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1 Higher CO₂ concentrations increase extreme event 2 risk in a 1.5°C world

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19 **Summary**

20 The Paris Agreement¹ aims to *'pursue efforts to limit the temperature increase to 1.5°C above pre-*
21 *industrial levels.'* However, it has been suggested that temperature targets alone are unable to limit the
22 risks associated with anthropogenic emissions^{2,3}. Here, using an ensemble of model simulations, we
23 show that atmospheric CO₂ increase - a more predictable consequence of emissions compared to global
24 temperature increase - has a significant impact on Northern Hemisphere summer temperature, heat stress,
25 and tropical precipitation extremes. Hence in an iterative climate mitigation regime aiming solely for a
26 specific temperature goal, an unexpectedly low climate response may have corresponding 'dangerous'
27 changes in extreme events. The direct impact of higher CO₂ concentrations on climate extremes therefore
28 substantially reduces the upper bound of the carbon budget, and highlights the need to explicitly limit
29 atmospheric CO₂ concentration when formulating allowable emissions. Thus, complementing global
30 mean temperature goals with explicit limits on atmospheric CO₂ concentrations in future climate policy
31 would reduce the adverse effects of high-impact weather extremes.

32 Main Body

33 The long-term goal of the Paris Agreement¹ aims to strengthen the global response to the threat of
34 climate change by: *‘Holding the increase in the global average temperature to well below 2°C above*
35 *pre-industrial levels and to pursue efforts to limit the temperature increase to 1.5°C above pre-industrial*
36 *levels, recognizing that this would significantly reduce the risks and impacts of climate change.’* To
37 achieve the warming goal, the agreement implements an explicit temperature-based goal-driven regime,
38 with which mitigation efforts are intended to be revised to over time. Therefore, no explicit long-term
39 emissions targets are directly associated with the agreement aims, aside from the need to ultimately
40 achieve a balance between net anthropogenic greenhouse gas sources and sinks into the atmosphere⁴.
41 Since the Paris Agreement, there has been a call for research into impacts associated with 1.5°C and 2°C
42 globally-average surface temperature anomalies^{5,6}. However, previous work^{2,3} suggests global mean
43 temperature targets alone are unable to comprehensively limit the risks from anthropogenic emissions, and
44 that global mean temperature is not the sole driver of changes in temperature and precipitation patterns^{7–11}
45 and extremes^{12–14}.

46 Here we isolate the direct effect of CO₂ concentration, investigating the difference between estimated
47 likely (>66% probability that the value will lie within this range) upper and lower bounds on CO₂
48 concentrations leading to 1.5°C of global warming in 2100 for adaptive pathways to meet the Paris
49 Agreement goals under climate response uncertainty¹⁵. We define the direct effect of CO₂ concentration
50 as all the effects of CO₂ on climate beside those occurring through ocean warming but including feedbacks
51 over land (e.g. from soil moisture), therefore excluding most of the global temperature change since
52 the latter is to a large extent determined by ocean temperatures. We run five Atmospheric Model
53 Intercomparison Project (AMIP) style ensembles¹⁶ with HadAM3P and MIROC5: a present day ensemble

54 (2006-2015); three ensembles with sea surface temperature (SST) levels equivalent to 1.5°C global mean
55 warming relative to 1850-1900 with low, best-estimate and high CO₂ concentrations; and a 2.0°C ensemble
56 with best-estimate CO₂ concentrations and SSTs for a 2.0°C world. By considering the difference between
57 high and low CO₂ concentration ensembles, we find substantial differences in the patterns of change in
58 temperature and precipitation, pointing to an increase in extreme event likelihood. These differences could
59 have important consequences if international climate policy does not seek to limit CO₂ concentrations
60 and instead concentrate exclusively on the global mean temperature target. This idealised experimental
61 design helps focus on the risks associated with different atmospheric CO₂ concentrations, consistent with
62 different climate responses, after global temperatures have stabilised at approximately the same level. We
63 also compare the results from HadAM3P and MIROC5 to results from CAM4 (see Methods for setups),
64 drawing similar conclusions.

65 The differences in global mean temperature and precipitation between the five ensembles are presented
66 in Table S1. Global mean temperature changes between the 1.5°C ensembles are small, as expected
67 due to the prescribed SSTs, with only a 0.12°C difference between low and high CO₂ ensembles in
68 HadAM3P and 0.11°C in MIROC5. Over land, these changes are slightly more pronounced. To correct
69 for the differences in global mean temperature between low, best-estimate and high ensembles, we use a
70 simple linear regression model (see Methods for details), fitting changes in variables to changes in global
71 mean temperature and radiative forcing. Using the regression coefficients and global mean temperature
72 difference between 1.5°C ensembles, we then adjust all variables to have a value associated with the global
73 mean temperature in the best-estimate ensemble. This means all differences between the 1.5°C ensembles
74 are due to the direct radiative forcing effect from differing CO₂ concentrations. All variables and figures
75 (unless specified) for the rest of this study use the corrected ensembles (i.e. with the effect of the global
76 mean temperature differences between the 1.5°C ensembles removed). The spatial patterns of change in

77 HadAM3P due to the direct radiative forcing effect are shown for several variables in Fig. S1 (and without
78 temperature correction in Fig. S2). It may be argued that the SST pattern and CO₂ concentrations are
79 physically inconsistent with each other, but the simulated atmosphere is still physically consistent within
80 the model with its SSTs and CO₂ boundary conditions. The top of atmosphere radiative imbalances in our
81 ensembles lie within the range of imbalances in the historical AMIP simulations (see Fig. S3), and so any
82 conclusions we draw possess the same caveats as conclusions drawn from using the AMIP simulations.

83 We consider three extreme indices: $\overline{TX90p}$, the number of days per season exceeding the 90th
84 percentile of daily maximum temperature in the present day ensemble; $\overline{WBGT95p}$, the number of degree
85 months per year above the 95th percentile of wet-bulb globe temperature (WBGT) in the same month
86 of the present day ensemble; and $\overline{R95p}$, the number of days per season where precipitation exceeds
87 the 95th percentile of daily precipitation on wet days (>1mm/day) in the present day ensemble (see
88 Methods for details). Differences between the high and low CO₂ concentration ensembles are shown for
89 HadAM3P, MIROC5 and CAM4 in Figs. 1, S5 and S7 respectively (and without temperature corrections
90 in Figs. S4, S6, S8). The stippling shows differences discernible against the decadal variability in the
91 model (see Methods for details). We test for statistical significance at the 10% level using a two-sample
92 Kolmogorov-Smirnov test; the hatching in Fig. S9 indicates significant changes in extreme indices in
93 HadAM3P (maps for the other ensembles are not shown, but display similar hatching regions). Due to
94 large ensemble sizes, we find significant changes across large regions of the globe. There is considerable
95 spread in extreme indices in the CMIP5 models at the same temperatures². We sub-sample the CMIP5
96 ensembles, computing the JJA TX90p when each model reaches 1.5°C under the RCP8.5 scenario. The
97 differences in the spatial patterns of TX90p between models with low and high global mean temperature
98 sensitivity to carbon emissions (see Fig. S10), support our findings that increases in CO₂ concentrations
99 lead to increases in extreme temperature.

100 The WBGT, a heat stress metric and proxy for human discomfort^{17,18}, allows us to better quantify the
101 impact of temperature extremes on humans. We consider changes in the extreme metric $\overline{WBGT95p}$ due
102 to increases in CO₂. In JJA, increases are co-located with high population density areas in the Eastern
103 US seaboard, central Europe, the Arabian Peninsula and North-East China and Korea. These regions are
104 likely to experience increases in likelihood and severity of humid heatwaves under climate change¹⁹. Our
105 results suggest these increases may be partly due to differences in atmospheric composition. Increases in
106 population over the coming decades, combined with rising CO₂ concentrations, mean more people will be
107 exposed to the extreme changes in WBGT²⁰ with potential economic impacts²¹ and potentially drastic
108 societal implications²².

109 Area mean changes from present day are summarized in Fig. 2 (and without temperature correction in
110 Fig. S11). Global mean changes in temperature and precipitation are shown, as well as land averaged
111 Northern Hemisphere extratropical (NH.ET), 30°N-90°N, and tropical, 30°S-30°N, JJA changes for the
112 mean and extreme indices. It is clear from the differences between high and low CO₂ ensembles that the
113 area averaged changes in JJA means and extremes are highly significant in both the NH extratropics and
114 tropics for $\overline{WBGT95p}$ and $\overline{TX90p}$. $\overline{R95p}$ changes are significant in HadAM3P (excluding the extreme
115 NH extratropical precipitation), but not in MIROC5.

116 We compare the difference in extreme indices due to the range in CO₂ concentrations consistent
117 with 1.5°C warming scenarios with the difference between the best-estimate CO₂ concentration cases
118 at 2.0°C and 1.5°C global mean warming. The CO₂ induced differences as a percentage of the global
119 mean warming induced differences for HadAM3P are shown for $\overline{TX90p}$ (Fig. 3a,b) and $\overline{WBGT95p}$
120 (Fig. 3c,d) (and without temperature correction in Fig. S12). In JJA, over the NH midlatitudes, there are
121 regions where the differences due to CO₂ are greater than those due to the extra 0.5°C of global mean
122 warming. For $\overline{R95p}$, we show the zonally averaged differences for ensemble means in the solid colours,

123 and individual ensembles member differences in the light colours (Fig. 3e,f) (and without temperature
124 correction in Fig. S12). In both DJF and JJA, the differences due to CO₂ in the tropics are on the same
125 order of magnitude as the differences due to the extra 0.5°C of global mean warming. The importance of
126 CO₂ on driving tropical extreme precipitation is contrary to the extratropics where global mean warming
127 is the main driver²³. The comparisons for MIROC5 (Fig. S13, and without temperature correction in Fig.
128 S14) vary from HadAM3P, with slightly lower magnitude, but nevertheless show substantial changes due
129 to CO₂ concentration differences compared to differences due to the extra 0.5°C of global mean warming.

130 These results highlight that increases in extreme indices due to the direct effect of CO₂ have important
131 implications on the upper limit of CO₂ that can be emitted in order to limit changes in specific costly
132 climate extremes. If we accept a level of extreme index increase consistent with 1.5°C of global mean
133 warming associated with the increased CO₂ in the best-estimate climate sensitivity case, we can ask what
134 the upper bound of the carbon budget is that will keep extreme event likelihoods at this level. This will
135 reduce the upper bound of the carbon budget, which is currently set by the uncertainty in the CO₂ emission
136 levels which lead to 1.5°C of global mean warming. This allowable emissions uncertainty arises due to
137 the global mean temperature response uncertainty²⁴.

138 Using the simple regression model (see Methods for details), we calculate how each extreme index
139 varies with global mean temperature and CO₂ concentration. Taking the extreme index value at 1.5°C of
140 global mean warming in the best-estimate climate sensitivity case, we can then compute pairs of values of
141 global mean temperature and CO₂ concentration resulting in that same extreme index value. This allows us
142 to determine an upper bound for the carbon budget consistent with this extreme index change, as illustrated
143 by Fig. 4 (see caption for details). For $\overline{TX90p}$ (Fig. 4a), the decrease in the upper bound of the carbon
144 budget is from 569 to 471 GtC, giving a $23^{+11}_{-12}\%$ decrease in the uncertainty range of the carbon budget.
145 For $\overline{WBGT95p}$ and $\overline{R95p}$ the new upper bounds are 450 GtC and 438 GtC, and reductions in uncertainty

146 are $28^{+6}_{-9}\%$ and $31^{+13}_{-17}\%$ respectively (Figs. 4b,c). Reductions due to regional $\overline{WBGT95p}$ changes are
147 shown in Fig. S15. Besides Western Africa, all regions display a decreased upper bound. We stress that
148 these reductions in the carbon budget are only estimates based on the results from the three models. The
149 use of a larger number of climate models run under a similar experiment, if they were available, would
150 allow us to quantify the reduction more accurately. With the full range of CMIP5 models, there would be
151 a greater spread in the uncertainty of the new upper bound of the carbon budget. However, based on the
152 magnitude of the reduction in the upper bound of the carbon budget from these three models, it seems
153 likely that the direct CO₂ effect is important, and should be taken into consideration when formulating
154 carbon budgets to avoid any given level of climate impacts on extremes.

155 We demonstrate significant differences in temperature and precipitation extremes between the higher
156 and lower likely CO₂ concentrations in a 1.5°C world. This highlights the importance of direct, local,
157 CO₂ forcing effects on regional climates and extremes. We must acknowledge these effects so that we
158 can avoid ‘dangerous’ changes in extremes which, in the context of the Paris Agreement, are defined as
159 impacts beyond those expected at 1.5°C. It also makes clear that impacts on extremes in a world warming
160 past 1.5°C would be different from those after stabilising at 1.5°C.

161 Differences in extremes at the same global mean temperature, due purely to differing CO₂ concentra-
162 tions, directly impact the use of the pattern scaling technique²⁵. These differences provide compelling
163 evidence that when using pattern scaling, we must account for CO₂ concentration and not just the changes
164 in global mean temperature (which is the dominant method used in academic and policy work).

165 This study supports findings²⁶ that geoengineering schemes aimed at reducing the global warming im-
166 pacts without reducing CO₂ concentration would not fully mitigate changes in extremes whose likelihoods
167 have increased by the direct effect of increasing CO₂ concentrations.

168 Whilst it is important to note these findings are from only three models, the use of large ensembles

169 and the statistical significance of the differences give weight to the reliability of our findings. Previous
170 work¹¹ shows the rapid regional precipitation response to increases in CO₂ is robust among CMIP5
171 models. Further work must be carried out to explore the spatial patterns and magnitudes of temperature
172 and precipitation differences in other models to better quantify how extremes change with different
173 atmospheric CO₂ concentrations. Despite this, it is clear that complementing global mean temperature
174 goals with explicit limits on atmospheric CO₂ concentrations would reduce the risk of unexpectedly high
175 changes in high-impact weather extremes.

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192 **Author contributions statement**

193 M.R.A. conceived the experiment. S.S ran the HadAM3P simulations, H.S. ran the MIROC5 simulations,
194 and U.B. ran the CAM4 simulations. H.S.B. performed the analysis with input from D.J.K., R.J.M., B.P.G.,
195 T.W., D.M. and M.R.A.. R.J.M. contributed to one of the figures. H.S.B. wrote the paper with comments
196 from all the authors.

197 **Additional information**

198 Reprints and permissions information is available at www.nature.com/reprints. The authors declare
199 no competing interests. Correspondence and requests for materials should be addressed to H.S.B.
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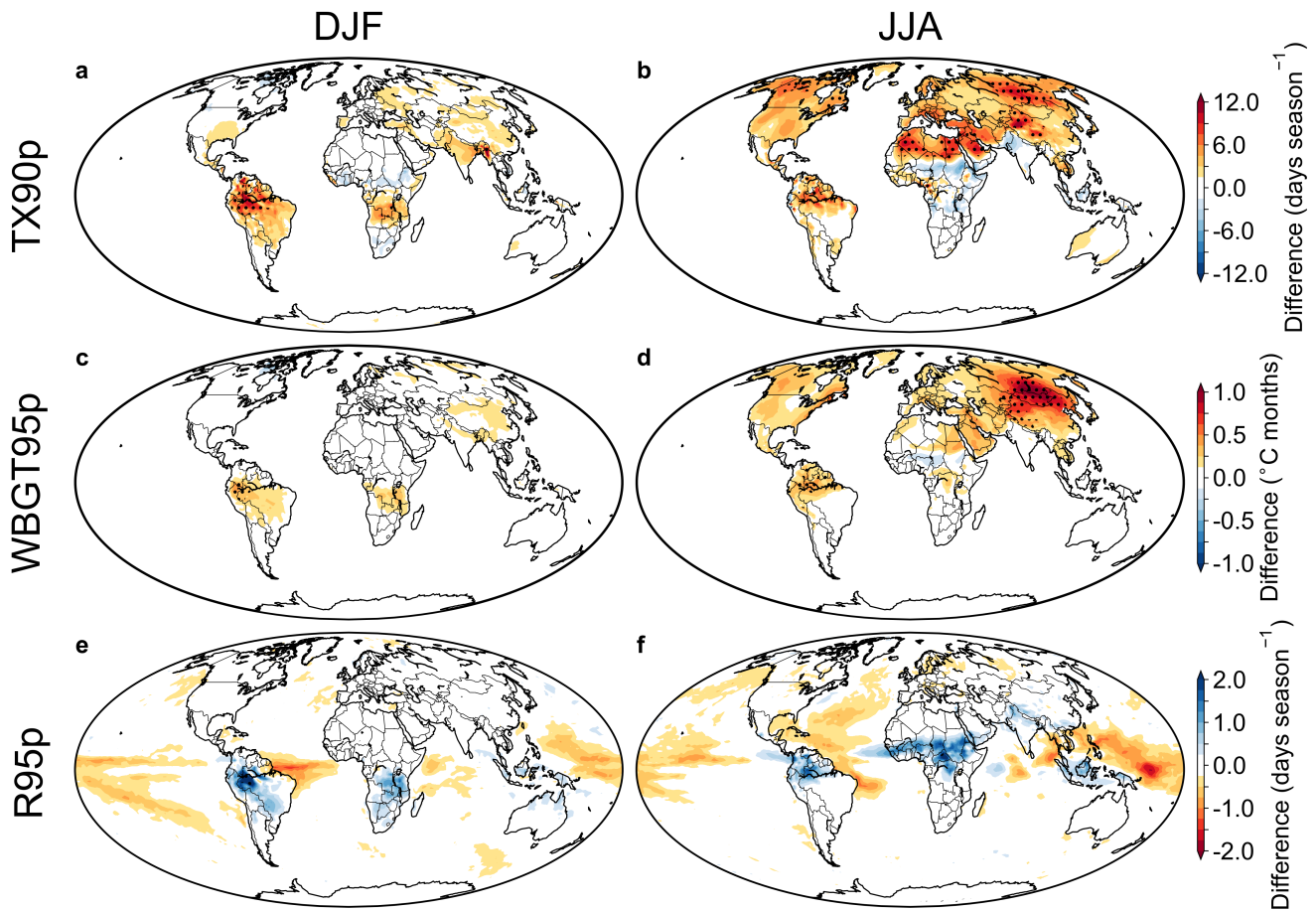


Figure 1. Changes associated with CO₂ concentration at 1.5°C warming in HadAM3P with temperature correction applied. Differences in $\overline{TX90p}$ (a, b), $\overline{WBGT95p}$ (c, d) and $\overline{R95p}$ (e, f) between high and low CO₂ forced runs. Stippling indicates statistical significance at the 10% level when testing for discernible impact (i.e. with an effective sample size of one, see Methods for details). Note the reversal of the colourbar for e, f.

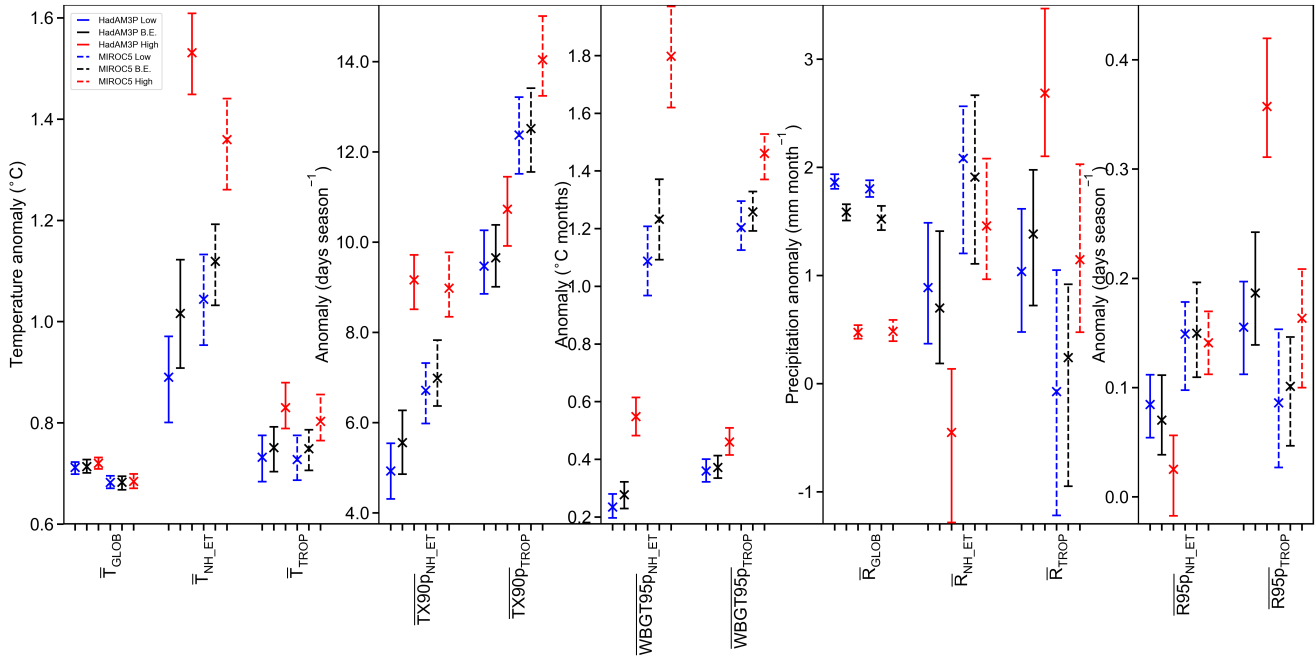


Figure 2. Area averaged differences associated with CO₂ concentration at 1.5°C warming with temperature correction applied. Whisker plots show the difference in indices between the low, best estimate and high CO₂ ensembles relative to the present day ensemble (solid lines: HadAM3P, dashed lines: MIROC5). Crosses mark the 50th percentile and the caps mark the 10th and 90th percentiles. Global mean temperature differences are calculated averaged over all seasons and the whole globe. ‘NH_ET’ indices are calculated for JJA 30°-90°N land, ‘TROP’ indices are calculated for JJA 30°S-30°N land. Precipitation anomalies are expressed as the mean change at each gridpoint.

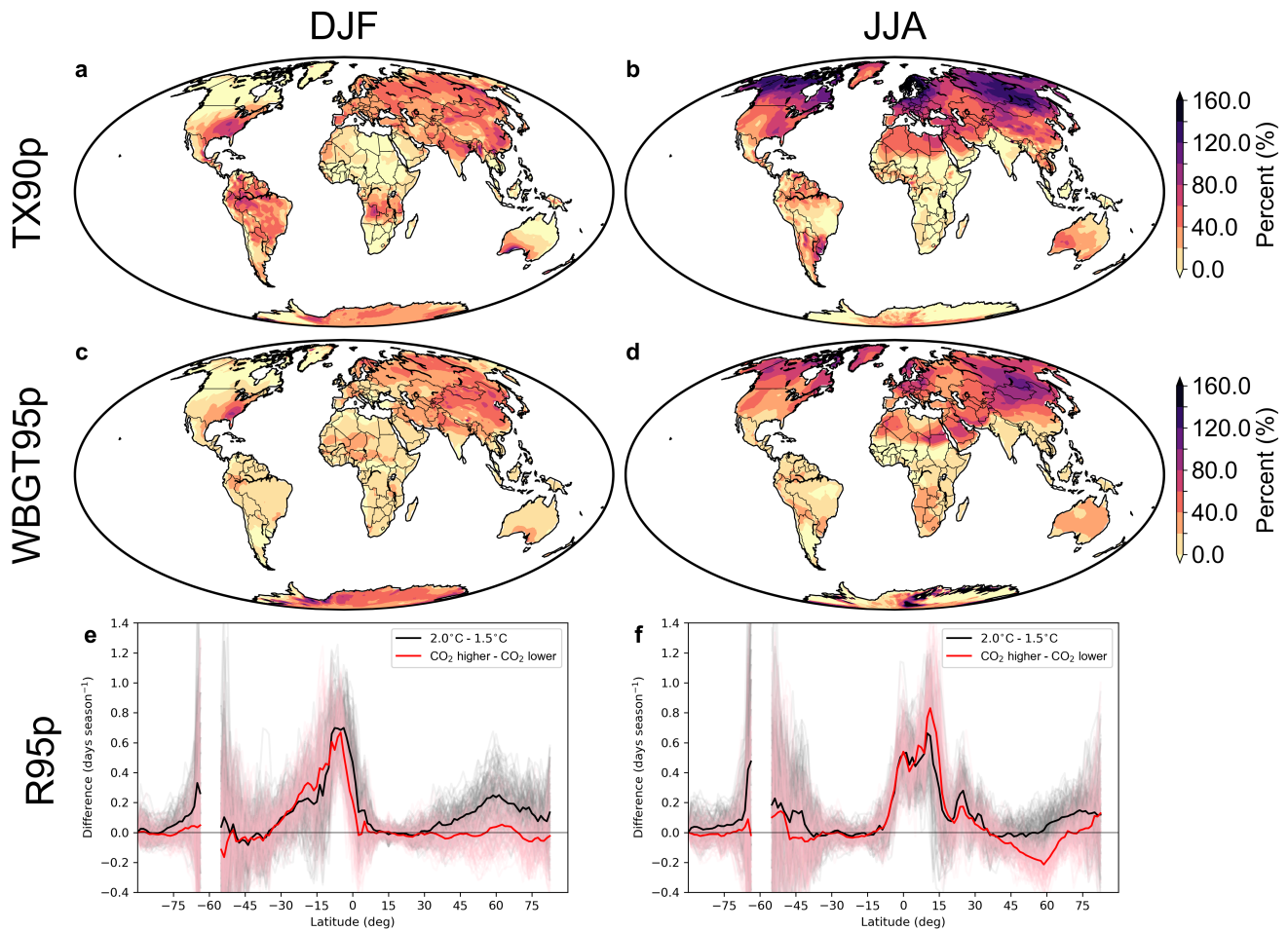


Figure 3. Comparison of changes due to CO_2 concentration with changes due to 2.0°C to 1.5°C warming difference in HadAM3P. Differences between high and low CO_2 , with temperature correction applied, as a percentage of the difference between 2.0°C and 1.5°C ensembles for $\overline{TX90p}$ (a, b) and $\overline{WBGT95p}$ (c, d). The differences between high and low CO_2 , and the differences between 2.0°C and 1.5°C for $\overline{R95p}$ (e, f). Plots show the zonally averaged mean ensemble $\overline{R95p}$ changes in solid colours, and individual ensemble members in the light colours.

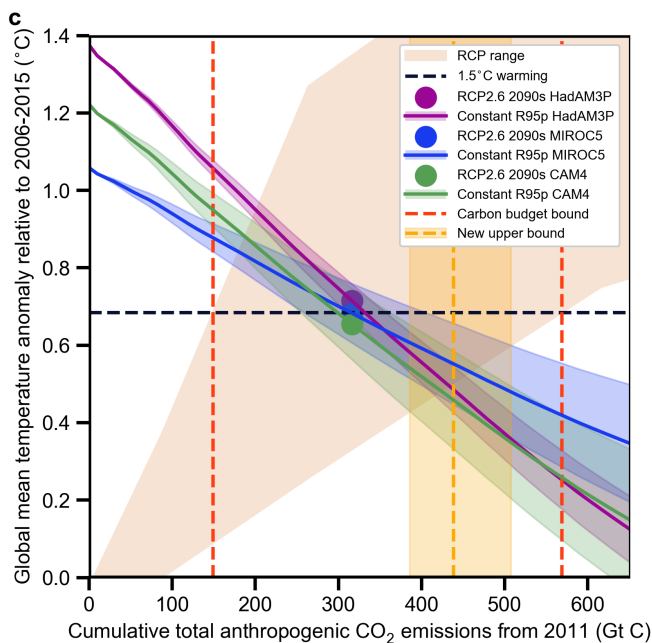
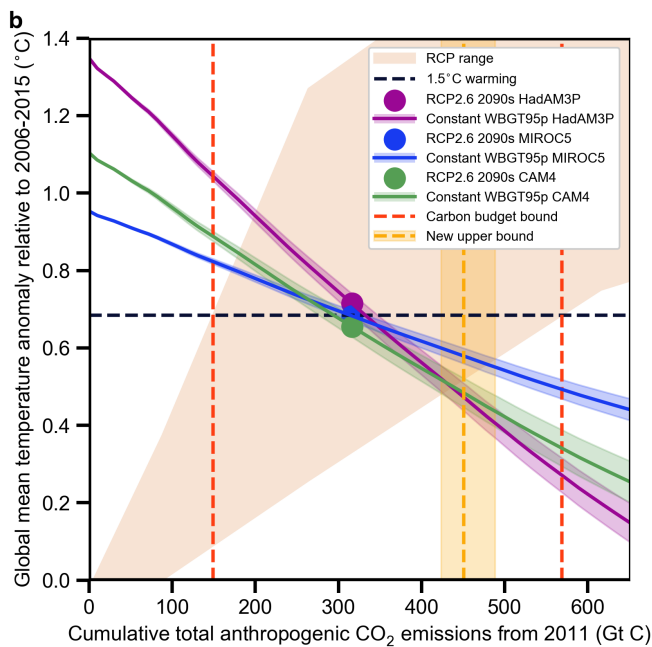
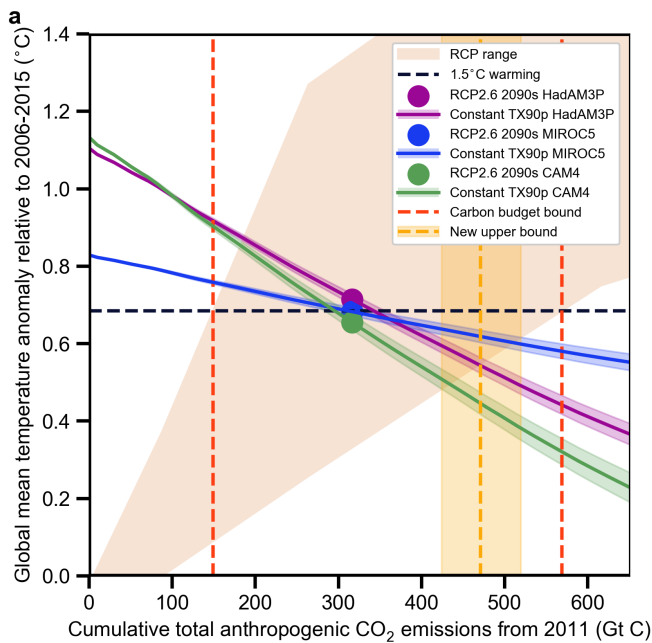


Figure 4. The reduction in the upper bound of the carbon budget. $\overline{TX90p}$ (JJA 30°-90°N land) (a), $\overline{WBGT95p}$ (JJA 30°-90°N land) (b) and $\overline{R95p}$ (JJA 30°S-30°N land) (c). The pink plume shows the 5-95 percentile of the CMIP5 response under all four RCPs relative to the 2006-2015 mean. The black dashed line marks the warming from the present decade equivalent to 1.5°C of mean global warming from the pre-industrial period (1850-1900). Dots indicate the warming observed in our best-estimate CO₂ concentration ensemble for each model (corresponding to the RCP2.6 2090-2099 mean). Curves show lines of constant extreme index, calculated from the regression model, with 5-95% uncertainty plumes. Red dashed lines mark the upper and lower uncertainty bounds of the carbon budget. The yellow dashed line is the new multi-model mean upper bound in the carbon budget, with uncertainty plume given by the pink plume's intersection with the model 5th percentile that gives the largest reduction in the carbon budget, to the intersection with the model 95th percentile that gives the smallest reduction in the carbon budget.

283 **Methods**

284 **Experimental design**

285 For this study, we make use of three climate models: HadAM3P, MIROC5²⁷ and CAM4²⁸. HadAM3P
286 is an atmosphere only, medium resolution, GCM developed by the UK Met Office. It is based upon the
287 atmospheric component of HadCM3^{29,30}. An improved version of HadAM3P using a more sophisticated
288 land-surface scheme is used here³¹. The model has been used extensively in the study of extreme events.
289 We run HadAM3P using the large-ensemble capability provided by the *climateprediction.net* volunteer
290 computing network^{31,32}, where members of the public are performing multi-thousand-member initial
291 condition ensemble general circulation model (GCM) simulations at 1.25° x 1.875° resolution. We
292 compare the results from HadAM3P to MIROC5 and CAM4 run at 1.4° x 1.4° and 1.9° x 2.5° resolutions
293 respectively.

294 The experimental setup follows the Half a degree Additional warming, Prognosis and Projected
295 Impacts (HAPPI; www.happimip.org) design³³. Forcing conditions are as in the DECK AMIP design,
296 including SSTs and sea ice¹⁶. The HAPPI experiments are designed to simulate conditions in the present
297 decade (2006-2015), and 1.5°C and 2.0°C warmer than pre-industrial (1861-1880) conditions. SSTs
298 for the 1.5°C case are calculated by adding to the observed 2006–2015 SSTs a change in SST (Δ SST)
299 between the decadal average of the modelled 2006–2015 period and the decadal average of the modelled
300 1.5°C world over 2091–2100³³. Hence the SST patterns are still time varying because they are based on
301 the 2006–2015 observations, but they have an additional warming added to them. As CMIP5 historical
302 simulations stopped in 2005, the decadal average of the 2005–2015 SSTs is estimated from RCP8.5
303 simulations, as this is the scenario that is closest to observations over this period. The decadal average of
304 the 2091–2100 SSTs is estimated from CMIP5 RCP2.6 simulations. The process to calculate the 2.0°C

305 SST pattern is similar and outlined in detail in reference [33]. Here we use the HAPPI Tier 1 experimental
306 design³³, which uses the multi-model mean patterns, thus across all three models used in this study all
307 the present day ensembles are run with an identical present day SST pattern, all the 1.5°C ensembles are
308 run with an identical 1.5°C SST pattern, and all the 2.0°C ensembles are run with an identical 2.0°C SST
309 pattern. Full details are discussed in reference [33].

310 For HadAM3P, we run five ensemble experiments, each over a 10-year period. A present day ensemble
311 is run using HAPPI present day setup³³ over the period 2006-2015 with an average CO₂ of 390.4ppm (90
312 ensemble members). A 1.5°C ensemble is run using HAPPI 1.5°C setup³³, which uses the RCP2.6 forcing
313 scenario boundary conditions from the last decade of the 21st century, with CO₂ fixed at 423.1ppm (71
314 members). Two further ensembles are run using the 1.5°C setup³³, but with CO₂ fixed at 395.8ppm and
315 550.0ppm (76 and 88 members) to represent the lower and higher likely CO₂ concentrations averaged over
316 2091-2100 in adaptive pathways that succeed in achieving warming below 1.5°C in 2100 for the assessed
317 ranges of climate response uncertainty¹⁵. The range of CO₂ concentration is intended to illustrate the
318 scale of the difference in concentrations that may be consistent with a 1.5°C world. Thus the range of
319 concentrations used does not affect the qualitative results of changes in extremes. As the concentrations are
320 only used to fit the regression model and not calculate changes in the carbon budget directly, the reduction
321 in the upper bound of the carbon budget is not influenced by the exact choice of CO₂ concentrations.
322 It is however influenced by the selection of models used in the study, due to their different climate
323 sensitivities. A similar range in concentrations can be obtained by converting the spread of individual
324 model temperatures in CMIP5 under RCP2.6 at the end of the century (2081-2100) to a concentration
325 range. The change in temperature at the end of the century from the pre-industrial baseline stated in AR5²⁴
326 is 1.6°C with a standard deviation of 0.4°C and 5-95% uncertainty range of 0.9-2.3°C. The radiative
327 forcing averaged over 2081-2100 is 2.60 Wm⁻², from which we compute the ratio of temperature to

328 radiative forcing. When multiplied by 1.5°C and converted to a concentration range, this gives a range of
329 400.2-510.2 ppm using the range of temperatures associated with one standard deviation from the mean,
330 or 381.6-624.7 ppm when using the 5-95% range. A 2.0°C ensemble is run with the HAPPI 2.0°C SST
331 pattern³³ and CO₂ fixed at 486.6ppm (96 members). All CO₂ concentrations are prescribed as a global
332 mean atmospheric concentration. Following previous work³⁴, initial condition perturbations are applied
333 between ensemble members via perturbations to the potential temperature. For MIROC5, we perform
334 exactly the same ensemble runs, with 50 members per experiment using the same CO₂ concentrations
335 as HadAM3P. For CAM4, we use the data from three experiments, a present day, 1.5°C, and 1.5°C with
336 a CO₂ concentration of 379.0ppm (thus for CAM4, the 1.5°C ensemble with CO₂ at 423.1ppm doubles
337 as the ‘best-estimate’ and ‘high’ ensembles and the 1.5°C with CO₂ at 379.0ppm becomes the ‘low’
338 ensemble), all using the HAPPI experimental design with 501 ensemble members.

339 Following standard AMIP design protocol¹⁶, soil moisture is allowed to vary freely in our simulations.
340 The primary aim of the study is to disentangle the direct CO₂ effect on regional climate from the ocean
341 warming, since the latter strongly affects global mean warming and, additionally, may cause some large
342 scale circulation changes. SST patterns can contribute to regional climate variability far further afield than
343 their perturbation, whereas soil moisture effects are more local and do not affect global mean temperature
344 significantly³⁵. Some of the regional effect (and indeed perhaps a large part of this effect over land areas³⁵)
345 is due to changes in soil moisture. Holding soil moisture constant would mask some of the changes due to
346 the direct effect of CO₂, preventing us from disentangling this effect from that of the global mean warming
347 of the ocean.

348 We select the HAPPI experimental design as it allows us to isolate the direct effect of changing CO₂
349 on climate extremes. However, other possible experimental designs do exist where it would be possible to
350 investigate this effect too^{36,37}. In the setup described in reference [36], 1.5°C warming is arrived at by

351 2100 under two different emissions scenarios, one keeping concentrations below 440 ppm, and the other
352 overshooting and then ramping down CO₂ concentrations to arrive at 1.5°C. In this setup however, there
353 are many other differences between the simulations at 1.5°C, e.g. sea ice-cover, and so the setup makes it
354 hard to disentangle the direct response on extremes due to differing atmospheric CO₂ concentrations. In
355 the setup described in reference [37], first the CO₂ emissions are determined that drive a particular coupled
356 model (CESM) to arrive at 1.5°C and 2.0°C global warming, and then the coupled model is run with this
357 CO₂ concentration. Unlike the HAPPI setup, this guarantees the physical consistency of the setup and
358 does not result in an SST response and CO₂ concentration that are potentially inconsistent. However, to
359 then investigate the impact of changed atmospheric CO₂ concentration on extremes, one would then have
360 to compare the impact across different models, making it hard to disentangle the direct effect of CO₂ from
361 all the other differences that may give rise to that difference in sensitivity (e.g. model physics and SST
362 anomaly pattern).

363 **Regression model**

364 To compute the contribution of changes in radiative forcing and changes in global mean temperature to
365 a change in a particular variable, we fit the changes in the 10-year ensemble mean variables between
366 present and 1.5°C runs to the model: $\Delta X = \alpha \Delta F + \beta \Delta \bar{T}$. ΔX is the change in variable. ΔF is the change
367 in the radiative forcing due to CO₂ from the present decade, $\Delta F = \frac{3.71}{\ln 2} \cdot \ln \left(\frac{C}{C_p} \right)^{24}$ where C and C_p are
368 the CO₂ concentration in the forced run and the present decade respectively in ppm. $\Delta \bar{T}$ is the change in
369 global mean temperature between the 1.5°C ensembles and the present day ensemble (not the change in
370 mean SST, thus $\Delta \bar{T}$ is different for the three 1.5°C ensembles). The regression framework allows us to
371 account for the change in variables from the increase in global mean temperature arising from increased
372 CO₂ concentrations, and so separate the effect of direct radiative CO₂ from the effect of the global mean
373 temperature increase. The fit parameters α and β indicate the dependence of the change in variable to

374 changes in CO₂ radiative forcing and global mean temperature respectively. Uncertainty estimates in
375 warming for a given radiative forcing are calculated from the covariance matrix of α and β which accounts
376 for the spread in 10-year ensemble member mean quantities.

377 To apply the global mean temperature change correction, for each variable we subtract $\beta (\bar{T}_H - \bar{T}_{BE})$
378 from the high CO₂ ensemble variable, and add $\beta (\bar{T}_{BE} - \bar{T}_L)$ to the low CO₂ ensemble variable (subscripts:
379 H - high ensemble, BE - best-estimate ensemble, L - low ensemble). For maps, β is calculated for each
380 individual grid-point, for area mean indices, β s are calculated from the area mean of the variable in
381 question.

382 **Extreme indices**

383 The extreme measures $\overline{TX90p}$ and $\overline{R95p}$ are taken from the dictionary of the European Climate As-
384 sessment and Dataset project (ECA&D), which has been commonly used in previous studies about
385 climate extremes³⁸⁻⁴⁰. Precise definitions of the full list of indices are available at the ECA&D web-
386 site (<http://eca.knmi.nl/indicesextremes/indicesdictionary.php>). The percentile thresholds for computing
387 $\overline{TX90p}$ and $\overline{R95p}$ are calculated from the present day ensemble. Due to the quantity of daily data, we
388 have not applied the five day filtering window when calculating the percentile threshold for $\overline{TX90p}$.

389 The simplified wet-bulb globe temperature (WBGT)⁴¹ is given by: $WBGT = 0.567T + 0.393e + 3.94$,
390 where T is the air temperature in degrees Celsius and e denotes the water vapour pressure in hPa. Water
391 vapour pressure is calculated from relative humidity by $rH = \frac{e}{E} \times 100\%$. The saturation water pressure, E
392 (in hPa), is approximated using the Magnus formula⁴²: $E(T) = 6.112hPa \times \exp\left(\frac{17.62T}{243.12^\circ C + T}\right)$. We define
393 the index $\overline{WBGT95p}$ for a particular month as the number of degree months per year at each grid point
394 above the 95th percentile of WBGT in the same month of the present day ensemble.

395 Significance testing

396 We test for a discernible signal against the decadal variability in the model by assessing the difference
397 between the means of the two sets of 10 year extreme index means for each ensemble member. To test
398 for discernible changes between the high and low CO₂ runs, the t-statistic does not depend on ensemble
399 size, and is given by $t = \frac{\mu_h - \mu_l}{\sqrt{\sigma_h^2 + \sigma_l^2}}$, where the μ_i are the ensemble means in the high and low CO₂ cases,
400 and the σ_i the standard deviations of the 10-year ensemble member means. This tests for a decadal
401 signal against the internal variability inherent in the climate system. We assess discernibility at the
402 10% significance level. We also test for significant changes in the extreme index distributions using a
403 two-sample Kolmogorov-Smirnov test at the 10% significance level.

404 Data availability

405 Model output data is being made available from the NERSC data portal: <http://portal.nersc.gov/c20c/data.html>.

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