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1 **A large ozone-circulation feedback and its implications for**
2 **global warming assessments**

3

4 Peer J. Nowack^{1*}, N. Luke Abraham^{2,1}, Amanda C. Maycock^{1,2}, Peter Braesicke^{2,1,3},
5 Jonathan M. Gregory^{2,4,6}, Manoj M. Joshi^{2,4,5}, Annette Osprey^{2,4} and John A. Pyle^{1,2}

6 ¹Centre for Atmospheric Science, Department of Chemistry, University of
7 Cambridge, Cambridge, United Kingdom

8 ²National Centre for Atmospheric Science, United Kingdom

9 ³now at: Karlsruhe Institute of Technology, IMK-ASF, Karlsruhe, Germany

10 ⁴Department of Meteorology, University of Reading, Reading, United Kingdom

11 ⁵now at: Centre for Ocean and Atmospheric Sciences, University of East Anglia,
12 Norwich, United Kingdom

13 ⁶Met Office Hadley Centre, Met Office, Exeter, United Kingdom

14

15 **State-of-the-art climate models now include more climate processes which are**
16 **simulated at higher spatial resolution than ever¹. Nevertheless, some**
17 **processes, such as atmospheric chemical feedbacks, are still computationally**
18 **expensive and are often ignored in climate simulations^{1,2}. Here we present**
19 **evidence that how stratospheric ozone is represented in climate models can**
20 **have a first order impact on estimates of effective climate sensitivity. Using a**
21 **comprehensive atmosphere-ocean chemistry-climate model, we find an**
22 **increase in global mean surface warming of around 1°C (~20%) after 75 years**
23 **when ozone is prescribed at pre-industrial levels compared with when it is**
24 **allowed to evolve self-consistently in response to an abrupt 4xCO₂ forcing.**
25 **The difference is primarily attributed to changes in longwave radiative**
26 **feedbacks associated with circulation-driven decreases in tropical lower**
27 **stratospheric ozone and related stratospheric water vapour and cirrus cloud**
28 **changes. This has important implications for global model intercomparison**
29 **studies^{1,2} in which participating models often use simplified treatments of**

30 **atmospheric composition changes that are neither consistent with the**
31 **specified greenhouse gas forcing scenario nor with the associated**
32 **atmospheric circulation feedbacks³⁻⁵.**

33 Starting from pre-industrial conditions, an instantaneous quadrupling of the
34 atmospheric CO₂ mixing ratio is a standard climate change experiment (referred to
35 as abrupt4xCO₂) in model intercomparison projects such as the Coupled Model
36 Intercomparison Project phase 5 (CMIP5)¹ or the Geoengineering Model
37 Intercomparison Project (GeoMIP)². One aim of these initiatives is to offer a
38 quantitative assessment of possible future climate change, with the range of
39 projections from participating models commonly used as a measure of uncertainty⁶.
40 Within such projects, stratospheric chemistry, and therefore stratospheric ozone, is
41 treated differently in individual models. In CMIP5 and GeoMIP, the majority of
42 participating models did not explicitly calculate stratospheric ozone changes^{2,4}. For
43 abrupt4xCO₂ experiments, modelling centres thus often prescribed stratospheric
44 ozone at pre-industrial levels^{2,5}. For transient CMIP5 experiments, it was instead
45 recommended to use an ozone field derived from the averaged projections of 13
46 chemistry-climate models (CCMs)³. This multi-model mean ozone dataset was
47 obtained from CCMVal-2 projections run under the SRES A1b scenario for well-
48 mixed greenhouse gases, in contrast to the representative concentration pathway
49 (RCP) scenarios used in CMIP5. To date, research on the impacts of contrasting
50 representations of stratospheric ozone has focused on regional effects, such as the
51 influence of possible future Antarctic ozone recovery on the position of the Southern
52 Hemisphere mid-latitude jet^{4,7}. However, its potential effect on the magnitude of
53 projected global warming has not received much attention.

54 Here, we present evidence which highlights that stratospheric chemistry-
55 climate feedbacks can exert a more significant influence on global warming
56 projections than has been suggested⁸. For a specific climate change experiment, we
57 show that the choice of how to represent key stratospheric chemical species alone
58 can result in a 20% difference in simulated global mean surface warming. Therefore,
59 a treatment of ozone that is not internally consistent with a particular model or
60 greenhouse gas scenario, as is the case for some CMIP5 simulations, could
61 introduce a significant bias into climate change projections.

62 The model used here is a HadGEM3-AO configuration of the UK Met Office's
63 Unified Model⁹ coupled to the UKCA stratospheric chemistry scheme¹⁰ (see
64 Methods). This comprehensive model set-up allows us to study complex feedback
65 effects between the atmosphere, land surface, ocean and sea-ice.

66 Fig. 1 shows the evolution of global and annual mean surface temperature
67 anomalies (ΔT_{surf}) from eight different climate integrations, two of which were carried
68 out with interactive stratospheric chemistry and six with different prescribed monthly-
69 mean fields of the following chemically and radiatively active gases: ozone, methane
70 and nitrous oxide (see Table 1 for details). Experiments with label A are pre-
71 industrial control runs. Experiment B is an abrupt4xCO₂ run with fully interactive
72 chemistry, and experiments labelled C are non-interactive abrupt4xCO₂ runs in
73 which the chemical fields were prescribed at pre-industrial levels. We conducted two
74 versions of each non-interactive experiment to test the effect of using zonal mean
75 fields (label 2, e.g. A2) instead of full 3D fields (label 1, e.g. A1). The time
76 development of ΔT_{surf} shows a clear difference of nearly 20% between the
77 abrupt4xCO₂ experiments B and C1/C2, indicating a much larger global warming in

78 C1/C2 as a consequence of missing composition feedbacks. The primary driver of
79 these differences is changing ozone, with methane and nitrous oxide making much
80 smaller contributions, see below. Fields averaged over the final 50 years of the
81 interactive experiment B were imposed from the beginning in the abrupt4xCO₂
82 experiments B1 and B2. These simulations show a close agreement with experiment
83 B in terms of ΔT_{surf} , implying that the global mean energy budget can be
84 comparatively well-reproduced with this treatment of composition changes, despite
85 the neglect of transient changes in their abundances.

86 We apply the linear regression methodology for diagnosing climate forcing
87 and feedbacks established by Gregory *et al.*¹¹ (see also Methods) to investigate the
88 sources of the differences between the abrupt4xCO₂ experiments with and without
89 the effects of interactive chemistry included. The method assumes a linear
90 relationship between the change in global and annual mean radiative imbalance at
91 the top of the atmosphere (TOA) and ΔT_{surf} . It has been shown to capture well the
92 response of models to many types of climate forcing^{11,12}. The slope obtained from
93 the regression is defined as the climate feedback parameter, α (Wm⁻²°C⁻¹). It
94 represents a characteristic quantity of a given model system, since its magnitude
95 approximates the ΔT_{surf} response to a radiative forcing introduced to the system. Fig.
96 2a shows the Gregory regression plot for each of the 75 years after the initial abrupt
97 4xCO₂ forcing is imposed. The slopes diagnosed for the chemically-similar
98 experiments B, B1 and B2 differ only slightly, however, in C1 and C2, which use the
99 pre-industrial ozone climatologies, there is a significant decrease in the magnitude of
100 α by ~20%, consistent with the larger ΔT_{surf} response. The prescribed chemical fields
101 drive the difference between experiments B1/B2 and C1/C2, so that the fundamental

102 difference in how the modelled climate system responds to the CO₂ forcing must be
103 connected to the changes in atmospheric composition and related further feedbacks.

104 To further investigate the differences, we decompose the TOA radiative fluxes
105 into clear-sky (CS) and cloud radiative effect (CRE) components. In addition, we
106 separate them further into shortwave (SW) and longwave (LW) contributions,
107 producing four components in total (see Methods)¹². Fig. 2b and 2c show Gregory
108 regressions for the two components found to be responsible for the majority of the
109 difference in α , namely the CS-LW ($\alpha_{cs,lw}$) and the CRE-LW ($\alpha_{cre,lw}$) components (see
110 Supplementary Fig. S1 for the smaller changes in the SW components). The
111 differences in $\alpha_{cs,lw}$ between B and C1/C2 are of the same sign as those for α , but
112 larger in magnitude, whereas the change in $\alpha_{cre,lw}$ is of the opposite sign and smaller
113 in magnitude.

114 The reasons for the changes in the CS-LW contribution to α can be
115 understood from the impact of the decrease in tropical and subtropical lower
116 stratospheric ozone between experiment A (and, by definition C1/C2) and B (Fig.
117 3a), which mainly arises as a result of an accelerated Brewer-Dobson circulation
118 (BDC, Supplementary Fig. S2), a ubiquitous feature in climate model projections
119 under increased atmospheric CO₂ concentrations^{4,13}. The increase in middle and
120 upper stratospheric ozone due to the slowing of catalytic ozone depletion cycles¹⁴
121 under CO₂-induced cooling¹⁵ of the stratosphere is also well understood. The local
122 decrease in ozone induces a significant cooling of the lower and middle tropical
123 stratosphere of up to 3.5°C in experiment B relative to C1 (Fig. 3b). An important
124 feedback resulting from this decrease in tropical tropopause temperature is a relative
125 drying of the stratosphere by ~4 ppmv in experiment B compared to C1/C2

126 (Supplementary Fig. S3). Since stratospheric water vapour is a greenhouse gas, this
127 amplifies the tropospheric cooling due to the tropical and subtropical decreases in
128 lower stratospheric ozone, and thus also contributes to changes in α (refs 16,17).

129 It is well-known that composition changes can modify the radiative balance of
130 the atmosphere. However, our results demonstrate that the choice of how to include
131 stratospheric composition feedbacks in climate models can be of first order
132 importance for projections of global climate change. We diagnose radiative effects
133 due to the differences in ozone and stratospheric water vapour between B and C1 of
134 -0.68 Wm^{-2} and -0.78 Wm^{-2} , respectively (see also Methods and Supplementary
135 Figure S4). The magnitude of this effect is related to the strong dependency of the
136 LW radiative impact of ozone and stratospheric water vapour changes on their
137 latitudinal and vertical structure. For instance, the low temperatures in the tropical
138 upper troposphere and lower stratosphere (UTLS) make ozone changes in this
139 region particularly important for the global energy budget^{18,19}. Consequently, climate
140 models need to capture ozone changes here realistically; the tropical UTLS is a
141 crucially sensitive region for climate models. However, trends in tropical tropopause
142 height under climate change differ between models and depend on the forcing
143 scenario²⁰. This suggests a potential mismatch between vertical temperature and
144 prescribed ozone profiles in climate models which do not calculate ozone
145 interactively. Such a mismatch would not only affect the direct radiative impact of
146 ozone, but could also trigger inconsistent local heating or cooling in the cold trap
147 region, which is crucial for the magnitude of the stratospheric water vapour feedback.

148 The magnitude of the overall feedback is expected to be strongly model-
149 dependent, see for example the study by Dietmüller *et al.* (ref. 8) with a less well

150 resolved stratosphere. The simulated BDC (and thus ozone) trends are closely
151 related to the degree of tropospheric warming (ref. 21), which differs between
152 models. The exact scaling of the ozone and water vapour response with tropospheric
153 warming, in turn, will depend on other model-dependent factors, including the
154 representation of gravity waves, the representation of the stratosphere, tropopause
155 dehydration, lightning NO_x , other Earth system feedbacks, as well as the model base
156 state²². Prescribing an ozone field which is neither consistent with the model nor with
157 the forcing scenario, as in some CMIP5 experiments, will also lead to an inconsistent
158 representation of the feedback. Consequently, further modelling studies are needed
159 to investigate how such inter-model differences affect the magnitude of this feedback
160 among a range of models.

161 The UTLS ozone changes are also key to understanding the differences in
162 $\alpha_{\text{cre,lw}}$ (Fig. 2c). To isolate the dominant changes from 50°N to 50°S, we use regional
163 Gregory regressions (Supplementary Fig. S5; ref. 23). We find a significant increase
164 in UTLS cirrus clouds in this region in B compared with C1 (Fig. 4 and
165 Supplementary Fig. S6), in agreement with the sensitivity of cirrus cloud formation to
166 atmospheric temperature (Fig. 3b; ref. 24). This reduces the magnitude of the
167 negative $\alpha_{\text{cre,lw}}$ in B compared to C1, consistent with the effects of high-altitude cirrus
168 clouds on the LW energy budget²⁴⁻²⁶. More studies are needed to quantify how this
169 effect could add to the large uncertainty in cloud feedbacks found in state-of-the-art
170 climate models^{12,24-26}. However, we highlight the large range in the magnitude of
171 $\alpha_{\text{cre,lw}}$ arising as a result of varying the treatment of ozone. This has obvious
172 implications for studies in which cloud feedbacks are compared between models
173 irrespective of their representation of stratospheric chemistry^{1,2,12}.

174 In conclusion, our results demonstrate the potential for considerable sensitivity
175 of global warming projections to the representation of stratospheric composition
176 feedbacks. We highlight the tropical UTLS as a key region for further study and
177 emphasize the need for similar studies; including other climate feedbacks and their
178 interactions in increasingly sophisticated Earth system models. Our results imply that
179 model- and scenario-consistent representations of ozone are required, in contrast to
180 the procedure applied widely in climate change assessments. These include
181 quadruple CO₂ experiments, where changes in ozone are often not considered, as
182 well as other CMIP5 and GeoMIP integrations where the majority of models specified
183 inconsistent ozone changes. We note that further increasing model resolution will not
184 address this fundamental issue. Consequently, we see a pressing need to invest
185 more effort into producing model- and scenario-specific ozone datasets, or to move
186 to a framework in which all participating models explicitly represent atmospheric
187 chemical processes.

188

189

190 **Methods**

191 **Model set-up**

192 A version of the recently developed atmosphere-ocean coupled configuration of the
193 Hadley Centre Global Environment Model version 3 (HadGEM3-AO) from the United
194 Kingdom Met Office has been employed here⁹. It consists of three submodels,
195 representing the atmosphere plus land surface, ocean and sea-ice.

196 For the atmosphere, the Met Office's Unified Model (MetUM) version 7.3 is
197 used. The configuration used here is based on a regular grid with a horizontal
198 resolution of 3.75° longitude by 2.5° latitude and comprises 60 vertical levels up to a
199 height of ~84 km, and so includes a full representation of the stratosphere. Its
200 dynamical core is non-hydrostatic and employs a semi-Lagrangian advection
201 scheme. Subgridscale features such as clouds and gravity waves are parameterised.

202 The ocean component is the Nucleus for European Modelling of the Ocean
203 (NEMO) model version 3.0 coupled to the Los Alamos sea ice model CICE version
204 4.0. It contains 31 vertical levels reaching down to a depth of 5 km. The NEMO
205 configuration used in this study deploys a tripolar, locally anisotropic grid which has
206 2° resolution in longitude everywhere, but an increased latitudinal resolution in
207 certain regions with up to 0.5° in the tropics.

208 Atmospheric chemistry is represented by the United Kingdom Chemistry and
209 Aerosols (UKCA) model in an updated version of the detailed stratospheric chemistry
210 configuration¹⁰ which is coupled to the MetUM. A simple tropospheric chemistry
211 scheme is included which provides for emissions of 3 chemical species and
212 constrains surface mixing ratios of 6 further species. This includes the surface mixing
213 ratios of nitrous oxide (280 ppbv) and methane (790 ppbv), which effectively keeps
214 their concentrations in the troposphere constant at approximately pre-industrial
215 levels. Changes in photolysis rates in the troposphere and the stratosphere are
216 calculated interactively using the Fast-JX photolysis scheme²⁷.

217 **Linear climate feedback theory**

218 The theory is based on the following equation described by Gregory *et al.*¹¹

$$219 \quad N = F + \alpha \Delta T_{\text{surf}}$$

220 where N is the change in global mean net TOA radiative imbalance (Wm^{-2}), F the
 221 effective forcing (Wm^{-2}), ΔT_{surf} the global-mean surface temperature change ($^{\circ}\text{C}$),
 222 and α the climate feedback parameter ($\text{W m}^{-2} \text{ } ^{\circ}\text{C}^{-1}$). Thus, α can be obtained by
 223 regressing N as a function of time against ΔT_{surf} relative to a control climate. Here,
 224 the positive sign convention is used, meaning that a negative α implies a stable
 225 climate system. The theory assumes that the net climate feedback parameter can be
 226 approximated by a linear superposition of processes which contribute to the overall
 227 climate response to an imposed forcing. This can be expressed in form of a linear
 228 decomposition of the α parameter into process-related parameters

$$229 \quad \alpha = \sum \lambda_i$$

230 with λ_i for example being $\lambda_{\text{water feedback}}$, λ_{clouds} etc. Similarly, one can decompose the
 231 climate feedback parameter into separate radiative components^{12,23,25}

$$232 \quad \alpha = \alpha_{\text{cs}} + \alpha_{\text{cre}} = \alpha_{\text{cs,sw}} + \alpha_{\text{cs,lw}} + \alpha_{\text{cre,sw}} + \alpha_{\text{cre,lw}}$$

233 providing individual shortwave (SW) and longwave (LW) components for clear-sky
 234 (CS) radiative fluxes and the cloud radiative effect (CRE). In this method, the CRE
 235 contains direct cloud radiative effects and indirect cloud masking effects, e.g. due to
 236 persistent cloud cover which masks surface albedo changes in the all-sky
 237 calculation^{25,26}.

238 Radiative Transfer Experiments

239 The radiative transfer calculations were carried out using a version of the Edwards
 240 and Slingo²⁸ offline radiative transfer code updated to use the correlated-k method

241 for calculating transmittances²⁹. This is identical to the radiation code used in the
242 coupled model simulations. The inferred all-sky radiative effects due to the changes
243 in ozone and stratospheric water vapour between experiments B and C1 were
244 diagnosed using a base climatology (temperature, pressure, humidity etc.) taken
245 from the last 50 years of C1 and perturbing around this state with the B minus C1
246 ozone or stratospheric water vapour fields over the same time period. The
247 calculations employ the fixed dynamical heating (FDH) method¹⁵, in which
248 stratospheric temperatures are adjusted to re-establish radiative equilibrium in the
249 presence of the imposed perturbation (see ref. 30 for details). The radiative forcing is
250 then diagnosed as the imbalance in the total (LW+SW) net (down minus up)
251 tropopause fluxes. Note that the changes in ozone and stratospheric water vapour
252 described in the study could be considered as a part forcing and part climate
253 feedback. For example, the increase in ozone in the mid and upper stratosphere in
254 Fig. 3a is linked to the CO₂ induced cooling at these levels, and may therefore not be
255 strongly correlated with surface temperature change. In contrast, the decrease in
256 ozone in the tropical mid- and lower-stratosphere is driven by the strengthening in
257 the Brewer-Dobson circulation, which is more closely linked to tropospheric
258 temperature change²¹. However, for the purposes of quantifying the radiative
259 contribution of the composition changes to the evolution of global climate in the
260 experiments, we impose them diagnostically in the offline code as a pseudo radiative
261 forcing agent.

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343

344 **Additional information**

345 Supplementary information is available in the online version of the paper. Reprints
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347 Correspondence and requests for materials should be addressed to P.J.N.

348

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361

362 Author contributions

363 P.J.N. conducted the research on a day-to-day basis; the model was developed by
364 N.L.A., J.M.G., M.M.J. and A.O.; N.L.A. and P.B. designed the initial experiment and
365 its subsequent evolution; major analysis and interpretation of results was performed
366 by P.J.N. and A.C.M.; P.J.N. led the paper writing, supported by A.C.M.; N.L.A., P.B.
367 and J.A.P. all contributed to the discussion and interpretation of results and write-up;
368 J.A.P. suggested the study.

369

370 Competing financial interests

371 The authors declare no competing financial interests.

372

373 Captions of Figures

374

375 Figure 1 | Temporal evolution of the annual and global mean surface

376 **temperature anomalies.** All anomalies ($^{\circ}\text{C}$) are shown relative to the average
377 temperature of experiment A. Solid lines show the interactive chemistry runs (A, B),
378 dashed lines the 3D climatology experiments (A1, B1, C1) and dotted lines the 2D
379 climatology experiments (A2, B2, C2). For clarity, lines for the abrupt4xCO₂
380 experiments start after year one so that they are not joined with those of the
381 corresponding control experiments. The last 50 years of the abrupt4xCO₂
382 experiments are highlighted in the inset panel with the straight lines marking the
383 average temperature in each set of experiments over the last 20 years.

384

385 **Figure 2 | Gregory regression plots. a**, For all radiative components, giving an

386 ~25% larger climate feedback parameter, α , in C1/C2 than in B. **b**, **c**, For the CS-

387 LW and CRE-LW components only. In particular in **c**, a clear evolution of the
388 atmospheric state B is observable as it starts off very close to C1 and C2 and
389 evolves towards B1 and B2. Radiative fluxes follow the downward sign convention
390 so that all negative (positive) changes in α imply a cooling (warming) effect. The
391 inset tables give the correlation coefficient (R_{corr}) and the α parameter obtained from
392 each regression.

393

394 **Figure 3 | Annual and zonal mean differences in ozone and temperature.** Shown
395 are averages over the last 50 years of each experiment. **a**, The percentage
396 differences in ozone between simulations B and A. By definition, these are identical
397 to the differences in the climatologies between B/B1/B2 and C1/C2/A/A1/A2. Note
398 that the climatologies of experiments B1/B2 and other 2D and 3D versions of each
399 set of experiment are only identical after zonal averaging. **b**, The absolute
400 temperature anomaly ($^{\circ}\text{C}$) between experiments B and C1. Apart from some areas
401 around the tropopause (hatched out), all differences in **b** are statistically significant at
402 the 95% confidence level using a two-tailed Student's t-test.

403

404 **Figure 4 | Cirrus cloud changes.** Zonal and annual mean frozen cloud fraction per
405 unit volume multiplied by factor 100 in the region 50°N - 50°S where the deviations in
406 $\alpha_{\text{cre,lw}}$ are found. The shading shows the difference B minus C1 averaged over the
407 last 50 years of both experiments. Contour lines (interval 2.5) denote the climatology
408 of C1. Note that the tropical cloud fraction increases at ~ 12 - 13 km mainly result from
409 the relatively warmer climate in C1. They therefore do not change $\alpha_{\text{cre,lw}}$, in contrast
410 to the increases in the UTLS, see also Figure S6. Non-significant differences (using

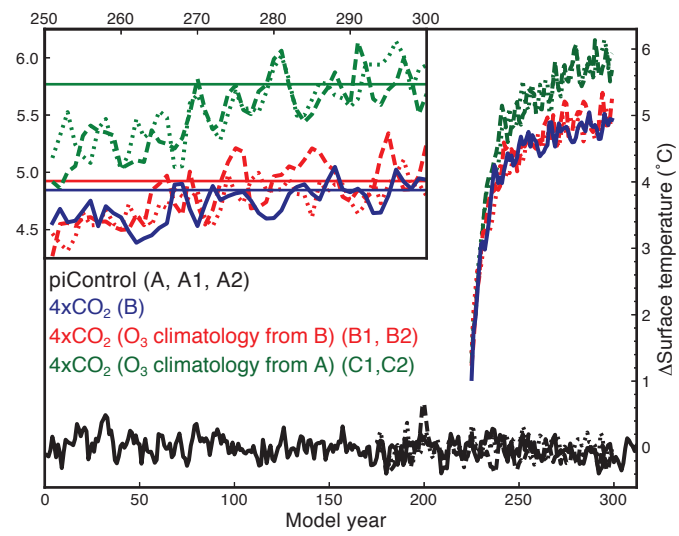
411 a two-tailed Student's t-test at the 95% confidence level or where the cloud fraction
 412 in both experiments is smaller than 5‰) are hatched out.

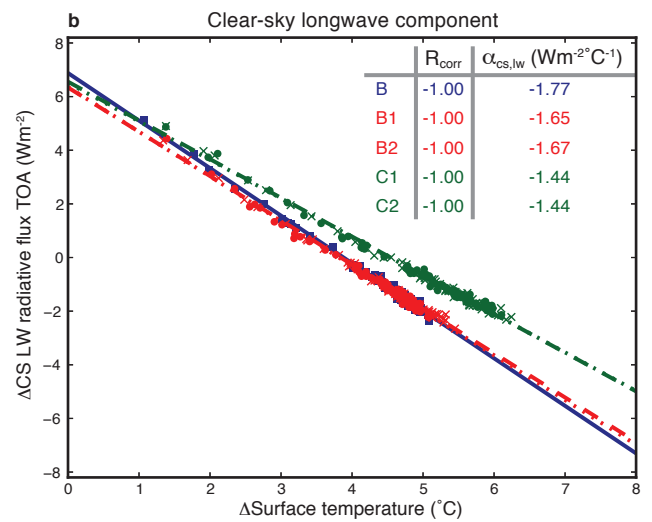
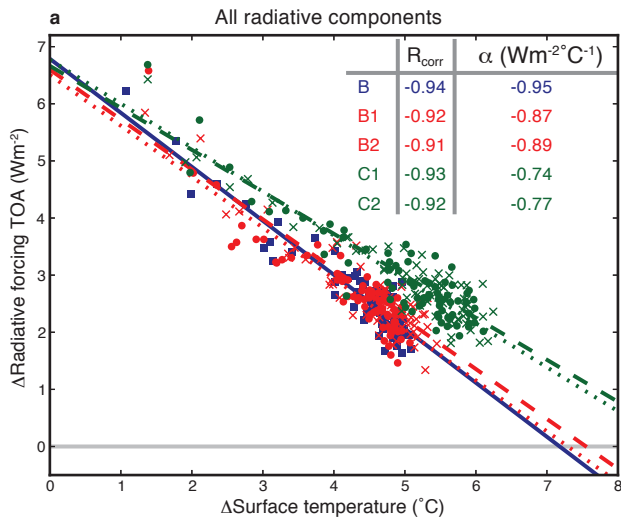
413

414 **Table 1 | Overview of the experiments.**

Experiment	Description	Initial Condition	Chemistry
A	piControl, (285 ppmv CO ₂)	Initialised from 900 year spin-up	Interactive
A1	piControl-1, (285 ppmv CO ₂)	Initialised from A (year 175)	Non-interactive, 3D climatologies from A
A2	piControl-2, (285 ppmv CO ₂)	Initialised from A (year 175)	Non-interactive, 2D climatologies from A
B	abrupt4xCO ₂ (1140 ppmv CO ₂)	Initialised from A (year 225)	Interactive
B1	abrupt4xCO ₂ (1140 ppmv CO ₂)	Initialised from A1 (year 50)	Non-interactive, 3D climatologies from B
B2	abrupt4xCO ₂ (1140 ppmv CO ₂)	Initialised from A2 (year 50)	Non-interactive, 2D climatologies from B
C1	abrupt4xCO ₂ (1140 ppmv CO ₂)	Initialised from A1 (year 50)	Non-interactive, 3D climatologies from A
C2	abrupt4xCO ₂ (1140 ppmv CO ₂)	Initialised from A2 (year 50)	Non-interactive, 2D climatologies from A

415 Climatologies for the non-interactive runs represent the seasonal cycle on a monthly-
 416 mean basis. 3D climatologies contain chemical fields of the most important
 417 radiatively active species (ozone, methane, and nitrous oxide) for all spatial
 418 dimensions (longitude, latitude, altitude). For 2D climatologies these fields were
 419 averaged over all longitudes, as it is commonly done for ozone climatologies used in
 420 non-interactive climate integrations^{3,5}.





$4\times\text{CO}_2$

$4\times\text{CO}_2$ (O_3 climatology from B)

$4\times\text{CO}_2$ (O_3 climatology from A)

B

B1

B2

C1

C2

