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Accepted Version

Rugenstein, M., Bloch-Johnson, J. ORCID:
<https://orcid.org/0000-0002-8465-5383>, Gregory, J., Andrews,
T., Mauritsen, T., Li, C., Frölicher, T. L., Paynter, D.,
Danabasoglu, G., Yang, S., Dufresne, J.-L., Cao, L., Schmidt,
G. A., Abe-Ouchi, A., Geoffroy, O. and Knutti, R. (2020)
Equilibrium climate sensitivity estimated by equilibrating
climate models. *Geophysical Research Letters*, 47 (4).
e2019GL083898. ISSN 0094-8276 doi:
<https://doi.org/10.1029/2019GL083898> Available at
<https://centaur.reading.ac.uk/87462/>

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To link to this article DOI: <http://dx.doi.org/10.1029/2019GL083898>

Publisher: American Geophysical Union

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Equilibrium climate sensitivity estimated by equilibrating climate models

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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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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 CO₂ 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 CO₂ 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

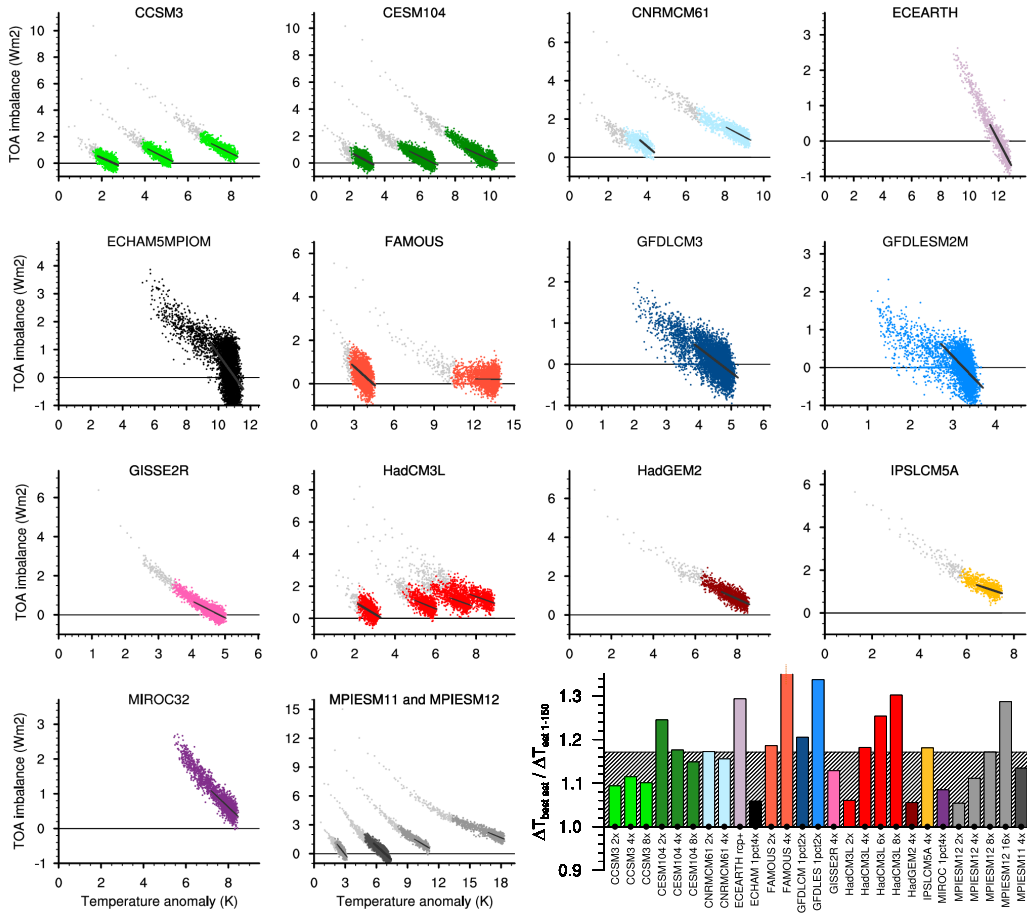


Figure 1. Evolution of global and annual mean top of the atmosphere (TOA) imbalance and surface temperature anomalies (14 small panels). The first 150 years of step forcing simulations are depicted in light gray. For experiments which are not step forcing simulations only the period after stabilizing 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 CO₂ 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% CO₂ 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 CO₂ 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 CO₂ 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 “ $\Delta T_{\text{est } 1-150}$ ”.

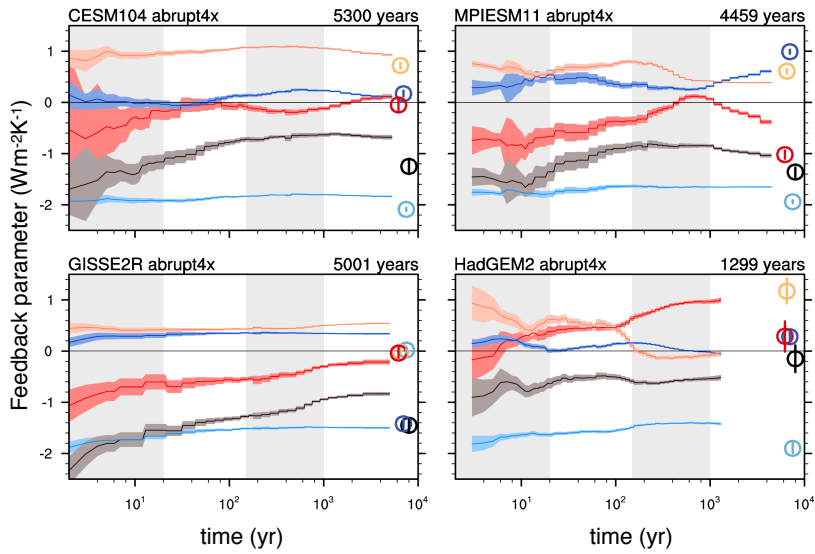
95 For simulations that have gradual forcings (e.g., *1pct2x*), we use 150 year long step forcing
 96 simulations of the same model to calculate $\Delta T_{\text{est } 1-150}$. The median increase of $\Delta T_{\text{best est}}$
 97 over $\Delta T_{\text{est } 1-150}$ is 17% for all simulations and 16% for the subset of CO₂ 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 CO₂ 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 CO₂ 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 CO₂ levels to estimate
 109 equilibrium warming for CO₂ doubling. We do so by assuming that the temperature scales
 110 with the forcing level, which depends logarithmically on the CO₂ 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 CO₂ 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

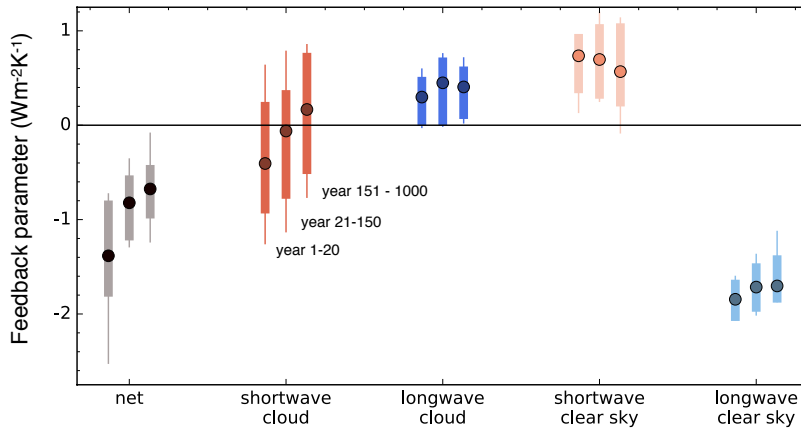


Figure 2. a) Time evolution of global feedbacks in four characteristic models. Net TOA feedback (gray) is the sum of its components: the cloud effects in the shortwave (red) and longwave (blue), and clear sky feedbacks in the shortwave (salmon) and longwave (light blue). Circles at the right of each 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”;

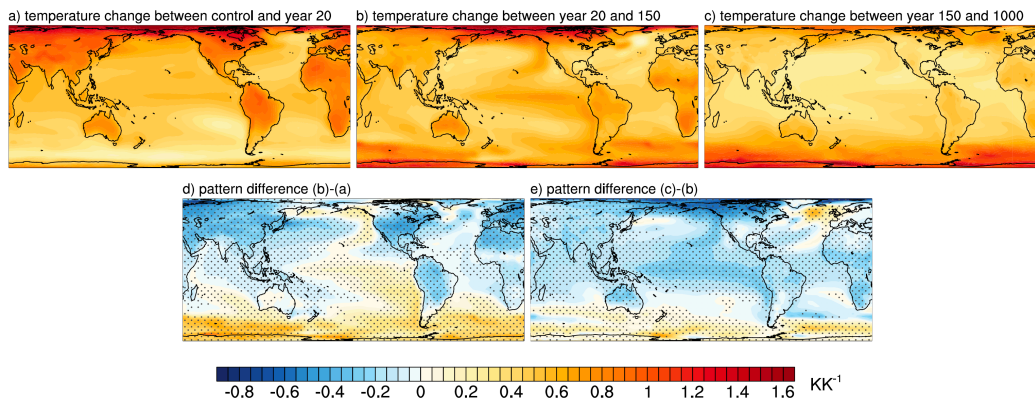


Figure 3. Multi-model mean normalized patterns of surface warming (local warming divided by global warming) between the average of (a) the control simulation and year 15-25, (b) year 15-25 and 140-160, (c) year 140-160 and 800-1000, and their differences (d and e) for the same models and simulations 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 CO₂ 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,

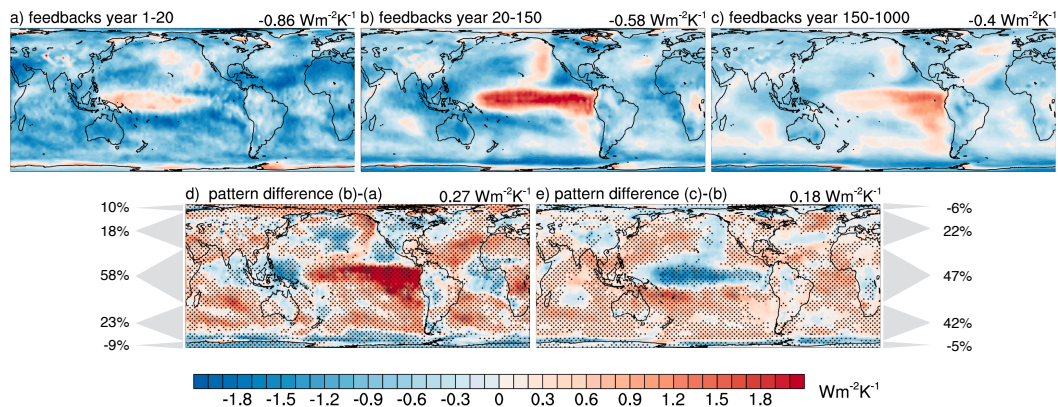


Figure 4. Time evolution of feedback patterns. Model-mean of local contribution to the change in global feedbacks (local TOA anomaly divided by global warming during the period indicated in the panel titles; see text for definitions) (a–c) and their differences (d, e). The global feedback value is shown in the panel title. Regionally aggregated contributions to the global values are indicated with percent numbers and gray triangles (22°S - 22°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 patterns 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 CO₂ (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.

325 **Acknowledgments**

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 and M. Rugenstein et al. (2019) for more details on each
328 simulation and available variables, not shown here.

329 We thank Urs Beyerle, Erich Fischer, Jeremy Rugenstein, Levi Silvers, and Martin Stolpe
330 for technical help and comments on the manuscript.

331 MR is funded by the Alexander von Humboldt Foundation. NCAR is a major facility spon-
332 sored by the U.S. National Science Foundation under Cooperative Agreement 1852977. TA
333 was supported by the Joint UK BEIS/Defra Met Office Hadley Centre Climate Programme
334 (GA01101). TLF acknowledges support from the Swiss National Science Foundation un-
335 der Grant PP00P2_170687, from the European Union's Horizon 2020 research and innova-
336 tion program under grant agreement no. 821003 (CCiCC) CL was supported through the
337 Clusters of Excellence CliSAP (EXC177) and CLICCS (EXC2037), University Hamburg,
338 funded through the German Research Foundation (DFG). SY was partly supported by Eu-
339 ropean Research Council under the European Community's Seventh Framework Programme
340 (FP7/20072013)/ERC grant agreement 610055 as part of the ice2ice project.

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