

# Equilibrium climate sensitivity estimated by equilibrating climate models

Maria Rugenstein<sup>1,2\*</sup>, Jonah Bloch-Johnson<sup>3</sup>, Jonathan Gregory<sup>3,4</sup>, Timothy  
Andrews<sup>4</sup>, Thorsten Mauritsen<sup>5</sup>, Chao Li<sup>2</sup>, Thomas L. Frölicher<sup>6,7</sup>, David  
Paynter<sup>8</sup>, Gokhan Danabasoglu<sup>9</sup>, Shuting Yang<sup>10</sup>, Jean-Louis Dufresne<sup>11</sup>, Long  
Cao<sup>12</sup>, Gavin A. Schmidt<sup>13</sup>, Ayako Abe-Ouchi<sup>14</sup>, Olivier Geoffroy<sup>15</sup>, and Reto  
Knutti<sup>1</sup>

<sup>1</sup>Institute for Atmospheric and Climate Science, ETH Zurich, CH-8092 Zurich, Switzerland

<sup>2</sup>Max-Planck-Institute for Meteorology, Bundestrasse 53, 20146 Hamburg, Germany

<sup>3</sup>NCAS, University of Reading, Reading

<sup>4</sup>UK Met Office Hadley Centre, FitzRoy Road, Exeter, EX1 3PB

<sup>5</sup>Stockholm University, SE-106 91 Stockholm, Sweden

<sup>6</sup>Climate and Environmental Physics, Physics Institute, University of Bern, Switzerland

<sup>7</sup>Oeschger Centre for Climate Change Research, University of Bern, Switzerland

<sup>8</sup>Geophysical Fluid Dynamics Laboratory, Princeton, New Jersey, USA

<sup>9</sup>National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307

<sup>10</sup>Danish Meteorological Institute, Lyngbyvej 100, DK-2100 Copenhagen, Denmark

<sup>11</sup>LMD/IPSL, Centre National de la Recherche Scientifique, Sorbonne Université, École Normale  
Supérieure, PSL Research University, École Polytechnique, Paris, France

<sup>12</sup>School of Earth Sciences, Zhejiang University, Hang Zhou, Zhejiang Province, 310027, China

<sup>13</sup>NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY 10025

<sup>14</sup>Atmosphere and Ocean Research Institute, The University of Tokyo, Japan

<sup>15</sup>Centre National de Recherches Météorologiques, Université de Toulouse, Météo-France, CNRS,  
Toulouse, France

## Key Points:

- 27 simulations of 15 general circulation models are integrated to near equilibrium
- All models simulate a higher equilibrium warming than predicted by using extrapolation methods
- Tropics and mid-latitudes dominate the change of the feedback parameter on different timescales

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\*Max-Planck-Institute for Meteorology, Bundestrasse 53, 20146 Hamburg, Germany

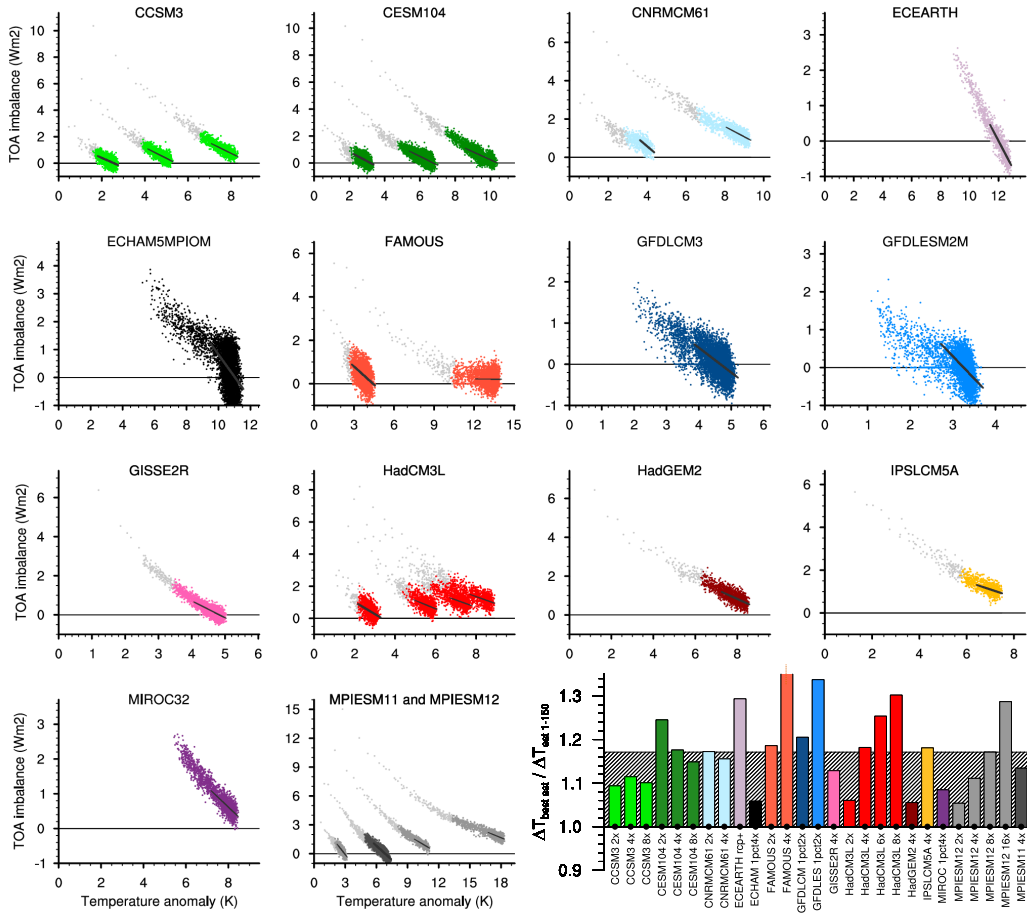
Corresponding author: Maria Rugenstein, [maria.rugenstein@mpimet.mpg.de](mailto:maria.rugenstein@mpimet.mpg.de)

**Abstract**

The methods to quantify equilibrium climate sensitivity are still debated. We collect millennial-length simulations of coupled climate models and show that the global mean equilibrium warming is higher than those obtained using extrapolation methods from shorter simulations. Specifically, 27 simulations with 15 climate models forced with a range of CO<sub>2</sub> concentrations show a median 17% larger equilibrium warming than estimated from the first 150 years of the simulations. The spatial patterns of radiative feedbacks change continuously, in most regions reducing their tendency to stabilizing the climate. In the equatorial Pacific, however, feedbacks become more stabilizing with time. The global feedback evolution is initially dominated by the tropics, with eventual substantial contributions from the mid-latitudes. Time-dependent feedbacks underscore the need of a measure of climate sensitivity that accounts for the degree of equilibration, so that models, observations, and paleo proxies can be adequately compared and aggregated to estimate future warming.

**1 Estimating equilibrium climate sensitivity**

The equilibrium climate sensitivity (ECS) is defined as the global- and time-mean, surface air warming once radiative equilibrium is reached in response to doubling the atmospheric CO<sub>2</sub> concentration above pre-industrial levels. It is by far the most commonly and continuously applied concept to assess our understanding of the climate system as simulated in climate models and it is used to compare models, observations, and paleo-proxies (Knutti et al., 2017; Charney et al., 1979; Houghton et al., 1990; Stocker, 2013). Due to the large heat capacity of the oceans, the climate system takes millennia to equilibrate to a forcing, but performing such a long simulation with a climate model is often computationally not feasible. As a result, many modeling studies use extrapolation methods on short, typically 150-year long, simulations to project equilibrium conditions (Taylor et al., 2011; Andrews et al., 2012; Collins et al., 2013; Otto et al., 2013; Lewis & Curry, 2015; Andrews et al., 2015; Forster, 2016; Calel & Stainforth, 2017). These so-called *effective* climate sensitivities (Murphy, 1995; Gregory et al., 2004) are often reported as ECS values (Hargreaves & Annan, 2016; Tian, 2015; Brient & Schneider, 2016; Forster, 2016). Research provides evidence for decadal-to-centennial changes of feedbacks (e.g., Murphy (1995); Senior and Mitchell (2000); Gregory et al. (2004); Winton et al. (2010); Armour et al. (2013); Proistosescu and Huybers (2017); Paynter et al. (2018)) but the behavior on longer timescales has



**Figure 1.** Evolution of global and annual mean top of the atmosphere (TOA) imbalance and surface temperature anomalies (14 small panels). The first 150 years of step forcing simulations are depicted in light gray. For experiments which are not step forcing simulations only the period after stabilizing  $\text{CO}_2$  concentrations is shown. The black line shows the linear regression of TOA imbalance and surface warming for the last 15% of warming. The panel on the lower right shows the ratio  $\Delta T_{\text{best est}} / \Delta T_{\text{est } 1-150}$ , see text for definitions. A dot at the lower end of the bar indicates with 90% confidence that  $\Delta T_{\text{best est}}$  and  $\Delta T_{\text{est } 1-150}$  obtained by resampling 10,000 times do not overlap. The gray hashed bar in the background is the median of all simulations (1.17). FAMOUS *abrupt4x* ends outside of the depicted range at 1.53. Table 1 specifies the model versions and names, length of simulations, and numerical values for different climate sensitivity estimates.

62 not been compared among models. Here, we utilize LongRunMIP, a large set of millennia-  
 63 long coupled general circulations models (GCMs) to estimate the true equilibrium warming,  
 64 study the centennial-to-millennial behavior of the climate system under elevated radiative  
 65 forcing, and test extrapolation methods. LongRunMIP is a model intercomparison project  
 66 (MIP) of opportunity in that its initial contributions were preexisting simulations, without  
 67 a previously agreed upon protocol. The minimum contribution is a simulation of at least  
 68 1000 years with a constant CO<sub>2</sub> forcing level. The collection consists mostly of doubling or  
 69 quadrupling step forcing simulations (“abrupt2x”, “abrupt4x”, ...) as well as annual incre-  
 70 ments of 1% CO<sub>2</sub> increases reaching and sustaining doubled or quadrupled concentrations  
 71 (“1pct2x”, “1pct4x”). Table 1 lists the simulations and models used here, while M. Rugen-  
 72 stein et al. (2019) documents the entire modeling effort and each contribution in detail.

73 The equilibration of top of the atmosphere (TOA) radiative imbalance and surface  
 74 temperature anomaly of the simulations are depicted in Fig. 1. Throughout the manuscript,  
 75 we show anomalies as the difference to the mean of the unforced control simulation with  
 76 pre-industrial CO<sub>2</sub> concentrations. Light gray dots indicate annual means of the first 150  
 77 years of a step forcing simulation, requested by the Coupled Model Intercomparison Project  
 78 Phase 5 and 6 protocols (CMIP5 and CMIP6; Taylor et al. (2011); Eyring et al. (2016))  
 79 and widely used to infer ECS (Andrews et al., 2012; Geoffroy, Saint-Martin, Olivié, et al.,  
 80 2013). We refer to this timescale as “decadal to centennial”. Colors indicate the “centen-  
 81 nial to millennial” timescale we explore here. The diminishing distances to the reference  
 82 line at TOA = 0 indicate that most simulations archive near-equilibrium by the end of the  
 83 simulations. However, even if a simulation has an equilibrated TOA imbalance of near zero,  
 84 the surface temperature, surface heat fluxes, or ocean temperatures can still show a trend  
 85 (discussed in M. Rugenstein et al. (2019)).

86 Throughout the manuscript, we use “ $\Delta T_{[specification]}$ ” for a true or estimated equilib-  
 87 rium warming, for a range of forcing levels not only CO<sub>2</sub> doubling (Table 1). We define the  
 88 best estimate of equilibrium warming,  $\Delta T_{\text{best est}}$ , as the temperature-axis intersect of the  
 89 regression of annual means of TOA imbalance and surface temperature anomaly over the  
 90 simulations’ final 15% of global mean warming (black lines in Fig. 1). The lower right panel  
 91 in Fig. 1 illustrates that all simulations eventually warm significantly more (measured by  
 92  $\Delta T_{\text{best est}}$ ) than predicted by the most commonly used method to estimate the equilibrium  
 93 temperature by extrapolating a least-square regression of the first 150 years of the same step  
 94 forcing simulation (Gregory et al., 2004; Flato et al., 2013), denoted here as “ $\Delta T_{\text{est } 1-150}$ ”.

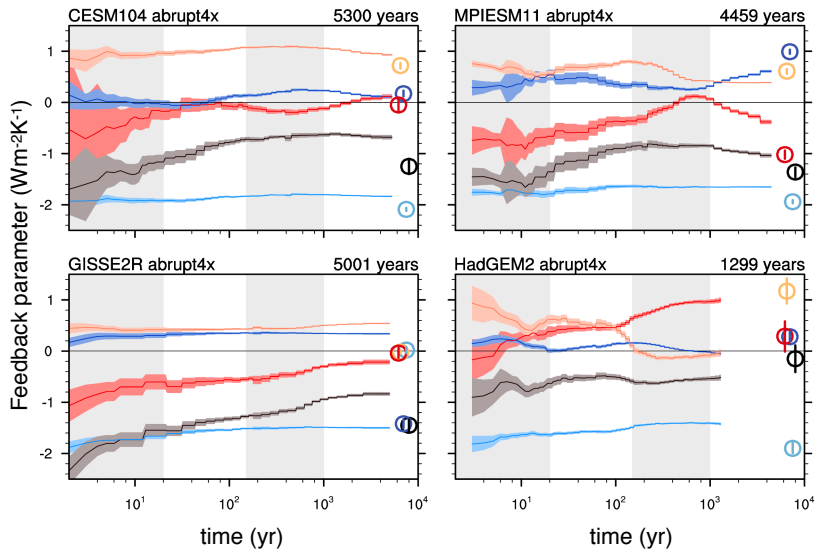
95 For simulations that have gradual forcings (e.g., *1pct2x*), we use 150 year long step forcing  
 96 simulations of the same model to calculate  $\Delta T_{\text{est } 1-150}$ . The median increase of  $\Delta T_{\text{best est}}$   
 97 over  $\Delta T_{\text{est } 1-150}$  is 17% for all simulations and 16% for the subset of CO<sub>2</sub> doubling and qua-  
 98 drupling simulations. While  $\Delta T_{\text{est } 1-150}$  implies a constant feedback parameter (the slope  
 99 of the regression line), other extrapolation methods allow for a time-dependent feedback pa-  
 100 rameter, but still typically underestimate  $\Delta T_{\text{best est}}$ : Using years 20-150 in linear regression  
 101 ( $\Delta T_{\text{est } 20-150}$ ; e.g., Andrews et al. (2015); Armour (2017)) results in a median equilibrium  
 102 warming estimate which is 7% lower than  $\Delta T_{\text{best est}}$ , both for all simulations and the subset  
 103 of CO<sub>2</sub> doubling and quadrupling. The two-layer model including ocean heat uptake efficacy  
 104 ( $\Delta T_{\text{EBM}-\epsilon}$ ; e.g., Winton et al. (2010); Geoffroy, Saint-Martin, Bellon, et al. (2013)) results  
 105 in a multi model median equilibrium warming estimate which is 9% lower than  $\Delta T_{\text{best est}}$ ,  
 106 again both for all simulations and the subset of CO<sub>2</sub> doubling and quadrupling. Both meth-  
 107 ods are described and illustrated in the supplemental material.

108  $\Delta T_{\text{best est}}$  of any forcing level can be scaled down to doubling CO<sub>2</sub> levels to estimate  
 109 equilibrium warming for CO<sub>2</sub> doubling. We do so by assuming that the temperature scales  
 110 with the forcing level, which depends logarithmically on the CO<sub>2</sub> concentration (Myhre et  
 111 al., 1998), and assuming no feedback temperature dependence (e.g. Mauritsen et al. (2018)  
 112 and Rohrschneider et al. (2019), see discussion below). The estimate of equilibrium warm-  
 113 ing for CO<sub>2</sub> doubling range from 2.42 to 5.83 K (excluding FAMOUS *abrupt4x* at 8.55K;  
 114 see Table 1 and Fig. 1). Note that the simulation *abrupt4x* of the model FAMOUS warms  
 115 anomalously strongly. As this simulation represents a physically possible result, we do not  
 116 exclude it from the analysis (see more details in SM Section 4). The results are qualitatively  
 117 the same if  $\Delta T_{\text{best est}}$  is defined by regressing over the last 20% instead of 15% of warming  
 118 or instead time averaging the surface warming toward the end of every simulation without  
 119 taking the information of the TOA imbalance into account. SM Section 1 discusses different  
 120 options and choices to determine  $\Delta T_{\text{best est}}$ .

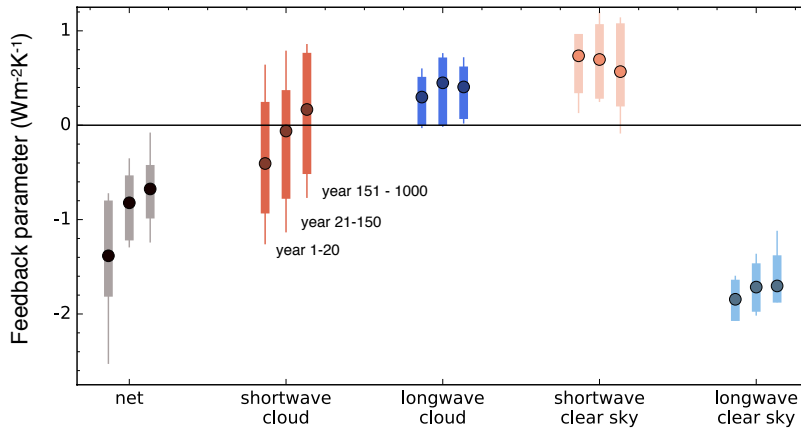
## 121 2 Global feedback evolution

122 Current extrapolation methods underestimate the equilibrium response because climate  
 123 feedbacks change with the degree of equilibration (Murphy, 1995; Senior & Mitchell, 2000;  
 124 Andrews et al., 2015; Knutti & Rugenstein, 2015; M. A. A. Rugenstein, Caldeira, & Knutti,  
 125 2016; Armour, 2017; Proistosescu & Huybers, 2017; Paynter et al., 2018). We define the  
 126 global net TOA feedback as the *local tangent* in temperature-TOA space ( $\delta\text{TOA}/\delta T$ ) com-

a) Time evolution of feedbacks in four models



b) Feedback components for different time periods

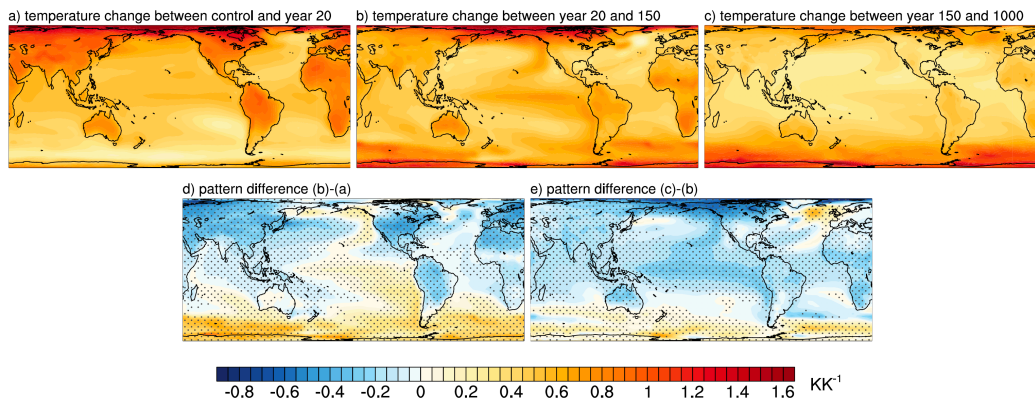


**Figure 2.** a) Time evolution of global feedbacks in four characteristic models. Net TOA feedback (gray) is the sum of its components: the cloud effects in the shortwave (red) and longwave (blue), and clear sky feedbacks in the shortwave (salmon) and longwave (light blue). Circles at the right of each panel indicate the feedbacks arising from internal variability; shading and vertical lines shows the 2.5-97.5% confidence intervals. Panel titles give the model name and length of the simulation. Time periods of 1-20 years and 150-1000 years are shaded gray. (b) Feedback evolution in the step forcing simulations of CCSM3, CESM104, CNRMCM6, ECHAM5MPIOM, FAMOUS, GISS2R, HadCM3L, HadGEM2, IPSLCM5A, MPIESM11, and MPIESM12, see Table 1 for naming convention. Lines show all simulations, dots represent median values and bars spans all but the two highest and two lowest simulations. SM Fig.4 and 5 show the feedback evolution for all available simulations.

127    puted by a least square regression of all global and annual means of netTOA imbalance and  
128    surface temperature anomaly within a temperature bin, which is moved in steps of 0.1 K  
129    throughout the temperature space to obtain the continuous local slope of the point cloud  
130    (sketched out in SM Fig. 2a). We decompose the net TOA imbalance into its clear sky and  
131    cloud radiative effects (CRE; e.g., Wetherald and Manabe (1988); Soden and Held (2006);  
132    Ceppi and Gregory (2017)) in the shortwave and longwave (Fig. 2a). The feedbacks change  
133    continuously – not on obviously separable timescales – in some models more at the begin-  
134    ning of the simulations (e.g., CESM104), in some models after 150 years (e.g., GISS2R) or,  
135    in some models, intermittently throughout the simulation (e.g., MPIESM11 or HadGEM2).  
136    The shortwave CRE dominates the magnitude and the timing of the net feedback change,  
137    and can be counteracted by the longwave CRE. The reduction of the shortwave clear sky  
138    feedback associated with ice albedo, lapse rate, and water vapor is a function of tempera-  
139    ture and occurs on centennial to millennial timescales. Longwave clear sky changes, when  
140    present, contribute to the increase of the sensitivity with equilibration time and temperature.  
141    The net feedback parameter can be composed of a subtle balance of different components at  
142    any time and the forced signal is not obviously linked to the feedback arising from internal  
143    variability, defined by regressing all available annual and global means of TOA imbalance  
144    and surface temperature anomalies (relative to the mean) of the control simulations (circles  
145    in Fig. 2a; Roe (2009); Brown et al. (2014); Zhou et al. (2015); Colman and Hanson (2017)).

146        Models which are more sensitive than other models – have feedbacks which are more  
147    positive – at the beginning of the simulation are generally also more sensitive towards the  
148    end. The model spread in the magnitude of feedbacks does not substantially reduce in time,  
149    while the feedback parameter change varies from negligible to an order of magnitude. We  
150    quantify the continuous changes across models by considering different time periods, namely  
151    years 1-20, 21-150, and 151-1000 (Fig. 2b), in each of which we regress all points. In addition  
152    to the increase of the feedback parameter between years 1-20 and 21-150, which has been  
153    documented for CMIP5 models (Geoffroy, Saint-Martin, Bellon, et al., 2013; Andrews et  
154    al., 2015; Proistosescu & Huybers, 2017; Ceppi & Gregory, 2017), there is a further increase  
155    from centennial to millennial timescales.

156        Previous research has shown that the change in feedbacks over time can come about  
157    through a dependence of feedback processes on the increasing temperature (Hansen et al.,  
158    1984; Jonko et al., 2013; Caballero & Huber, 2013; Meraner et al., 2013; Bloch-Johnson et  
159    al., 2015), due to evolving surface warming patterns and feedback processes (“pattern effect”;



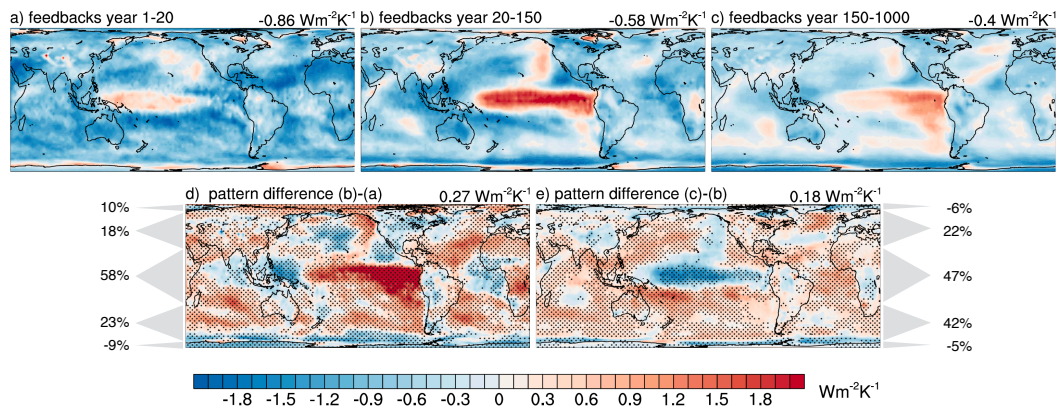
**Figure 3.** Multi-model mean normalized patterns of surface warming (local warming divided by global warming) between the average of (a) the control simulation and year 15-25, (b) year 15-25 and 140-160, (c) year 140-160 and 800-1000, and their differences (d and e) for the same models and simulations as in Fig. 2b. For models contributing several simulations, these are averaged. Stippling in panel d and e indicates that 9 out of 11 models agree in the sign of change.

160 Senior and Mitchell (2000); Winton et al. (2010); Armour et al. (2013); M. A. A. Rugenstein,  
 161 Gregory, et al. (2016); Gregory and Andrews (2016); Haugstad et al. (2017); Paynter et al.  
 162 (2018)), or both at the same time (Rohrshneider et al., 2019). There is no published method  
 163 which clearly differentiates between time/pattern and temperature/state dependence and  
 164 simulations with several forcing levels are needed to disentangle them. The relationship  
 165 between forcing and CO<sub>2</sub> concentrations is a matter of debate (Etminan et al., 2016) and  
 166 further complicates the analysis, as time, temperature, and forcing level dependence might  
 167 compensate to some degree (Gregory et al., 2015). As not all models contributed several  
 168 forcing levels, we focus in the following on robust pattern changes in surface temperatures  
 169 and feedbacks, which occur in most or all simulations irrespective of their overall tempera-  
 170 ture anomaly or forcing level.

### 171 **3 Pattern evolution of surface warming and feedbacks**

172 The evolution of surface warming patterns during the decadal, centennial, and mil-  
 173 lennial periods displays a fast establishment of a land-sea warming contrast, Arctic am-  
 174 plification, and the delayed warming over the Southern Ocean that have been studied on  
 175 annual to centennial timescales (Fig. 3; Senior and Mitchell (2000); Li et al. (2013); Collins  
 176 et al. (2013); Armour et al. (2016)). Arctic amplification does not change substantially,





**Figure 4.** Time evolution of feedback patterns. Model-mean of local contribution to the change in global feedbacks (local TOA anomaly divided by global warming during the period indicated in the panel titles; see text for definitions) (a–c) and their differences (d, e). The global feedback value is shown in the panel title. Regionally aggregated contributions to the global values are indicated with percent numbers and gray triangles ( $22^{\circ}\text{S}$ – $22^{\circ}\text{N}$ ,  $22^{\circ}\text{S}/\text{N}$ – $66^{\circ}\text{S}/\text{N}$ ,  $66^{\circ}\text{S}/\text{N}$ – $90^{\circ}\text{S}/\text{N}$ , representing 40%, 27%, and 4% of the global surface area respectively). Model and simulations selection, weighting, and stippling is the same as in Fig. 3. SM Fig. 6–12 shows all TOA components.

177 whereas Antarctic amplification strengthens by approximately 50% on centennial to millen-  
 178 nial timescales (Salzmann, 2017; M. Rugenstein et al., 2019). The warming in the northern  
 179 North Atlantic reflects the strengthening of the Atlantic meridional overturning circulation,  
 180 after the initial decline (Stouffer & Manabe, 2003; Li et al., 2013; M. A. A. Rugenstein,  
 181 Sedláček, & Knutti, 2016; Rind et al., 2018; Jansen et al., 2018).

182 In the Pacific, at all times, the temperatures in absolute terms are higher in the West  
 183 compared to the East Pacific. The eastern equatorial Pacific warms more than the warm  
 184 pool in most simulations, a phenomenon reminiscent of the positive phase of the El-Niño-  
 185 Southern-Oscillation (ENSO) (“ENSO-like warming” (Song & Zhang, 2014; Andrews et al.,  
 186 2015; Luo et al., 2017; Tierney et al., 2019)). This tendency can last several millennia, but  
 187 significantly reduces or stops in most simulations after a few hundred years. Similar to the  
 188 Equatorial east Pacific, the south east Pacific warms more than the warm pool (Zhou et al.,  
 189 2016; Andrews & Webb, 2018). However, models display a large variance in the timescales  
 190 of warming in these two regions, i.e. the warm pool can initially warm faster or slower than  
 191 the south east Pacific.s Across the Pacific, the change in surface warming pattern is reminis-  
 192 cent of the Interdecadal Pacific Oscillation (IPO; Fig. 3d). In many models, the reduction

193 of the Walker circulation coincides with the decadal to centennial ENSO/IPO-like warming  
194 pattern, but it does not obviously coincide with surface warming pattern changes on the  
195 millennial timescale, indicating that subtropical ocean gyre advection and upwelling play a  
196 more prominent role on longer timescales (Knutson & Manabe, 1995; Song & Zhang, 2014;  
197 Fedorov et al., 2015; Andrews & Webb, 2018; Luo et al., 2017; Zhou et al., 2017; Kohyama  
198 et al., 2017). The mechanisms and spread of model responses in the Pacific are still under  
199 investigation.

200 Feedbacks defined as the local tangent in temperature-TOA space as used in Fig. 2a  
201 contain a signal from both the internal variability and the forced response. In order to  
202 isolate the forced response, we take the difference of the means at the beginning and end of  
203 the time periods discussed above. We call this definition of feedbacks the *finite difference*  
204 *approach*, as it represents a change *across* a time period (SM Fig. 2b). Fig. 4 shows the local  
205 contribution to the global net TOA changes (defined as the local change in TOA imbalance  
206 divided by the global temperature change.) for the same time periods and models as used in  
207 Fig. 3. In the initial years, the atmosphere restores radiative balance through increased ra-  
208 diation to space almost everywhere, except in the western-central Pacific (Fig. 4a), whereas  
209 on decadal to centennial timescales, the structure of the feedbacks mirrors the surface tem-  
210 perature evolution and develops a pattern reminiscent of ENSO/IPO (Fig. 4b). The cloud  
211 response dominates the pattern change, although for CMIP5 models, changes on decadal  
212 and centennial timescales have been attributed to changing lapse rate feedbacks as well (SM  
213 Fig. 6-8 and Andrews et al. (2015); Andrews and Webb (2018); Ceppi and Gregory (2017)).  
214 For the millennial timescales, our models show that feedbacks become less negative almost  
215 everywhere, switching from slightly negative to positive in parts of the Southern Ocean and  
216 North Atlantic region, and become less destabilizing in the Tropical Pacific (Fig. 4c). The  
217 feedback pattern change from decadal to centennial timescales (Fig. 4d) is reversed in many  
218 regions on centennial to millennial timescales (Fig. 4e), particularly in the entire Pacific  
219 basin, the Atlantic, and parts of Asia and North America. This “pattern flip” is dominated  
220 by longwave CRE (SM Fig. 8) and mirrors, in the Pacific, the reduction in ENSO/IPO-like  
221 surface warming patterns discussed for the surface temperature evolution.

222 Note that the local temperature is not part of the calculation of the local contribution  
223 in feedback changes. Due to the far-field effects of local feedbacks (e.g., Rose et al. (2014);  
224 Kang and Xie (2014); M. A. A. Rugenstein, Caldeira, and Knutti (2016); Zhou et al. (2016,  
225 2017); Ceppi and Gregory (2017); Liu et al. (2018); Dong et al. (2019)), the relation between

226 the local feedback contribution (Fig. 4) and the local temperatures (Fig. 3) is not straight-  
227 forward. There is strong correspondence between changes of TOA fluxes and temperature  
228 patterns in the Pacific on decadal to millennial timescales: Stronger (weaker) local warming  
229 coincides with a more positive (negative) local feedback contribution. However, there is  
230 no clear correspondence directly after the application of the forcing, or over land and the  
231 Southern Ocean through time. SM Fig. 13 and 14 show overlays of Fig. 3 and 4 for a better  
232 comparison. A local correspondence does not necessarily indicate a strong local feedback  
233 (i.e. local TOA divided by local surface temperature change), as both the local TOA and  
234 the surface in one region could be forced by another region. A closer investigation of local  
235 and far-field influence of feedbacks is under investigation (Bloch-Johnson et al., in revision).

236 Although the spatial patterns of changing temperature and radiative feedbacks vary  
237 among models, the large scale features discussed here occur robustly across most models  
238 and forcing levels, and also occur in the *1pct2x* and *1pct4x* simulations, which are not  
239 included in the figures.

#### 240 **4 Regions accounting for changing global feedbacks**

241 We quantify the contribution of the tropics, extra-tropics, and polar regions to the  
242 global feedback change (Fig. 4d,e) by adding up all feedback contributions of the respective  
243 areas indicated by the gray triangles and expressing them as percentages of the total. We  
244 note that the total is the global feedback parameter, i.e., the slope of the point clouds in  
245 Fig. 1 which is indicated on the top right of each panel. These percentages reflect the role  
246 played by TOA fluxes in each region, which is not the same as the role played by surface  
247 warming in each region, as noted above. Whereas the tropics account for the bulk of the  
248 change (58% on decadal to centennial and 47% on centennial to millennial timescales), the  
249 mid-latitudes become more important with time (Northern and Southern Hemisphere com-  
250 bined for 41% on decadal to centennial and for 66% on centennial to millennial timescales).  
251 The high latitudes, dominated by the shortwave clear sky feedback (SM Fig. 12), play only  
252 a minor role in influencing the global response at all timescales. The regional accounting  
253 of global feedback changes permits us to test competing explanations regarding the spatial  
254 feedback pattern by placing them in a common temporal framework. Primary regions con-  
255 trolling the global feedback evolution have been suggested to be the Southern Hemisphere  
256 mid to high latitudes (Senior & Mitchell, 2000), the Northern Hemisphere subpolar regions  
257 (Rose & Rayborn, 2016; Trossman et al., 2016), and the Tropics (Jonko et al., 2013; Mer-

258 aner et al., 2013; Block & Mauritsen, 2013; Andrews et al., 2015; Ceppi & Gregory, 2019),  
259 especially in the Pacific (Andrews & Webb, 2018; Ceppi & Gregory, 2017).

260 The simulations robustly shows a delayed warming in the Southern Hemisphere relative  
261 to the Northern Hemisphere throughout the millennia-long integrations, which correlates  
262 with the time evolution of net TOA and shortwave CRE (not shown). This behavior lends  
263 support to the hypothesis of Senior and Mitchell (2000) who propose that feedbacks change  
264 through time due to the slow warming rates of the Southern Ocean relative to the upper  
265 atmospheric levels. This reduced lapse rate increases atmospheric static stability (and thus,  
266 the shortwave cloud response) in the transient part of the simulation, but decreasingly less  
267 so towards equilibrium.

268 The extra-tropical cloud response in the model-mean is non-negligible in the Southern  
269 Ocean and North Atlantic on decadal to centennial timescales, as proposed by Rose and  
270 Rencurrel (2016) and Trossman et al. (2016). However, it comes to dominate the global  
271 response only on centennial to millennial timescales and when both hemispheres are consid-  
272 ered.

273 We find that the longwave clear sky feedback does moderately increase in many mod-  
274 els as the temperature or the forcing level increases, mainly in the tropics and Northern  
275 Hemisphere mid-latitudes (Fig. 2a, SM Fig. 4, SM Fig. 5). This is in accordance with the  
276 proposed argument that the tropics govern the global feedback evolution because the water  
277 vapor feedback increases with warming (Jonko et al., 2013; Meraner et al., 2013; Block &  
278 Mauritsen, 2013; Andrews et al., 2015), possibly following the rising tropical tropopause  
279 (Meraner et al., 2013; Mauritsen et al., 2018).

280 Recent work has focused on the relative influence of the Pacific, specifically the relative  
281 influence of temperatures of the warm pool versus compared to other regions. Feedbacks in  
282 regions of atmospheric deep convections have a far-field and global effect, while feedbacks  
283 in regions of atmospheric subsidence have only a local or regional influence (Barsugli &  
284 Sardeshmukh, 2002; Zhou et al., 2017; Andrews & Webb, 2018; Ceppi & Gregory, 2019;  
285 Dong et al., 2019). With the available fields in the LongRunMIP archive, we cannot quan-  
286 tify the relative importance of water vapor and lapse rate feedbacks. However, the short and  
287 longwave cloud response (SM Fig. 6–8) in the models qualitatively agree with the proposed  
288 change of tropospheric stability patterns on decadal to centennial timescales (Andrews &  
289 Webb, 2018; Ceppi & Gregory, 2017), especially in the Pacific region. In contrast, on centen-

290 nial to millennial timescales, the tropical Pacific response becomes less important compared  
291 to the mid-latitudes and the net tropical CRE does not change anymore (SM Fig. 6).

## 292 5 Implications

293 We demonstrate that the evolution of the global feedback response is dominated by the  
294 mid-latitudes on centennial to millennial and the tropics on decadal to centennial timescales.  
295 The global net feedback change is a result of a subtle balance of different regions and different  
296 TOA components at all times; even more so in single simulations than in the model mean  
297 shown here. This motivates process-based feedback studies in individual models as well  
298 as multi-model ensembles to draw robust conclusions and increase physical understanding  
299 of processes. To relate the timescales and model behavior to the observational record and  
300 paleo proxies a better understanding of a) the atmospheric versus oceanic drivers of surface  
301 temperature patters in both, the coupled climate models and the real world and b) the local  
302 and far field interactions of tropospheric stability, clouds, and surface temperatures need  
303 to be achieved. Note that climate models have typical and persistent biases in regions we  
304 identify as important, mainly the Equatorial Pacific, Southern Ocean and ocean upwelling  
305 regions. The pattern effect of the real world might act on timescales which are different  
306 than the ones of the climate models.

307 Our results show that radiative feedbacks, usually called “fast”, act continuously less  
308 stabilizing on the climate system as the models approach equilibrium. As a result, the  
309 equilibrium warming is higher than estimated with common extrapolation methods from  
310 short simulations for all models and simulations in the LongRunMIP archive. ECS has  
311 been historically used as a model characterization (Charney et al., 1979), but some studies  
312 propose that it is not the most adequate measure for estimating changes expected over the  
313 next decades and until the end of the century (e.g., Otto et al. (2013); Shiogama et al. (2016);  
314 Knutti et al. (2017)). Alternative climate sensitivity measures are the effective climate  
315 sensitivity computed on different timescales, the transient climate response to gradually  
316 increasing CO<sub>2</sub> (TCR), or the transient climate response to cumulative carbon emissions  
317 (e.g., Allen and Frame (2007); Millar et al. (2015); Gregory et al. (2015); Grose et al. (2018)).  
318 Beyond not being an accurate indicator of the equilibrium response, these alternative climate  
319 sensitivity measures capture the models in different degrees of equilibration. We show that  
320 it is an open question how different measures of sensitivity relate to each other. A recent  
321 study shows that  $\Delta T_{\text{est } 1-150}$  correlates better than TCR with end-of-21st-century warming

322 across model (Grose et al. (2018), see also Gregory et al. (2015)). Thus, we underscore  
323 the need of comparing models, observations, and paleo proxies on well-defined measures of  
324 climate sensitivity, which ensure they are in the same state of equilibration.

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326 Fields shown in this paper can be accessed on [https://data.iac.ethz.ch/longrunmip/  
327 GRL/](https://data.iac.ethz.ch/longrunmip/GRL/). See [www.longrunmip.org](http://www.longrunmip.org) and M. Rugenstein et al. (2019) for more details on each  
328 simulation and available variables, not shown here.

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