the north of the MD as
265 expected, where a very weak anticyclone is established over the cold high associated with the
266 wetter ground; although the effect is weaker than in the *trough zone* experiment, there is still an
267 appreciable increase in the strength of the zonal circulation in the north quadrant of the MD. There
268 is little change to the precipitation, except for a slight increase in the north over the increased
269 soil moisture and a reduction in the west. On reflection, we should expect little difference to
270 the large-scale structure of the MD, but the strongest contrast is likely to be meridional given
271 the nature of our perturbation; therefore, we now consider some latitude-height cross-sections
272 for the 500%-minus-1% difference composites. These are given for potential vorticity, relative
273 humidity, and temperature in Fig. 8. It is clear (and unsurprising) that the effect of changing
274 *arable zone* soil moisture is felt substantially less by the MD than changing *trough zone* soil
275 moisture, since the *arable zone* soil moisture perturbation is some distance from the MD core.
276 The most prominent effect of wetting the soil there is to set up a wet, cool boundary layer; this,
277 in turn, acts to vertically extend the warm core of the MD while slightly reducing moisture in the
278 upper troposphere. Computation of mean CAPE (not shown) for each sub-experiment suggests a
279 slight increase around the centre with increasing soil moisture. There is no real evidence of this

280 apparent strengthening, however, in the precipitation or lower-tropospheric wind fields – the only
281 appreciable increase in magnitude is of the 700 hPa PV maximum.

282 It is also important to consider how varying soil moisture affects MDs as a function of their
283 lifetime. For example, one would suppose the impact to be quite minimal while most of the MD
284 is over the ocean. To test this, we can explore how selected fields from the trough experiment
285 ensemble sets vary as a function of depression lifetime (simply a normalised time axis: 0% is the
286 time of MD genesis, 100% is the time of MD lysis) - this is given for four fields in Fig. 9, in
287 which the colours red, yellow, green, and blue represent fractional changes to trough soil moisture
288 of 1%, 80%, 120%, and 500% respectively. Each field is computed over a box of side length
289 250 km centred on the MD centre. The topmost field in the figure is maximum CAPE found in
290 the quadrant of the aforementioned box that contains the next track point of the MD. There is a
291 marked region (roughly 40-70% through the MD lifetime) where the average maximum CAPE
292 in all sub-experiments is significantly higher than during the rest of the lifetime, and it is in this
293 region that a change in soil moisture has the strongest effect, with the extreme sub-experiments'
294 ensemble members almost having zero overlap. We also note that here, as well as in the other
295 fields, predictability is rapidly lost (i.e. the ensemble spread significantly widens) once the MD
296 starts to dissipate, and further that in this regime the effect of varying soil moisture becomes
297 negligible. In this particular instance, it is also true that during the spin-up phase of the MD, there
298 is no obvious correlation between increased soil moisture and enhanced CAPE. The reader's eye
299 may be drawn to this phase in particular both for its low CAPE and the fact that it continues to
300 drop in all cases before it hits land. Inspection of contemporaneous reanalyses suggests that this
301 system existed as a tropical low for a few days in the head of the Bay of Bengal (eroding CAPE),

⁰Delineated into NW, NE, SE, and SW; that is, if the MD is propagating WNW, CAPE is computed in the NW quadrant.

302 and – as can be seen from Fig. 5 – remained there for a little longer thereafter (eroding it further,
303 as seen in Fig. 9).

304 Related to CAPE, but not shown, is convective inhibition (CIN). Changes in soil moisture have
305 been shown to affect CIN (e.g. Clark and Arritt 1995), which typically reaches minimum mag-
306 nitude just ahead of the depression centre (Hunt et al. 2016a). Applying the same analysis that
307 we did for CAPE, we find that in the 1% case, CIN is significantly much more negative (less con-
308 ductive to convection) and that this extreme is much longer lasting in the vicinity of the centre when
309 compared to the other cases. The remaining cases did not differ significantly from each other.

310 Second from top in Fig. 9 is the mean total precipitable water in the area surrounding the MD
311 centre. This field is less variable than CAPE but still displays a clear maximum across all sub-
312 experiments at approximately 60% of the MD lifetime before rapidly falling away. As with max-
313 imum CAPE, there is significant correlation between trough soil moisture and mean total precip-
314 itable water as well as a significant difference between the values of the extreme sub-experiments
315 during the middle period where the MD is at its strongest, followed by a complete loss of corre-
316 lation, significance, and predictability after this point; although unlike CAPE, the correlation and
317 significance are retained during spin-up. Second from bottom is the mean lower/mid-tropospheric
318 temperature anomaly (averaged 850-400 hPa), here the picture is much the same as for total pre-
319 cipitable water, although the correlation is no longer significant at the 95% confidence level, and
320 the ensemble spread does not widen as much during lysis. Finally, at the bottom is maximum
321 relative vorticity in the lower troposphere (900-800 hPa); whilst this is an inherently variable field,
322 and consequently although there is arguably some correlation between it and soil moisture during
323 the period of maximum intensity, it is not significant, nor is the difference between the two ex-
324 treme sub-experiments significant more than occasionally. That having been said, any semblance
325 of correlation vanishes, as with the other fields, during the dissipation phase.

326 **4. Discussion and conclusion**

327 Monsoon depressions are responsible for the majority of the precipitation incident throughout
328 the summer across northern peninsular India and the monsoon trough region. Previous work has
329 established the possibility of at least a correlative connection between antecedent soil moisture
330 and the behaviour of incident MDs, but this is the first study to investigate the nature of that
331 relationship. Soil moisture, in two key areas where it has previously been identified as variable
332 and of meteorological importance, was varied through multiples of the climatology in a selected
333 NWP case study run in the Met Office Global Unified Model.

334 We have presented the results of a set of idealised sensitivity tests, each with multiple ensemble
335 members, initialised from the analysis of a typical depression chosen in August 2014. Whilst
336 we have framed these tests in the context of a single MD, significant differences have emerged
337 between the ensembles due to the imposition of soil moisture anomalies; we hope that this will
338 motivate further study of other events to explore the climatological relationship between MDs and
339 soil moisture.

340 We found that both the structure and propagation of the MD was significantly sensitive to
341 changes in soil moisture in the monsoon trough region: wetter conditions there caused a strength-
342 ening of the MD with increased central PV and a warmer thermal core, as well as a more pro-
343 nounced westward axial tilt. Such cases were also found to travel further inland before dissipating.
344 Further, we found that these changes were greatest (among variables associated with MD strength:
345 CAPE, TPW, mid-tropospheric temperature, and lower-tropospheric vorticity) during the period
346 when the MD is most intense, and that varying soil moisture has no noticeable effect on the MD
347 during its spin-down.

348 In the other experiment, soil across South Asia was kept desiccated while moisture in the sub-
349 Himalayan *arable zone* was varied. This had a lesser effect on both the structure and track of the
350 case study, although some significant differences persisted: tracks in the wetter cases terminated
351 later, and there was some weak strengthening of the MD in the middle and upper troposphere.

352 We also noted that in the wetter *trough zone* experiments, the ensemble composite MD became
353 more axially confined (as well as more intense), mimicking MD behaviour over the ocean (Hunt
354 et al. 2016a). This suggests that added soil moisture in this region provides more moisture to the
355 lower troposphere and subsequently enhances convective activity related to the MD. This is further
356 enhanced by increased lower-level convergence to the west of the centre.

357 This leaves us with several questions for further study. Firstly, how exactly does a monsoon
358 depression interact with the boundary layer? It has been indicated both here and in previous
359 work that MDs are very efficient at moving water from the surface through the PBL and into the
360 troposphere, despite not having particularly high wind speeds (by definition MDs lie at between 5
361 and 7 on the Beaufort Scale). This could be appropriately investigated by examination of a case
362 study in a mesoscale-resolution NWP model. Secondly, how would an incident MD respond to
363 horizontal gradients in soil moisture, rather than the block changes performed in this study; for
364 example with increasing (and decreasing) values both along track and across track? Thirdly, even
365 though we have spoken of CISK as the energy source for MDs, the precise role of CISK, and
366 its magnitude, remains uncertain. Uncovering the true MD spin-up mechanism would provide
367 invaluable direction for future research on the topic, and could be investigated using mechanism-
368 denial experiments in a suitable NWP framework (cf. Craig and Gray 1996).

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376 APPENDIX

377 **A1. Overview of the land surface scheme used in the model**

378 Four soil layers are used, for both the thermodynamic and hydrological subroutines, at depths
379 from the surface of 10, 25, 65, and 200 cm respectively; the prognostic total soil water in each
380 layer is given by:

$$M = \rho_w \Delta z \Theta_u \quad (\text{A1})$$

381 where ρ_w is the density of water, Δz is the thickness of the layer, and Θ_u is the liquid water
382 concentration (for the sake of this discussion, we neglect frozen water, although it is catered for in
383 the scheme). This is subject to the transport equation:

$$\frac{dM_n}{dt} = W_{n-1} - W_n - E_n, \quad (\text{A2})$$

384 where subscript n denotes the layer, W_n and W_{n-1} the diffusion terms in the layer and that immedi-
385 ately below it, and E_n is the evapotranspiration (including interaction with roots). The evapotran-
386 spiration function is controlled by land usage and vegetation data embedded in JULES, whereas
387 the diffusion terms are prescribed by the Darcy equation:

$$W = K \left(\frac{\partial \Psi}{\partial z} + 1 \right), \quad (\text{A3})$$

388 where K is the hydraulic conductivity and Ψ is the soil water suction function. Within MOSES
 389 these are respectively described by the Clapp-Hornberger relationships (Clapp and Hornberger
 390 1978):

$$\Psi = \Psi_s S_u^{-b} \quad (\text{A4})$$

$$K = K_s S_u^{2b+3}, \quad (\text{A5})$$

392 where Ψ_s , K_s and b are empirical constants that can be set on model initialisation. For this study,
 393 the default values used operationally by the Met Office were used.

394 There are then two boundary conditions: at the surface, the flux (aside from evaporation) is
 395 computed as the summation of canopy throughfall, snowmelt, and surface runoff; underneath the
 396 bottom (Nth) layer, the drainage (W_N) is set to equal the hydraulic conductivity.

397 Finally, the evaporation to the atmosphere from soil at the surface is given by:

$$E = \rho C_H U_1 [q_{\text{sat}}(T_\star, p_\star) - q_1] \left[f_a + (1 - f_a) \frac{g_s}{g_s + C_H U_1} \right] \quad (\text{A6})$$

398 where f_a is the tile saturation fraction (e.g. 1 for ice, lake, ocean, 0 for dry rock), ρ is the density
 399 of air, g_s is the surface soil conductivity, U is the wind speed, C_H is the surface flux heat exchange
 400 coefficient, q is specific humidity; and the subscripts \star , 1, and sat refer to the surface, lowest
 401 atmospheric model level, and saturation respectively.

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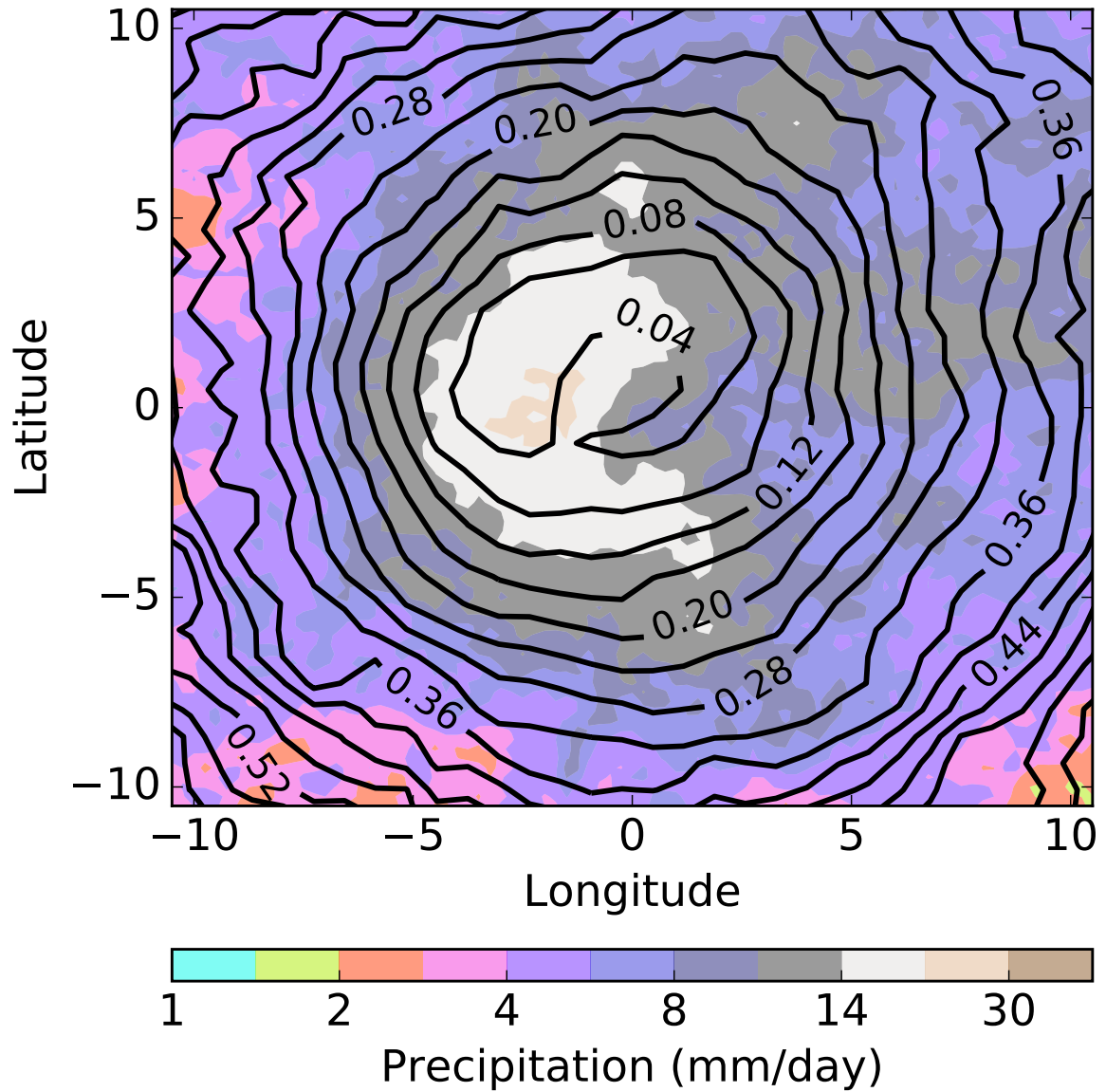
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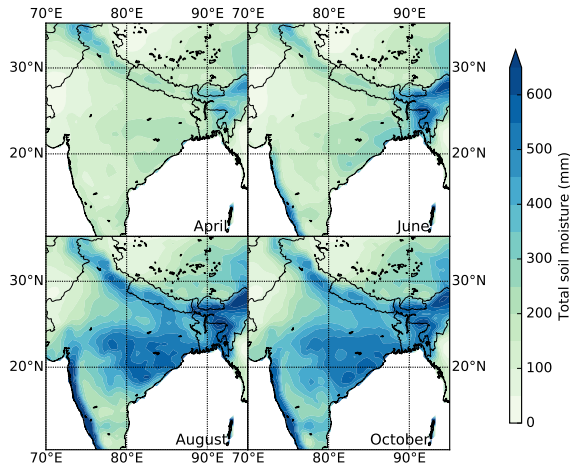
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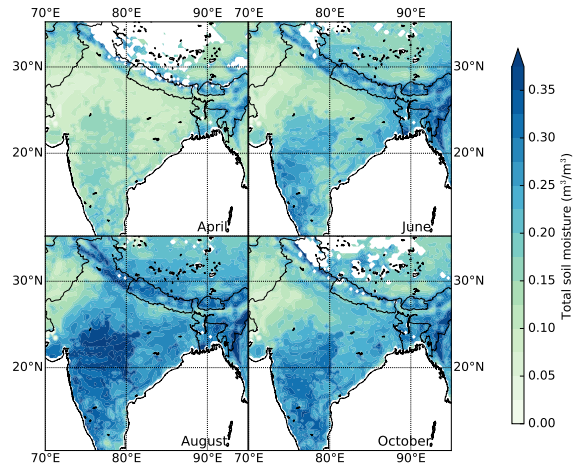
613 **Fig. 9.** Selected fields as a function of normalised depression lifetime for the trough experiment,
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615 - blue. From top to bottom, they are: the maximum CAPE (J kg^{-1}) found in the advance
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617 between 850 and 400 hPa; and maximum relative vorticity (10^{-5}s^{-1}). The thick, solid lines
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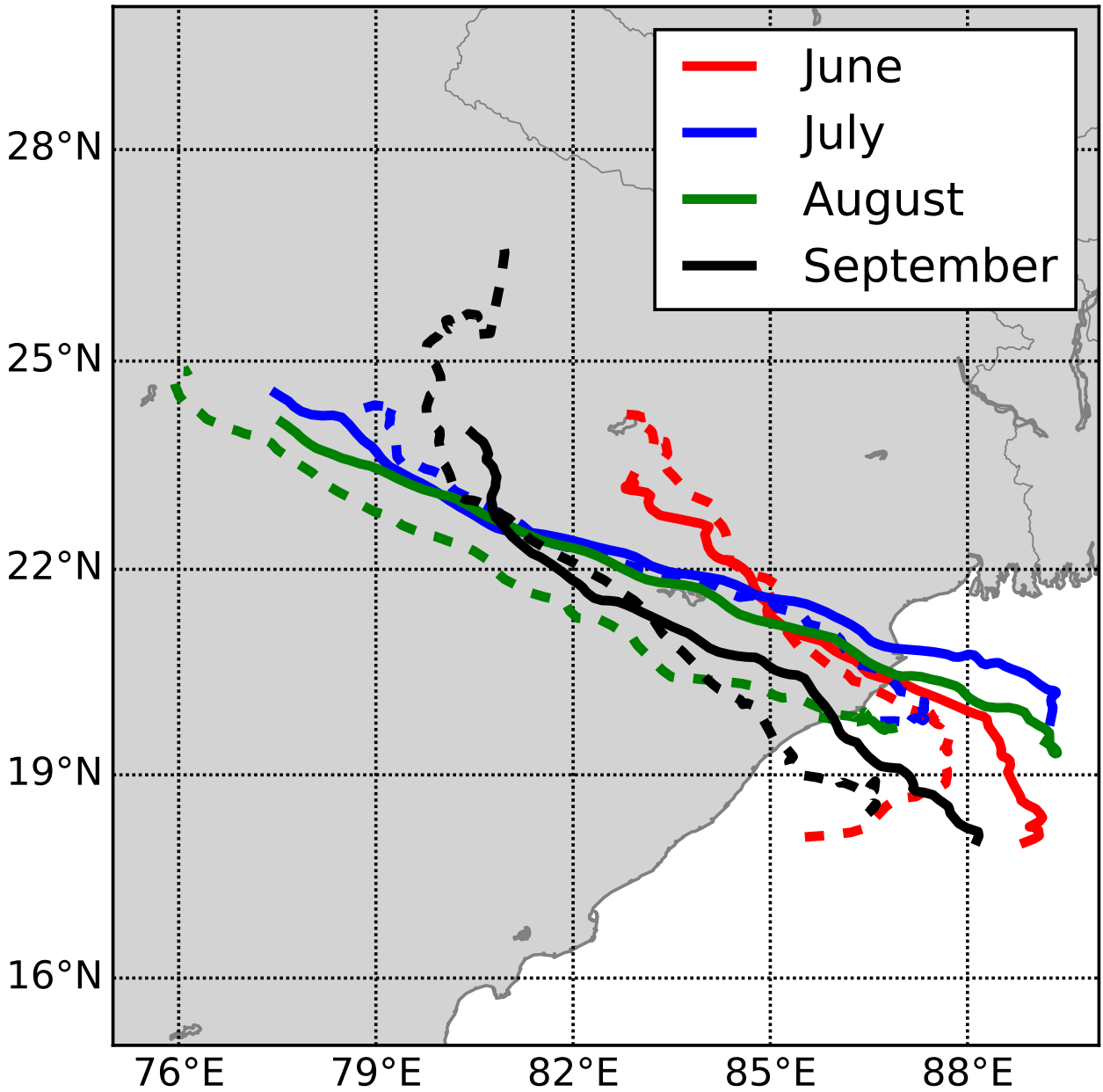


(a) NOAA Climate Prediction Center (1948-2016)

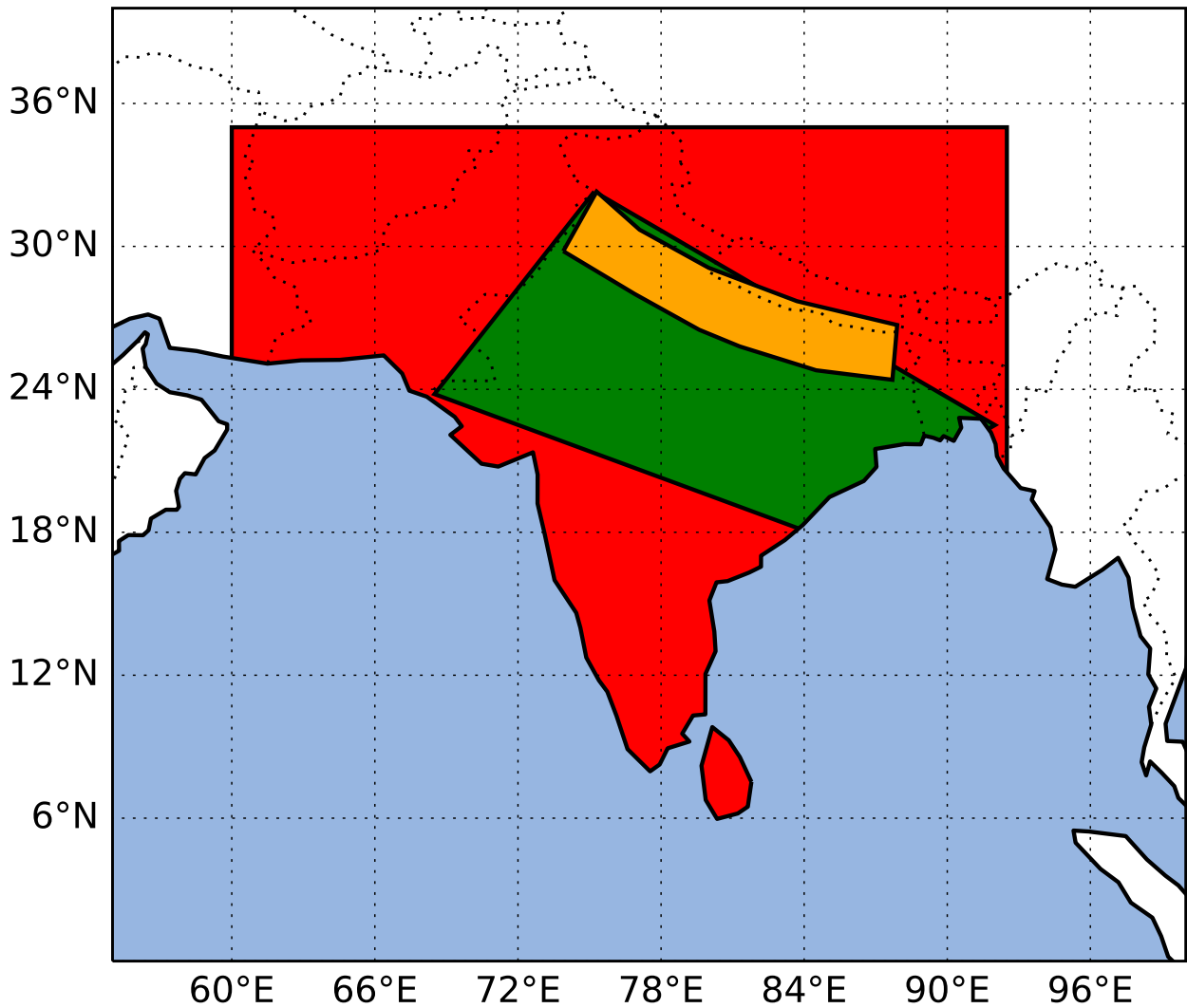


(b) ESA Climate Change Initiative (1991-2014)

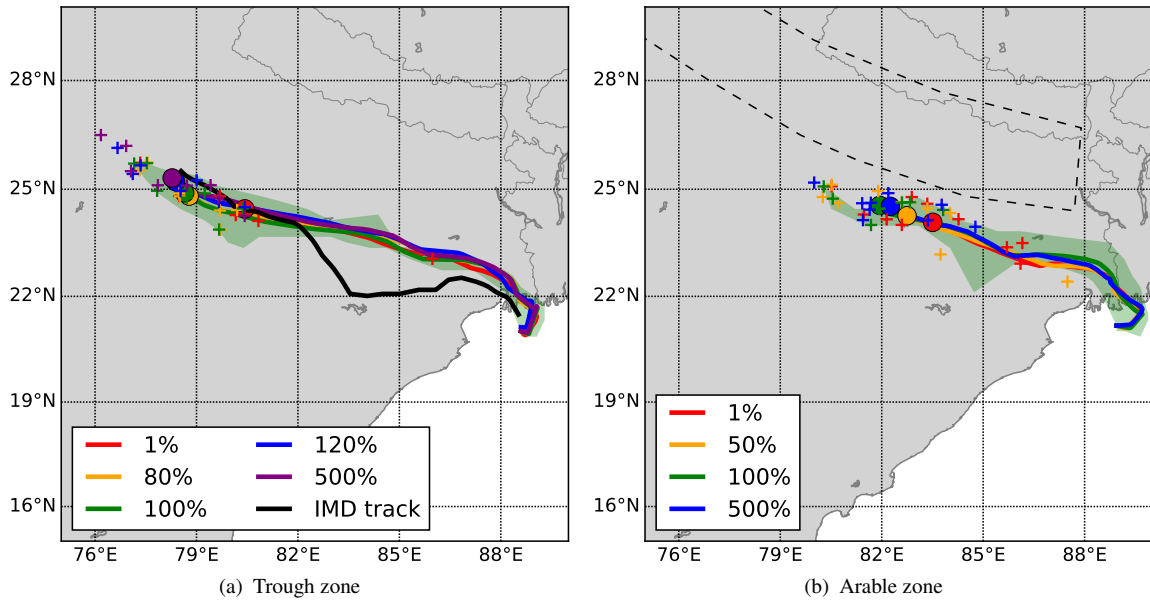
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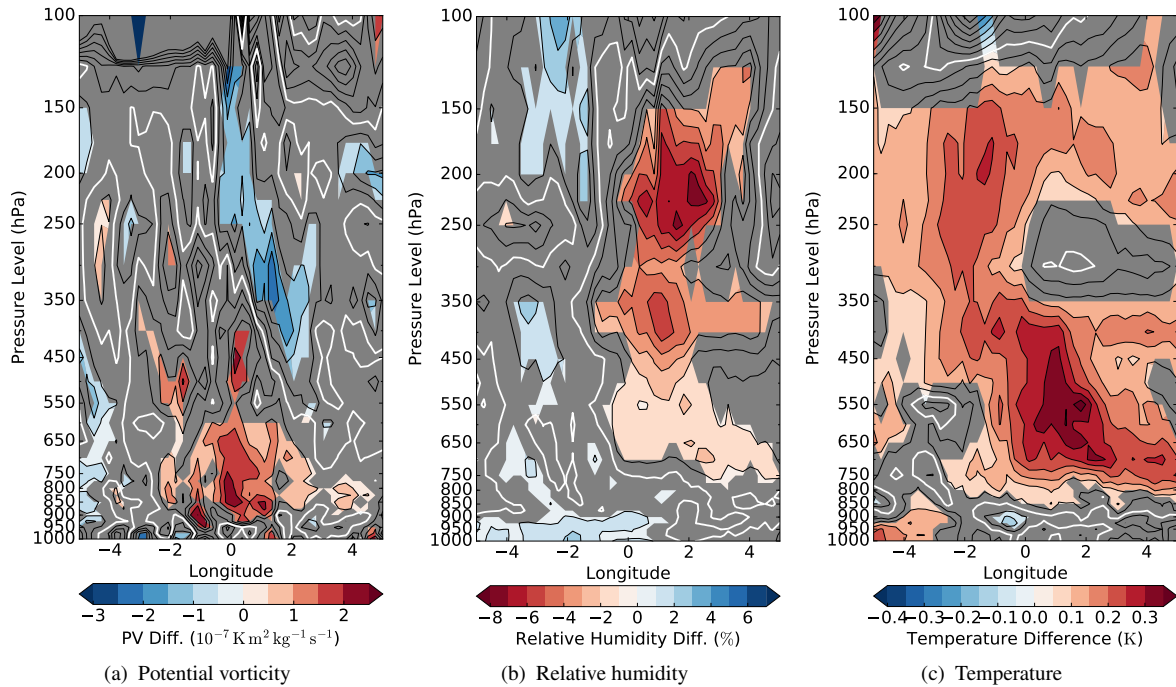
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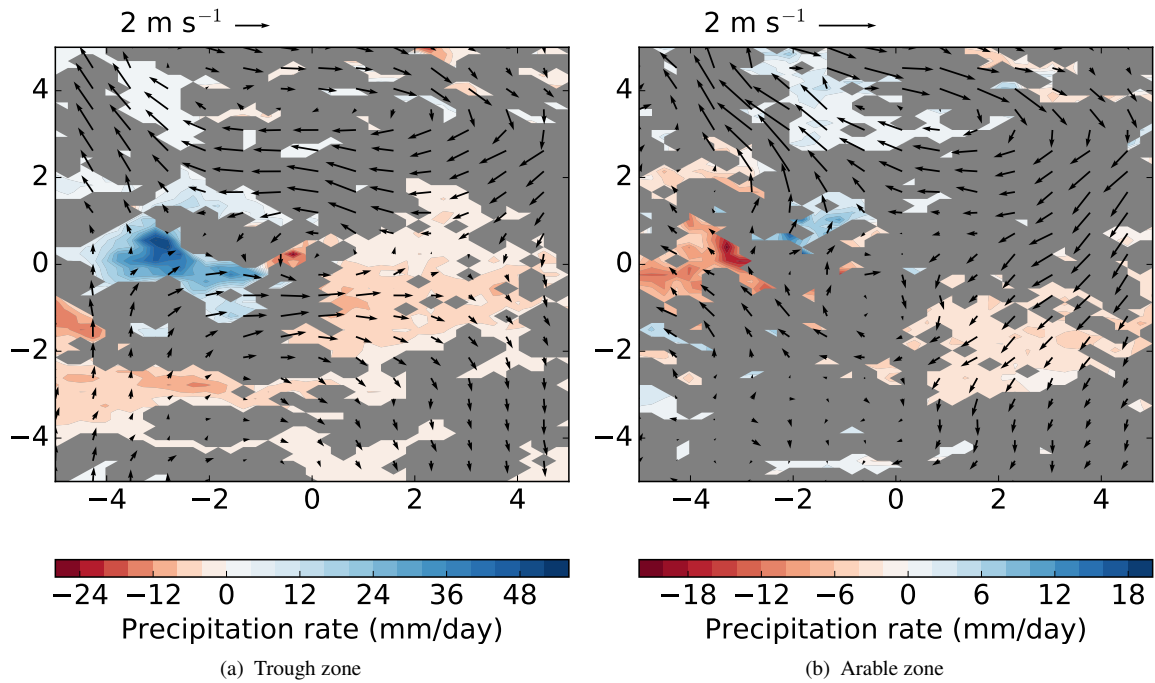
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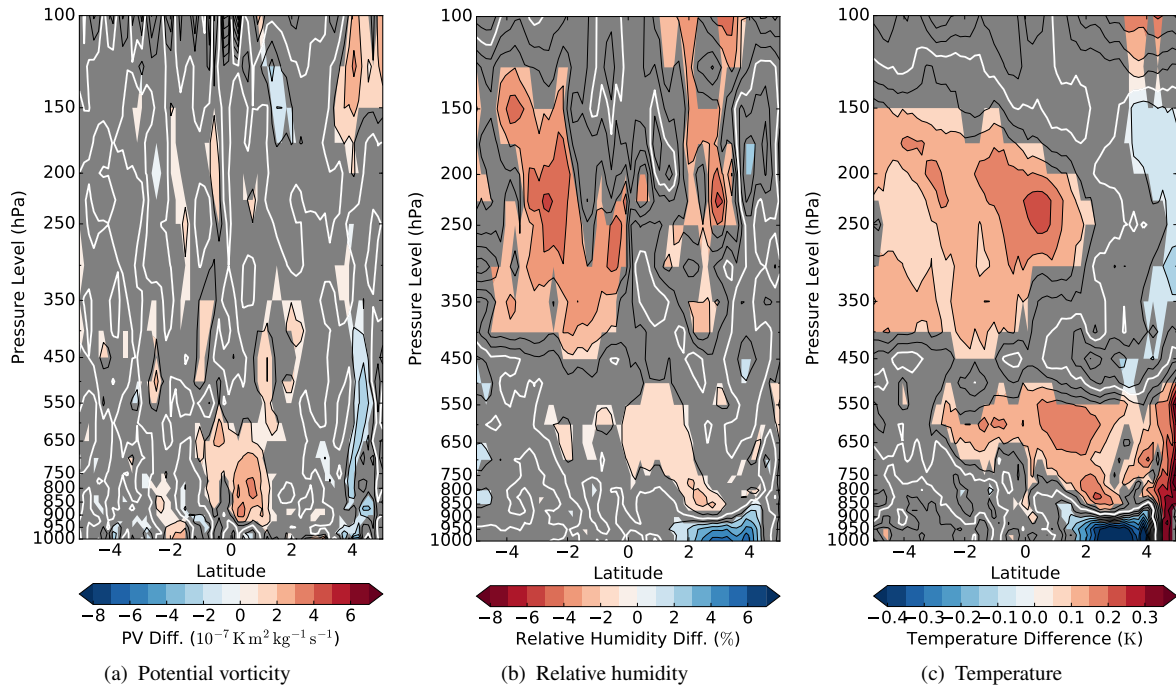
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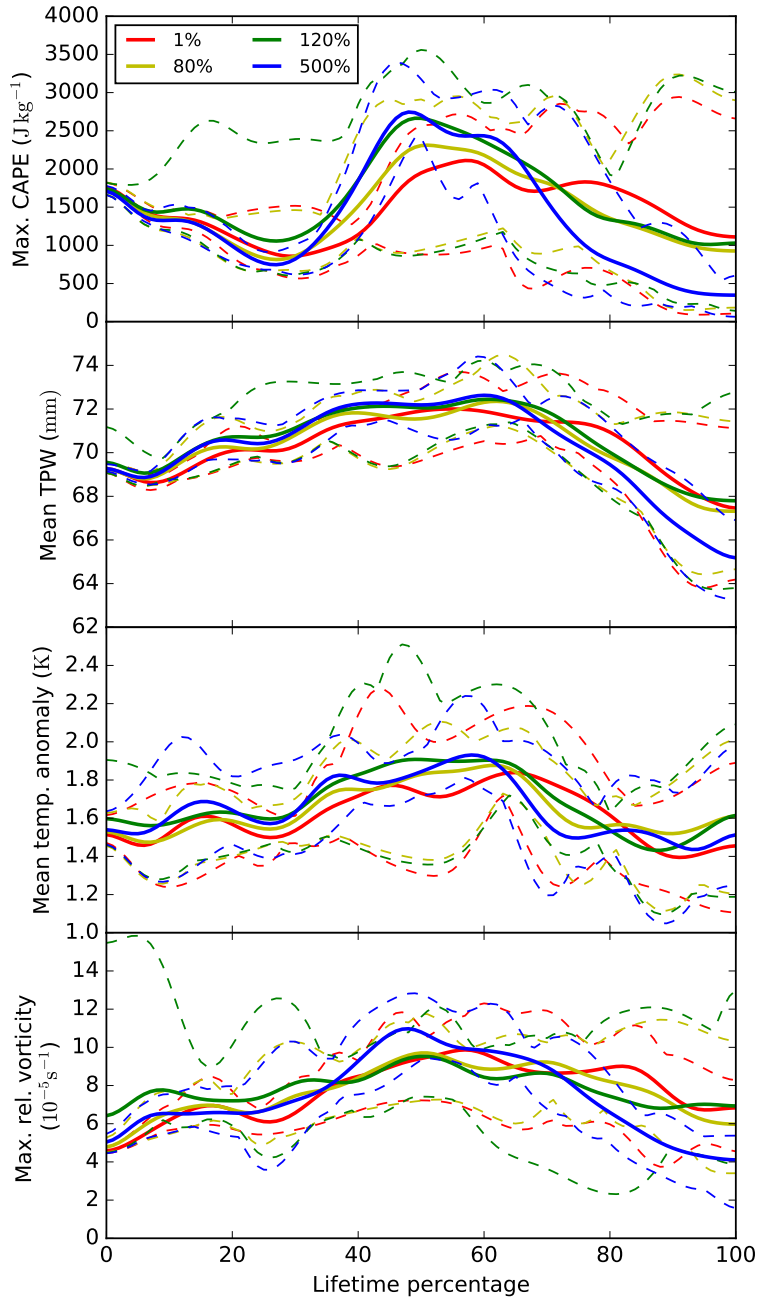
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