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1 **On the effect of historical SST patterns on radiative feedback**

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48 **Abstract**

49 We investigate the dependence of radiative feedback on the pattern of sea-surface temperature
50 (SST) change in fourteen Atmospheric General Circulation Models (AGCMs) forced with observed
51 variations in SST and sea-ice over the historical record from 1871 to near-present. We find that over
52 1871-1980, the Earth warmed with feedbacks largely consistent and strongly correlated with long-
53 term climate sensitivity feedbacks (diagnosed from corresponding atmosphere-ocean GCM abrupt-
54 4xCO₂ simulations). Post 1980 however, the Earth warmed with unusual trends in tropical Pacific
55 SSTs (enhanced warming in the west, cooling in the east) that drove climate feedback to be
56 uncorrelated with – and indicating much lower climate sensitivity than – that expected for long-term
57 CO₂ increase. We show that these conclusions are not strongly dependent on the AMIP II SST dataset
58 used to force the AGCMs, though the magnitude of feedback post 1980 is generally smaller in eight
59 AGCMs forced with alternative HadISST1 SST boundary conditions. We quantify a ‘pattern effect’
60 (defined as the difference between historical and long-term CO₂ feedback) equal to 0.44 ± 0.47 [5-
61 95%] $\text{W m}^{-2} \text{K}^{-1}$ for the time-period 1871-2010, which increases by $0.05 \pm 0.04 \text{ W m}^{-2} \text{K}^{-1}$ if calculated
62 over 1871-2014. Assessed changes in the Earth’s historical energy budget are in agreement with the
63 AGCM feedback estimates. Furthermore satellite observations of changes in top-of-atmosphere
64 radiative fluxes since 1985 suggest that the pattern effect was particularly strong over recent
65 decades, though this may be waning post 2014 due to a warming of the eastern Pacific.

66 1. Introduction

67 1.1. Background

68 A common starting point for quantifying the sensitivity of the Earth's climate to external
69 perturbations is consideration of the global-mean energy budget, $N = F + \lambda T$, where N is the net
70 downward radiative flux at the top-of-atmosphere (TOA) (units W m^{-2}), F the effective radiative
71 forcing (units W m^{-2}), λ the climate feedback parameter (units $\text{W m}^{-2} \text{K}^{-1}$, a negative number in this
72 paper, but the opposite convention is also used) and T the surface-air-temperature change (units K)
73 relative to an unperturbed steady state in which $N=F=0$. Applied to non-steady states, such as the
74 Earth's historical record (since the 1800s), λ is determined via either (i) differences (denoted by Δ)
75 between two climate states (often present-day and pre-industrial) according to $\lambda = (\Delta N - \Delta F)/\Delta T$ (e.g.
76 Gregory et al., 2002; Otto et al., 2013; Sherwood et al., 2020), or (ii) regression in the differential
77 form $\lambda = d(N - F)/dT$ if the timeseries of N , F and T are known (Gregory et al. 2004; Gregory et al.
78 2020).

79 Until recently it was often assumed that λ was - to a good approximation - a constant property of the
80 climate system, such that feedbacks that applied over the historical record also applied to the
81 Earth's long-term response, as quantified by the canonical equilibrium climate sensitivity (ECS, units
82 K) to a forcing from a doubling of CO_2 (F_{2x}) over pre-industrial levels. Thus ECS was estimated directly
83 from historical changes in N , T and F , according to $\text{ECS} = -F_{2x}/\lambda = -F_{2x} \Delta T / (\Delta N - \Delta F)$ (e.g. Gregory et
84 al., 2002; Otto et al., 2013, amongst many others).

85 However, it is now recognised that λ varies in time since a forcing is applied and with the strength
86 and/or type of that forcing (e.g. Senior and Mitchell, 2000; Hansen et al., 2005; Andrews et al. 2012;
87 Armour et al., 2013; Geoffroy et al., 2013; Rose et al. 2014; Gregory et al. 2015; Andrews et al. 2015;
88 Marvel et al. 2016; Rugenstein et al. 2016; Richardson et al., 2019; Dong et al. 2020; Bloch-Johnson
89 et al., 2021; Rugenstein and Armour, 2021). Hence λ is an 'effective feedback parameter' that applies
90 only to the climate change over which it was calculated. More specifically, over the historical record
91 λ is thought to be more stabilizing (more negative, climate sensitivity smaller) than might operate in
92 the long-term future, and so λ estimated from historical climate change would understate ECS (e.g.
93 Gregory and Andrews, 2016; Zhou et al., 2016; Armour, 2017; Proistosescu & Huybers, 2017;
94 Andrews et al., 2018; Marvel et al., 2018; Silvers et al., 2018; Lewis and Curry, 2018; Gregory et al.
95 2020; Sherwood et al. 2020; Dong et al. 2021).

96 The reason for the underestimate of long-term ECS is that climate feedbacks setting λ , such as cloud
97 and lapse-rate changes, vary with the pattern of surface warming. Proxy reconstructions of past
98 equilibrium climates and atmosphere-ocean general circulation model (AOGCM) simulations of long-
99 term climate change show an 'ENSO-like' temperature pattern with strong temperature changes in
100 the eastern Pacific as well as the Southern Ocean, whereas observed historical warming shows more
101 pronounced warming in the western equatorial Pacific relative to the tropical mean and cooling in
102 the eastern Pacific and Southern Ocean over recent decades (e.g. Collins et al., 2013; Li et al., 2013;
103 Andrews et al., 2015; Gregory and Andrews, 2016; Zhou et al., 2016; Dong et al., 2019; Sherwood et
104 al., 2020; Rugenstein et al. 2020; Olonscheck et al., 2020; Fueglistaler and Silvers, 2021; Watanabe et
105 al. 2021; Power et al. 2021; Tierney et al. 2019; 2020).

106 Thus, more-stabilizing feedbacks have occurred over the historical record because enhanced
107 warming in the western Pacific warm pool – a region of deep ascent and convection – results in a
108 stronger negative lapse-rate feedback widely across the tropics due to efficient warming of the free
109 troposphere, which in turn causes increased cloudiness (a negative cloud feedback) in the eastern

110 tropical Pacific due to remotely controlled increased lower tropospheric stability. In contrast, less-
111 stabilizing feedbacks are expected in the future as enhanced warming in the eastern Pacific –
112 characterised by descending air and marine low cloud decks which are capped under a temperature
113 inversion and form over the relatively cool sea-surface-temperatures (SSTs) – results in a positive
114 cloud feedback, without an accompanying negative lapse-rate feedback since the warming is
115 ‘trapped’ in the boundary layer (e.g. Zhou et al., 2016, Andrews and Webb, 2018, Ceppi and Gregory,
116 2017; Dong et al. 2019).

117 The dependence of radiative feedback on the pattern of surface temperature change has been
118 termed a ‘pattern effect’ (Stevens et al., 2016), which distinguishes it from other feedback variations
119 that might occur for example as a function of the magnitude of ΔT (e.g. Block & Mauritsen, 2013;
120 Caballero and Huber, 2013; Bloch-Johnson et al., 2021). Armour (2017) and Andrews et al. (2018)
121 proposed a method to account for the pattern effect in estimates of ECS derived from historical
122 climate changes via a modification of the energy budget approach. Their method requires an
123 estimate of the difference in feedback, $\Delta\lambda$, due to the pattern effect between historical climate
124 change and long-term ECS, so that $ECS = -F_{2x}/(\lambda_{hist} + \Delta\lambda)$, where λ_{hist} is the historical value. Since $\Delta\lambda$ is
125 found to be positive, it increases the best estimate of ECS and substantially lifts the upper
126 uncertainty bound, but has only a small impact on the lower bound (Armour, 2017; Andrews et al.,
127 2018; Sherwood et al. 2020).

128 One way of estimating the pattern effect, $\Delta\lambda$, is to contrast λ_{hist} in an Atmospheric GCM (AGCM)
129 simulation forced by observed historical SST and sea-ice variations (termed an *amip-piForcing*
130 simulation, see Section 2) with λ_{4xCO2} in the coupled AOGCM *abrupt-4xCO2* simulation with the same
131 AGCM, so that $\Delta\lambda = \lambda_{4xCO2} - \lambda_{hist}$ (Andrews et al. 2018). Note that here *abrupt-4xCO2* is being used as
132 a surrogate² for long-term ECS. We assume other impacts on λ , such as the nature of the forcing
133 agent – so called ‘efficacies’ (Hansen et al., 2005; Marvel et al. 2016; Richardson et al., 2019) –
134 primarily occur due to forcing-specific impacts on historical SST patterns that will be included in the
135 historical record, rather than any dependence on the actual forcing agent concentration in the
136 atmosphere (which will be excluded in our design, because forcing levels are fixed at pre-industrial
137 levels in *amip-piForcing*). On the other hand, *abrupt-4xCO2* experiments contain larger warming
138 than the historical record, so any state dependence on T (e.g. Block & Mauritsen, 2013; Caballero
139 and Huber, 2013; Bloch-Johnson et al., 2021) might erroneously be included in pattern effect
140 estimates using this method.

141 The principal advantage of using *amip-piForcing* simulations in the calculation of the pattern effect is
142 that λ_{hist} will be consistent with the SST patterns that occurred over the historical record. In contrast,
143 one could use AOGCM historical simulations for λ_{hist} , but when AOGCMs are free to simulate their
144 own historical SST patterns they struggle to reproduce the observed recent decadal trends in
145 tropical Pacific SST patterns (Gregory et al. 2020; Fueglistaler and Silvers, 2021; Watanabe et al.
146 2021; Dong et al., 2021) and the associated magnitude of λ_{hist} , thus underestimating the pattern
147 effect (Gregory et al., 2020; Dong et al. 2021). This AOGCM bias in the pattern effect has important
148 implications, which we return to in the Discussion, but our focus in this manuscript is on the

² We use λ_{4xCO2} rather than an equilibrium feedback, λ_{eqm} , because in practice equilibrium is difficult to achieve in AOGCMs due to the millennial timescales required to equilibrate the deep ocean. The feedback parameter associated with ECS is therefore often approximated from short (~ 150 years) *abrupt-4xCO2* experiments (Andrews et al. 2012). Technically this is still an ‘effective feedback parameter’ and associated ‘effective climate sensitivity’ (EffCS), but in practice it is found to provide a suitable analogue for long-term feedbacks in climate projections (Grose et al., 2018) and ECS (Sherwood et al. 2020) and so this distinction is not considered further (see Rugenstein et al. (2020) and Rugenstein and Armour (2021) for further discussion).

149 historical pattern effect as simulated by AGCMs *given* the observed SSTs, thus avoiding the issue of
150 AOGCM biases in historical SST patterns. Note that while our focus is on the atmospheric response
151 to a given SST pattern, causality can work in both directions. For example cloud feedback has been
152 shown to have an impact on the pattern of tropical Pacific SST changes (Chalmers et al., 2022).

153 *amip-piForcing* simulations also show multi-decadal variations in λ_{hist} (Gregory and Andrews 2016;
154 Zhou et al., 2016; Andrews et al., 2018; Fueglistaler and Silvers, 2021; Dong et al. 2021). In particular
155 λ_{hist} is generally most negative (pattern effect largest) over the most recent decades. This is because
156 variations in atmospheric feedback are well explained by changes in SSTs in regions of tropical deep
157 convection relative to the tropical-mean (Fueglistaler and Silvers, 2021) or global-mean (Dong et al.
158 2019). Since the late 1970s, regions of deep convection have warmed by about +50% more than the
159 tropical-mean (Fueglistaler and Silvers, 2021), and the eastern Pacific has cooled despite
160 temperatures increasing globally (e.g. Hartmann et al. 2013; Power et al. 2021; and see our Figures 4
161 and 9). Hence under this configuration of tropical Pacific SST change, we would expect negative
162 feedback from the mechanisms described above (e.g. Zhou et al., 2016, Andrews and Webb, 2018,
163 Ceppi and Gregory, 2017; Dong et al. 2019).

164 A limitation of the *amip-piForcing* experiment for quantifying λ_{hist} is that it may include a structural
165 dependence on the underlying SST patterns and sea-ice in the Atmospheric Model Intercomparison
166 Project (AMIP) II boundary condition data set (Gates et al., 1999; Hurrell et al., 2008; Taylor et al.,
167 2000) used to force the *amip-piForcing* simulations (Andrews et al., 2018; Lewis and Mauritsen,
168 2021; Zhou et al., 2021; Fueglistaler and Silvers, 2021). Different SST reconstructions have slightly
169 different patterns of SST change over the historical period, and λ_{hist} may be affected. Indeed Lewis
170 and Mauritsen (2021) and Fueglistaler and Silvers (2021) showed that warming in the tropical
171 western Pacific relative to the tropical-mean is less pronounced in other SST datasets, and so we
172 might expect less negative feedbacks ($\Delta\lambda$ less positive) if the AGCMs were forced with non-AMIP II
173 datasets.

174 Consistent with this expectation, Andrews et al. (2018) noted that in one AGCM the magnitude of
175 λ_{hist} was reduced by $\sim 0.2 \text{ W m}^{-2} \text{ K}^{-1}$ when the AMIP II SSTs were replaced by HadISST2.1 SSTs (sea-ice
176 remaining unchanged) in an *amip-piForcing* simulation. Partly because of this, Sherwood et al.
177 (2020) and Forster et al. (2021) assessed the historical pattern effect to be smaller and more
178 uncertain ($\Delta\lambda = 0.5 \pm 0.5 \text{ W m}^{-2}$) than simply taking the *amip-piForcing* based model distribution
179 reported by Andrews et al. (2018) ($\Delta\lambda = 0.64 \pm 0.40 \text{ W m}^{-2}$). Subsequently, Lewis and Mauritsen
180 (2021) and Zhou et al. (2021) also found λ_{hist} to be less negative ($\Delta\lambda$ smaller) when using other SST
181 datasets than AMIP II used in *amip-piForcing* simulations discussed here.

182 1.2. Aims and motivating questions

183 Andrews et al. (2018) provides much of the published quantitative analysis on λ_{hist} to observed SST
184 patterns and $\Delta\lambda$, but only six AGCMs from only four different modelling centres were considered.
185 Hence, a first motivation of this manuscript is to revisit their numbers with a broader set of models
186 by utilizing the new *amip-piForcing* simulations from the Cloud Feedback Model Intercomparison
187 Project phase 3 (CFMIP, Webb et al. 2017) contribution to the Coupled Model Intercomparison
188 Project phase 6 (CMIP6, Eyring et al., 2016). The larger ensemble totalling 14 models when
189 combined will provide a more robust quantification of the magnitude and spread of λ_{hist} and $\Delta\lambda$ to a
190 broader set of model physics and climate sensitivities (Zelinka et al. 2020; Meehl et al. 2020; Flynn
191 and Mauritsen, 2020).

192 Secondly, the limited set of models in Andrews et al. (2018) prevented them from robustly exploring
193 and quantifying the relationship between λ_{hist} and $\lambda_{4\times\text{CO}_2}$ across models. In other words, it is not
194 known whether feedbacks acting over the historical record in AGCMs are correlated to feedbacks
195 acting on long-term ECS. For example is there a relationship between the two that could form the
196 basis of an emergent constraint? Do different parts of the historical record relate better to
197 feedbacks acting on long-term ECS than other parts, and why? As we will show, feedbacks over
198 different parts of the historical record have different relationships to $\lambda_{4\times\text{CO}_2}$, and this is important for
199 understanding what can and cannot be directly constrained from the historical record.

200 Thirdly, λ_{hist} and $\Delta\lambda$ have been shown to vary substantially on decadal timescales with λ_{hist} being most
201 negative (pattern effect largest) over recent decades since ~ 1980 (Gregory and Andrews 2016; Zhou
202 et al., 2016; Andrews et al., 2018; Gregory et al. 2020; Dong et al. 2021). This is consistent with the
203 findings of Fueglistaler and Silvers (2021), who identified ~ 1980 as the point in which the Earth
204 begins to warm with a particular (even “*peculiar*”) configuration of tropical Pacific SSTs where
205 “*regions of deep convection warm about +50% more than the tropical average*” driving large
206 negative cloud feedbacks. Hence we are motivated to separate λ_{hist} and $\Delta\lambda$ into a ‘before’ and ‘after’
207 1980. This separation leads into our next motivating question.

208 Fourthly, are observations of recent decadal change since the 1980s consistent with the AGCMs? If
209 so, what does a strong pattern effect in the presence of a substantial rate of global warming ($\sim 0.19 \text{ K}$
210 dec^{-1} , Tokarska et al., 2020) imply for the efficiency of ocean heat uptake and is there any
211 relationship between them? Loeb et al. (2020; 2021) identified a marked change in the Earth’s
212 radiation budget post 2014 associated with the 2015/2016 El Niño event and a change in sign in the
213 Pacific Decadal Oscillation (PDO) index. Such a shift in tropical Pacific SST patterns (a shift to
214 warming in the eastern Pacific) should favour more positive feedbacks. We ask whether evidence of
215 this is now potentially emerging in the satellite record.

216 Finally and fifthly, a limitation of the *amip-piForcing* approach, as discussed in Section 1.1, is that λ_{hist}
217 and $\Delta\lambda$ derived from these experiments includes a structural dependence on the SST patterns and
218 sea-ice in the AMIP II boundary condition data set used to force the AGCMs (Andrews et al., 2018;
219 Lewis and Mauritsen, 2021; Zhou et al., 2021; Fueglistaler and Silvers, 2021). To investigate this
220 further, we supplement the new *amip-piForcing* simulations with sensitivity tests with eight AGCMs
221 forced with historical HadISST1 (Rayner et al., 2003) SSTs as per Lewis and Mauritsen (2021).

222 In summary, previous studies have shown that historical climate feedback (λ_{hist}) varies on decadal
223 timescales in *amip-piForcing* simulations and is larger in magnitude (climate sensitivity smaller) than
224 that seen in long-term *abrupt-4xCO2* simulations associated with ECS, giving rise to a ‘pattern
225 effect’. This is accentuated over recent decadal climate change. Here we make use of observations
226 of the Earth’s energy budget from 1985 and a new suite of *amip-piForcing* simulations from
227 CFMIP3/CMIP6 (giving us a combined ensemble of 14 models), as well as targeted HadISST1 versus
228 AMIP II SST dataset sensitivity tests with eight AGCMs, to address the above question.

229 The manuscript is organised as follows: Section 2 describes the model and observational data.
230 Section 3 presents the model results. Section 4 brings in the observational data. Section 5 presents a
231 summary, discussion and outlook.

232

233

234

235 2. Methods and Data

236 2.1 *amip-piForcing*

237 To provide estimates of λ_{hist} consistent with the observed variations in SST patterns we turn to
238 AGCMs forced with observed monthly variations in SSTs and sea-ice, while keeping all forcing agents
239 such greenhouse gases and aerosols etc. constant at pre-industrial levels. Since the radiative forcing
240 is constant ($\Delta F=dF=0$) by construction, λ_{hist} can be diagnosed via $\lambda_{\text{hist}} = dN/dT$ (or $\Delta N/\Delta T$ if using finite
241 differences between climate states) (Andrews, 2014; Gregory and Andrews, 2016, Zhou et al., 2016;
242 Silvers et al., 2018; Andrews et al., 2018). Such an experimental design is now referred to as *amip-*
243 *piForcing* (Gregory and Andrews, 2016). The experimental protocol builds on the Atmospheric Model
244 Intercomparison Project (AMIP) design (Gates et al. 1999) that has long been used in climate
245 modelling, but extends back to 1870 (rather than 1979 in AMIP) and forcing agents are kept at pre-
246 industrial levels. As per AMIP, the underlying SST and sea-ice dataset used to force the AGCMs is the
247 AMIP II boundary condition data set (Gates et al., 1999; Hurrell et al., 2008; Taylor et al., 2000). A
248 description of the *amip-piForcing* protocol for CFMIP3/CMIP6 is given in Webb et al. (2017). When
249 forced with observed monthly SSTs and sea-ice, AGCMs generally reproduce the observed
250 relationships between surface temperature patterns, cloudiness and radiative fluxes well (Allan et
251 al., 2014; Loeb et al. 2020), lending some credibility to the radiative effects of their simulated
252 pattern effects to different SST patterns.

253 The *amip-piForcing* simulations used in this study are summarised in Table 1. They reflect a
254 combination of new CFMIP3/CMIP6 simulations with the latest generation of models archived in the
255 CMIP6 database and those used in Andrews et al. (2018) with some updates (see below). The
256 exception is MPI-ESM1-2-LR (Mauritsen et al., 2019); this is a CMIP6 generation model but its *amip-*
257 *piForcing* simulation is not currently included in the CMIP6 database. Note that this model contains
258 the ECHAM6.3 atmospheric model, so the results ought to be very similar to the older ECHAM6.3
259 simulations used in Andrews et al. (2018) and Lewis and Mauritsen (2021), though the models are
260 not identical owing to differences in atmospheric composition and land surface properties (see
261 Mauritsen et al., 2019, regarding the transition from MPI-ESM1.1 to MPI-ESM1.2). Furthermore, the
262 newer MPI-ESM1-2-LR simulations include a longer time-period than the ECHAM6.3 simulations
263 (Table 1).

264 The CFMIP3/CMIP6 *amip-piForcing* simulations begin in year 1870, but we discard the first year to
265 be consistent with the earlier Andrews et al. (2018) ensemble which started in January 1871. The
266 CFMIP3/CMIP6 simulations end in Dec 2014, whereas the simulations in the original Andrews et al.
267 (2018) ensemble (largely) ended in Dec 2010. In part to address this, some of the Andrews et al.
268 (2018) simulations have been rerun, including CAM4, GFDL-AM3 and GFDL-AM4 simulations, which
269 now end in Dec 2014 or later (see Table 1). Another difference to Andrews et al. (2018) is that we
270 now have an *abrupt-4xCO2* AOGCM simulation with GFDL-AM4 which they did not consider, to
271 permit a quantification of the pattern effect in that model. In contrast, we exclude the Andrews et
272 al. (2018) CAM5.3 simulation from our analysis since there is no *abrupt-4xCO2* AOGCM simulation to
273 compare against.

274 The models used, time-periods covered and number of ensembles are detailed in Table 1. Where
275 ensembles exist, an ensemble-mean dT and dN is created before analysis. Note that it makes little
276 difference to λ if, alternatively, individual members are first analysed and then the results ensemble-
277 meaned (Gregory and Andrews, 2016; Lewis and Mauritsen, 2021). All models share a common
278 1871-2010 time-period and so the principal analysis is restricted to this time-period, but we consider
279 the additional years to 2014 too. All data are global-annual-ensemble-means and expressed as

280 anomalies relative to an 1871-1900 baseline; the timeseries data has been made available (see Data
281 Availability Section).

282 Unless otherwise stated all uncertainties in multi model ensemble-mean results represent a 5-95%
283 confidence interval, calculated as 1.645σ across models assuming a gaussian distribution. We do not
284 attempt to adjust our uncertainty for the number of independent models, n , used in the ensemble
285 (i.e. dividing by square root of n). Our approach is similar to a "statistical indistinguishable ensemble"
286 approach (Annan and Hargraves, 2011; 2017) though likely overstates the uncertainty in the true
287 value if the ensemble shares characteristics of a "truth centred paradigm" (Sanderson and Knutti,
288 2012).

289 2.2 HadSST-piForcing

290 To test the sensitivity of the *amip-piForcing* results to the underlying SST dataset, we repeat the
291 *amip-piForcing* simulations with eight AGCMs (see Table 1) but replace the AMIP II boundary
292 condition SST dataset with HadISST1 (Rayner et al. 2003). All other aspects of the simulations,
293 including sea-ice, are identical to the *amip-piForcing* simulations. This is the same experimental
294 design as Lewis and Mauritsen (2021), and we include their ECHAM6.3 simulations here (which again
295 ought to be similar to the MPI-ESM1-2-LR simulations). The simulations cover a common time-period
296 across models of 1871-2010, like in *amip-piForcing*, but some models are also extended further (see
297 Table 1). We refer to these simulations as *hadSST-piForcing*, but note only the SSTs are from the
298 HadISST1 dataset (hence 'hadSST' rather than 'hadISST'), the sea-ice remains as per *amip-piForcing*.
299 Like *amip-piForcing*, all data are global-annual-ensemble-means and expressed as anomalies relative
300 to an 1871-1900 baseline, and the timeseries data has been made available (see Data Availability
301 Section).

302 Lewis and Mauritsen (2021) provide a summary of the source observational inputs used to construct
303 the AMIP II and HadISST1 SST datasets and how they differ. In addition, we note that AMIP II uses
304 HadISST1 SSTs (Rayner et al. 2003) prior to November 1981 and version 2 of the National Oceanic
305 and Atmospheric Administration (NOAA) weekly optimum interpolation (OI.v2) SST analysis
306 (Reynolds et al. 2002) thereafter. The merging procedure rebases the HadISST1 SSTs to avoid
307 discontinuities in the merged dataset (Hurrell et al. 2008). Hence AMIP II and HadISST1 might be
308 expected to be more similar before 1981, and diverge afterwards.

309 2.3 abrupt-4xCO2

310 A corresponding *abrupt-4xCO2* simulation using each AGCM's coupled AOGCM is used to determine
311 the model's long-term sensitivity metrics (F_{4x} , λ_{4xCO2} and $ECS = -0.5 * F_{4x} / \lambda_{4xCO2}$) from regression of
312 global-annual-mean dN against dT over 150 years of the simulations (see Andrews et al., 2012). We
313 also use λ_{4xCO2} diagnosed from years 1-20 and years 21-150 of the *abrupt-4xCO2* simulation following
314 Andrews et al. (2015), which approximately separates the two principal timescales of the climate
315 response: the mixed-layer and deep-ocean (see Geoffroy et al. 2013 and Andrews et al. 2015).
316 *abrupt-4xCO2 data* is available on the CMIP5 database (Taylor et al., 2012) for CCSM4, GFDL-CM3
317 and HadGEM2-ES. All other abrupt-4xCO2 data is available on the CMIP6 database, except for
318 HadCM3 and MPI-ESM1.1. For ECHAM6.3/MPI-ESM1.1, *abrupt-4xCO2* global-annual mean dN and
319 dT timeseries data are provided by Andrews et al. (2018). HadAM3 data is taken from Andrews et al.
320 (2018) and Andrews et al. (2015); while a mean of seven realizations, this simulation is only 100
321 years long so the calculations are over years 1-100 for λ_{4xCO2} and years 1-20 or 21-100 for the
322 separation of timescales in this model.

323 Note when aligning each AGCM to its AOGCM, sometimes the AGCM and AOGCM model names
324 differ in the literature. We indicate where this is applicable in Table 1. This does not apply to the
325 newer CFMIP3/CMIP6 simulations which publish their AGCM and AOGCM simulations under
326 consistent names.

327 *2.4 Observations of recent decadal climate change*

328 To understand Earth's recent decadal climate change since ~1985 we turn to its observed global-
329 mean energy budget (i.e. dT , dN and dF). For dT we use the HadCRUT5 analysis dataset (Morice et al.
330 2021) (the current version is HadCRUT.5.0.1.0). This is an improvement on previous HadCRUT
331 products and extends coverage in data sparse regions (see Morice et al. 2021). For dF we use the
332 best estimate historical ERF timeseries produced by IPCC AR6 (Forster et al. 2021). For dN we use
333 various versions of the DEEP-C satellite based reconstruction of the Earth's radiation balance from
334 1985 to near-present. These are described in detail in Allan et al. (2014) and Liu et al. (2015; 2017;
335 2020), but as we will use various versions of this product we give a brief overview here. The DEEP-C
336 dataset is derived by merging satellite observations of top-of-atmosphere radiative flux timeseries
337 from ERBE WFOV (Earth Radiation Budget Experiment Satellite wide field of view) and ECMWF
338 reanalysis (ERA-Interim/ERA5) since 1985 with CERES (Clouds and the Earth's Radiant Energy
339 System) satellite observed fluxes since March 2000. Hence prior to March 2000 it is largely informed
340 by ERBE WFOV and ERA reanalysis, then aligns with CERES from March 2000. AMIP and high
341 resolution AGCM simulations and reanalyses are used in the merging process to bridge the gaps
342 between products and avoid discontinuities in the timeseries, including a gap in the satellite record
343 during 1993 and 1999 (Allan et al. 2014). It is important to note that substantial uncertainty in
344 decadal changes in dN associated with the merging process affects the record and this is
345 conservatively estimated to be as high as 0.5 Wm^{-2} for changes applying across the whole record (Liu
346 et al. 2020). However, uncertainty in the CERES period since March 2000 is much smaller based on
347 assessment of instrument drift (Loeb et al. 2021). Various versions of the DEEP-C dataset exist which
348 parallel updates to the underlying products and update the merging process. We use the latest
349 version (DEEP-C v5, Liu and Allan 2022) for our principal analysis, which is based on CERES EBAF v4.1
350 and ERBS WFOV v3, alongside ERA5 reanalysis and AMIP6 simulations (Liu and Allan, 2022). To
351 illustrate structural uncertainties in our analysis we also use previous versions (v2, v3 and v4) of the
352 DEEP-C datasets. The availability of datasets is provided in the Data Availability Section.

353

354 **3. Historical feedback and pattern effect in amip-piForcing and hadSST-piForcing simulations**

355 Figure 1a shows the multi-model ensemble mean dT timeseries in the *amip-piForcing* and *hadSST-*
356 *piForcing* simulations, alongside an observed estimate from HadCRUT5 analysis dataset. The AGCM
357 design reproduces the observed historical dT variability well (the correlation coefficient, r , between
358 observed and both simulated dT timeseries is 0.97). However the AGCMs do not reproduce the
359 observed trends precisely, notably omitting some observed warming in the most recent decades
360 (Figure 1a; see also Andrews, 2014; Gregory and Andrews, 2016; Andrews et al. 2018). This is
361 because in the AGCM design only the prescribed SSTs and sea-ice are evolving according to the
362 observed dataset; the forcing agents (e.g. greenhouse gases, aerosols etc.) are prescribed at their
363 preindustrial level and land temperatures are free to evolve. Hence the AGCMs are free to simulate
364 their own land surface temperature variability (which may be different from that in the observed
365 historical record) and trends, which are found to be smaller than that observed because the AGCMs
366 do not include a land surface temperature change that arises as a consequence of increases in
367 greenhouse gases and other forcing agents independent of SST changes (see Andrews, 2014;
368 Gregory and Andrews, 2016; Andrews et al., 2018).

369 As dT increases, dN reduces (Figure 1b), i.e. the climate loses more heat to space as a consequence
 370 of the climate response and feedbacks in the system. Figure 1c and 1d show the difference in the dT
 371 and dN timeseries between the *amip-piForcing* and *hadSST-piForcing* ensemble-mean response. For
 372 most of the time the differences vary approximately about zero. However, larger differences are
 373 evident from 1981 onwards, when the dN response in *amip-piForcing* is substantially larger than that
 374 in *hadSST-piForcing* (Figure 1b and 1d), up to $\sim 0.5 \text{ W m}^{-2}$ in some years (Figure 1d). This is consistent
 375 with 1981 being the year in which the AMIP II boundary condition source dataset switches from
 376 HadISST1 to OI.v2 SST (see Section 3.2). This motivates us to separate the historical record into two
 377 time-periods either side of 1980, i.e. 1871-1980 and 1981-2010 (Section 3.2).

378 However, we first consider feedback and the pattern effect that arises when calculated over the
 379 historical record as a whole, rather than any time-period within (Section 3.1). This is useful for
 380 informing studies that use the entire observed historical record to estimate ECS via energy budget
 381 constraints (e.g. Andrews et al., 2018; Sherwood et al. 2020; Forster et al. 2021). It also allows a
 382 direct comparison of our results using a broad ensemble of models to the narrower range of model
 383 results reported by Andrews et al. (2018) and Lewis and Mauritsen (2021).

384 3.1 Considering the historical record as a whole

385 Figures 1e and 1f show the $\lambda_{\text{hist}} = dN/dT$ relationship in the ensemble-mean *amip-piForcing* and
 386 *hadSST-piForcing* simulation for 1871-2010. λ_{hist} is determined from ordinary least square linear
 387 regression on global-annual-mean dN and dT timeseries data. λ_{hist} values for individual models are
 388 given in Table 2 alongside their *abrupt-4xCO2* sensitivity metrics. Across the fourteen model
 389 ensemble of *amip-piForcing* simulations $\lambda_{\text{hist}} = -1.65 \pm 0.46 \text{ W m}^{-2} \text{ K}^{-1}$, slightly smaller in magnitude
 390 but with similar spread to the Andrews et al. (2018) ensemble (they reported $\lambda_{\text{hist}} = -1.74 \pm 0.48 \text{ W m}^{-2} \text{ K}^{-1}$).
 391 Like in Andrews et al. (2018), the spread in λ_{hist} is extremely similar to the spread in λ_{4xCO2} from
 392 the coupled AOGCM *abrupt-4xCO2* ensemble (Table 2). The pattern effect, $\Delta\lambda = \lambda_{4xCO2} - \lambda_{\text{hist}}$ between
 393 *amip-piForcing* and *abrupt-4xCO2* (with λ_{4xCO2} from years 1-150 of *abrupt-4xCO2*) is $\Delta\lambda = 0.70 \pm 0.47$
 394 $\text{W m}^{-2} \text{ K}^{-1}$ across the ensemble (Table 3), which is slightly larger in magnitude but with more spread
 395 than that reported by Andrews et al. (2018) ($0.64 \pm 0.40 \text{ W m}^{-2} \text{ K}^{-1}$).

396 Tables 2 and 3 present the equivalent λ_{hist} and $\Delta\lambda$ values when the AGCMs are forced with HadISST1
 397 SSTs instead (*hadSST-piForcing*) and Figure 2 shows the relationship to *amip-piForcing*. $\lambda_{\text{hist}} = -1.43 \pm$
 398 $0.43 \text{ W m}^{-2} \text{ K}^{-1}$ in *hadSST-piForcing* (Table 2), which is smaller in magnitude but with similar spread to
 399 the *amip-piForcing* results above. Subsetting to the eight AGCMs with both simulations, λ_{hist} is $0.26 \pm$
 400 $0.16 \text{ W m}^{-2} \text{ K}^{-1}$ smaller in magnitude in *hadSST-piForcing* but well correlated ($r=0.94$) with *amip-*
 401 *piForcing* values (Figure 2a, red points). The regression slopes of the red line in Figures 2a (slope =
 402 0.9 ± 0.2) and 2b (slope = 1.1 ± 0.4) are statistically consistent with unity, implying there is little
 403 AGCM dependence in the difference between λ_{hist} from *amip-piForcing* and *hadSST-piForcing*.
 404 Hence, given the the strong correlation and close approximation of being parallel to the one-to-one
 405 line (Figure 2, red points), we suggest a simple offset given by the difference ($0.26 \pm 0.16 \text{ W m}^{-2} \text{ K}^{-1}$)
 406 well approximates the relationship between λ_{hist} over 1871-2010 in *amip-piForcing* and *hadSST-*
 407 *piForcing*.

408 Despite λ_{hist} being smaller in magnitude in *hadSST-piForcing*, $\Delta\lambda = 0.44 \pm 0.31 \text{ W m}^{-2} \text{ K}^{-1}$ is still large
 409 and positive across the *hadSST-piForcing* ensemble (Table 3). The smaller uncertainty than the *amip-*
 410 *piForcing* pattern effect likely reflects the narrower diversity of model physics in the smaller *hadSST-*
 411 *piForcing* ensemble, for example we do not have *hadSST-piForcing* experiments for the models with
 412 the largest (CESM2) or smallest (MIROC6) pattern effects in *amip-piForcing*. If we subset the *amip-*
 413 *piForcing* ensemble to just those eight models with corresponding *hadSST-piForcing* experiments (Fig

414 2b, red points), then the spread (as measured by 1.645σ in Table 3) across models in $\Delta\lambda$ reduces
415 from 0.47 to 0.28, which is similar to the spread found in *hadSST-piForcing*.

416 That a large pattern effect is present in the *hadSST-piForcing* simulation over the historical record is
417 not in contradiction with the results of Lewis and Mauritsen 2021 (LM2021), who reported a
418 ‘negligible unforced historical pattern effect’ with ECHAM6.3 when forced with HadISST1 SSTs. This is
419 because LM2021 calculated their pattern effect by comparing λ from *hadSST-piForcing* to λ derived
420 from a coupled AOGCM historical simulation, or approximations of it from years 1-70 of 1%CO₂ or
421 years 1-50 of *abrupt-4xCO₂* simulations. This necessarily gives a smaller pattern effect because it
422 excludes many of the SST variations and patterns effects seen on longer timescales in CO₂ forced
423 simulations (Senior and Mitchell, 2000; Gregory et al. 2004; Andrews et al. 2012; Armour et al.,
424 2013; Geoffroy et al., 2013; Andrews et al. 2015; Rugenstein et al. 2016). While this might be useful
425 for trying to quantify different mechanisms of the pattern effect (e.g. forced or unforced), it is a
426 quantity we are less interested in, as we want to know the λ of relevance to long-term ECS and
427 projections of the late 21st century. Therefore contrasting to λ_{4xCO_2} from years 1-150 is the most
428 relevant metric (Sherwood et al., 2020), as we have done here.

429 Following Andrews et al. (2018) we decompose λ into its component longwave (LW) clear-sky,
430 shortwave (SW) clear-sky and cloud radiative effect (CRE, equal to all-sky minus clear-sky fluxes)
431 terms in Figure 3. Deviations away from the one-to-one line indicate a difference in *amip-piForcing*
432 and *abrupt-4xCO₂* λ (i.e. the pattern effect). Tables of the individual model results are given in the
433 Supplementary Tables 1 - 3. It confirms the basic premise that historical LW clear-sky and cloud
434 feedbacks are more stabilizing than under *abrupt-4xCO₂*, consistent with the mechanistic and
435 process understanding that the pattern effect arises predominantly from a lapse-rate (which affects
436 LW clear-sky fluxes) and cloud feedback dependence on SST patterns (e.g. Zhou et al., 2016,
437 Andrews and Webb, 2018, Ceppi and Gregory, 2017; Dong et al. 2019). Figure 3 also suggests there
438 is a small compensation to the total pattern effect from SW clear-sky feedbacks, likely from sea-ice.
439 That is, AGCMs forced with AMIP II boundary condition sea-ice changes have a slightly more positive
440 feedback than found in their coupled *abrupt-4xCO₂* simulations, though the difference is small
441 (Figure 3). Consequently, a simple attribution of the difference in total feedback between *amip-*
442 *piForcing* and *abrupt-4xCO₂* to an SST driven pattern effect (as we have done here) will slightly
443 understate the actual effect, though the term is small and we neglect it from now on. We discuss
444 sea-ice uncertainties further below.

445 MIROC6 is the only model in the *amip-piForcing* ensemble to have near zero pattern effect (Table 3
446 and note the single black dot on the one-to-one line in Figure 3). The reason for this different
447 behaviour remains unclear. One could speculate that there is a relationship between a model’s
448 climate sensitivity and its pattern effect, given that MIROC6 has the lowest ECS of all models
449 consider here (ECS=2.6K, Table 2). However, we note that there is little correlation between ECS and
450 $\Delta\lambda$ across models ($r=0.4$) and that several other models with low ECS have large $\Delta\lambda$.

451 Alternatively, it could be that MIROC6’s atmospheric physics are largely insensitive to different SST
452 patterns and/or that its AOGCM *abrupt-4xCO₂* warming pattern is more similar to the historical
453 record than other models. Both are potentially possible. For example, λ_{hist} for 1871-1980 and 1980-
454 2010 separately (next Section and Table 2) shows that MIROC6 does simulate a pattern effect, but
455 achieves a near zero pattern effect over the historical record as a whole by having a smaller (relative
456 to other models) pattern effect over recent decades, offset by a negative pattern effect over the
457 earlier period. In addition - and in contrast to other models - MIROC6 simulates a negative LW clear-
458 sky pattern effect (red dot below the one-to-one line, Figure 3) which offsets its positive cloud
459 feedback pattern effect.

460 The model with the largest pattern effect is CESM2 (Table 3). This occurs because of a particularly
461 large cloud feedback sensitivity to SST patterns (grey dot furthest from the one-to-one line, Figure
462 3). Zhu et al. (2022) argue that an issue in CESM2's cloud microphysics related to cloud ice number
463 leads to an unrealistically large cloud sensitivity to warming in this model. Whether this is
464 responsible for the model's large pattern effect is unclear. Mixed-phase clouds have not typically
465 been associated with the pattern effect, though might clearly be of relevance to pattern effects over
466 the Southern Ocean (Dong et al. 2020; Bjordal et al. 2020). It would be interesting in future work to
467 identify the different cloud types associated with the pattern effect and sensitivity experiments with
468 CESM2 to investigate which aspects of the cloud feedback change with different cloud microphysics
469 schemes.

470 Many of our *amip-piForcing* (eleven models) and *hadSST-piForcing* (five models) simulations
471 continue to Dec 2014 (Table 1), so we consider how this extended period affects the overall
472 assessment of the historical pattern effect. In the eleven *amip-piForcing* simulations, $\lambda_{\text{hist}} = -1.65 \pm$
473 $0.48 \text{ W m}^{-2} \text{ K}^{-1}$ over 1871-2010, but this increases in magnitude to $\lambda_{\text{hist}} = -1.71 \pm 0.51 \text{ W m}^{-2} \text{ K}^{-1}$ if
474 calculated over 1871-2014 (Supplementary Table 4). An increase occurs in every model and the
475 magnitude of change across the ensemble is $0.07 \pm 0.06 \text{ W m}^{-2} \text{ K}^{-1}$ (Supplementary Table 4). In the
476 five *hadSST-piForcing* simulations, $\lambda_{\text{hist}} = -1.47 \pm 0.45 \text{ W m}^{-2} \text{ K}^{-1}$ over 1871-2010, but this increases in
477 magnitude to $\lambda_{\text{hist}} = -1.52 \pm 0.42 \text{ W m}^{-2} \text{ K}^{-1}$ if calculated over 1871-2014 (Supplementary Table 4). The
478 magnitude of the increase ($0.05 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$) is thus slightly smaller in this dataset
479 (Supplementary Table 4).

480 While we have focused on the SST driven pattern effect, a remaining structural uncertainty in
481 assessing total feedback differences between $\lambda_{4\times\text{CO}_2}$ and λ_{hist} relates to the sea-ice dataset used to
482 force the AGCMs. Andrews et al. (2018) provided a sensitivity test (see their Supplementary
483 Material) by repeating the *amip-piForcing* simulation in two AGCMs but forced with HadISST2.1
484 (Titchner and Rayner, 2014) SSTs and sea-ice. They found that the historical feedback parameter
485 increased by $\sim 0.6 \text{ W m}^{-2} \text{ K}^{-1}$ when forced with HadISST2.1 compared to AMIP II, and attributed most
486 of this change to differences in the sea-ice datasets rather than SST. They noted that HadISST2.1 has
487 substantially more pre-industrial Antarctic sea-ice concentration (see Titchner and Rayner, 2014),
488 and so generated more sea-ice loss (more positive feedback) over the historical period (Andrews et
489 al. 2018), as well containing large discontinuities in the timeseries. The historical sea-ice trends and
490 associated feedbacks over the Southern Ocean in the HadISST2.1 dataset are difficult to reconcile
491 with those found in AOGCMs and our physical understanding of them (Schneider et al. 2018). We do
492 not pursue this further, but simply highlight that dataset assumptions made about pre-industrial sea-
493 ice concentrations in Antarctica can have substantial impacts on diagnosed feedbacks in AGCMs and
494 remains an outstanding uncertainty in assess total feedback differences. Fortunately, in *amip-*
495 *piForcing* the difference in SW clear-sky feedback (which will be strongly impacted on by sea-ice
496 feedbacks) is similar to that seen in $\lambda_{4\times\text{CO}_2}$ (Figure 3) so this can be ignored if the focus is solely on SST
497 driven feedbacks in the atmosphere.

498 In summary, for warming since the 1800s (using either 1871-2010 or 1871-2014), both *amip-*
499 *piForcing* and *hadSST-piForcing* suggest a substantial pattern effect between radiative feedbacks
500 operating over historical climate change and long-term ECS.

501 3.2 Considering the historical record before and after 1980

502 We now return the divergence in dN response between *amip-piForcing* and *hadSST-piForcing*
503 simulations around 1980 (Figure 1d). As well as the change in behaviour discussed above, 1980
504 provides a convenient separation of historical feedbacks and the pattern effect for two other

505 motivating reasons: (i) Fueglistaler and Silvers (2021) identify ~1980 as the point in which the Earth
506 begins to warm with a particular configuration of tropical Pacific SSTs where regions of deep
507 convection warm substantially more than the tropical mean, driving large negative cloud feedbacks
508 and consistent with a large pattern effect over this period (Gregory and Andrews 2016; Zhou et al.,
509 2016; Andrews et al., 2018; Gregory et al. 2020); and (ii) Fueglistaler and Silvers (2021) also identify
510 ~1980 as a useful approximation of when the satellite era was integrated into the global observing
511 system, and so developing an understanding of feedbacks and the pattern effect specifically from
512 1980 onwards will aid interpretation of our most comprehensive observations of climate change and
513 how they might relate to the future change (next Section).

514 Figure 4 compares the surface temperature trend over the two time-periods 1871-1980 and 1981-
515 2010 in *amip-piForcing* and *hadSST-piForcing*. Differences between the two SST reconstructions are
516 extremely subtle. For the earlier 1871-1980 time period, warming is more uniform, in part because
517 of the longer time-period considered which will smooth out variability. Since 1981 in contrast, there
518 has been strong western Pacific warming with eastern Pacific cooling, despite temperatures
519 increasing in the global mean. Hence, we might expect a small pattern effect prior to 1980 and a
520 large pattern effect post 1980 (e.g. Gregory and Andrews, 2016; Zhou et al., 2016, Andrews and
521 Webb, 2018, Ceppi and Gregory, 2017; Dong et al. 2019, Fueglistaler and Silvers 2021).

522 Figures 1g and 1h show the $\lambda_{\text{hist}} = dN/dT$ relationship in the ensemble-mean *amip-piForcing* and
523 *hadSST-piForcing* simulation for 1871-1980 (grey points) and 1981-2010 (blue points). Results for
524 individual models are given in Table 2. Figures 1g and 1h confirms the basic premise that λ_{hist}
525 strengthens in magnitude post 1980, consistent with the change in SST patterns (Figure 4).

526 For the earlier time-period, 1871-1980, $\lambda_{\text{hist}} = -1.14 \pm 0.33 \text{ W m}^{-2} \text{ K}^{-1}$ in *amip-piForcing* is similar to λ_{hist}
527 $= -1.25 \pm 0.37 \text{ W m}^{-2} \text{ K}^{-1}$ in *hadSST-piForcing* (Table 2) – suggesting little sensitivity of the results to
528 these two SST datasets over this time period. This is unsurprising given that the datasets are similar
529 (though not identical) prior to this period (Section 2.2 and Figure 4). For the eight AGCMs that
530 performed both simulations Figure 2a shows the relationship between λ_{hist} in *amip-piForcing* and
531 *hadSST-piForcing*. For all time-periods λ_{hist} in *amip-piForcing* and *hadSST-piForcing* are found to be
532 well correlated ($r \geq 0.84$, Figure 2a). For the earlier 1871-1980 results, the λ_{hist} values fall close to the
533 one-to-one line (blue dots, Figure 2) and within the range of $\lambda_{4\times\text{CO}_2}$ (grey shaded areas in Figure 2).
534 This suggests that for 1871-1980 λ_{hist} is broadly independent of the two SST datasets (consistent with
535 their common basis) and that the pattern effect is small for this time period. Indeed, the 1871-1980
536 pattern effect is small but positive ($\Delta\lambda = 0.19 \pm 0.35 \text{ W m}^{-2} \text{ K}^{-1}$ in *amip-piForcing* and $0.26 \pm 0.28 \text{ W m}^{-2}$
537 K^{-1} in *hadSST-piForcing*, Table 3 and Figure 2b).

538 In contrast, for 1981 onwards (i.e. 1981-2010), λ_{hist} is generally far from the $\lambda_{4\times\text{CO}_2}$ range (i.e. a large
539 pattern effect) and away from the one-to-one line (i.e. a dependence on the SST dataset) (Figure 2a;
540 grey points). Indeed, λ_{hist} over 1981-2010 is substantially stronger in magnitude than over 1871-1980
541 ($\lambda_{\text{hist}} = -2.33 \pm 0.72 \text{ W m}^{-2} \text{ K}^{-1}$ in *amip-piForcing* over 1981-2010, Table 2; Figure 2a) and the pattern
542 effect is large ($\Delta\lambda = 1.38 \pm 0.75 \text{ W m}^{-2} \text{ K}^{-1}$, Table 3; Figure 2b), although somewhat weaker in
543 magnitude in *hadSST-piForcing* ($\Delta\lambda = 1.12 \pm 0.69 \text{ W m}^{-2} \text{ K}^{-1}$, Table 3; Figure 2b). For 1981-2010, λ_{hist} is
544 generally weaker in *hadSST-piForcing* (Table 2; Figure 3a) by $0.26 \pm 0.48 \text{ W m}^{-2} \text{ K}^{-1}$ across the eight
545 AGCMs using both SST datasets.

546 These results are generally consistent with Fueglistaler and Silvers (2021) and Lewis and Mauritsen
547 (2021) who both point to the AMIP II SST dataset as having larger (relative) western tropical Pacific
548 warming than in other SST datasets, and hence from the process understanding we would expect a
549 more negative feedback (and larger pattern effect) in *amip-piForcing*, as found above. The one

550 exception is GFDL-AM4, which simulates a more negative λ_{hist} under HadISST1 SSTs than AMIP II
551 from 1981-2010, and so a larger pattern-effect over this period under HadISST1 SSTs (Tables 2 and 3
552 and the single grey dots in Figures 2a and 2b which sit on the other side of the one-to-one line from
553 the other models). The reasons for this remain unclear.

554 In summary we have shown that a division around 1980 usefully separates historical climate change
555 into two time-periods: (i) pre 1981 the Earth warmed over most of the historical record with an
556 averaged warming pattern that is relatively uniform, and feedbacks largely consistent with long-term
557 ECS feedbacks (i.e. a relatively small pattern effect), and (ii) post 1980 where the Earth warmed with
558 a particular configuration of strong SST gradients that drove feedbacks much more stabilizing than
559 those seen in long-term ECS feedbacks (i.e. large pattern effect), albeit with a sensitivity of the
560 magnitude of this result to the SST dataset considered.

561 *3.3 Relationships between historical and ECS feedbacks*

562 We now consider whether feedbacks over the historical period in *amip-piForcing* are related to
563 $\lambda_{4\times\text{CO}_2}$. This is in contrast to the previous sections which only quantified their difference (i.e. the
564 pattern effect).

565 Firstly, we note that the spread in feedbacks across models over the earlier (1871-1980) time-period
566 in *amip-piForcing* are well correlated with the spread in feedbacks across models in *abrupt-4xCO₂*.
567 ($r=0.69$, Figure 5a). In contrast, feedbacks over the most recent decades (1981-2010) are only weakly
568 correlated with $\lambda_{4\times\text{CO}_2}$ ($r=0.27$). Secondly, feedback over the full historical record (1871-2010) is only
569 weakly correlated with feedback from the 1871-1980 time-period ($r=0.45$, Figure 5b). In contrast,
570 1871-2010 feedback is strongly correlated with feedback over the most recent 1980-2010 decades
571 ($r=0.91$, Figure 4b). This strong correlation between 1981-2010 and the 1871-2010 feedback arises
572 because the spread for 1871-2010 is dominated by the spread for 1981-2010.

573 Given that the feedbacks applying in 1871-1980 and in 1981-2010 are different, we infer that
574 variation in the pattern of SST over these two periods is dominated by different effects. Because the
575 feedbacks of 1871-1980 are correlated with *abrupt-4xCO₂*, the difference between the two periods
576 could be explained by CO₂ being the dominant influence in 1871-1980 SST patterns, while something
577 else (e.g. perhaps variability, aerosol, volcanism) dominates during 1981-2010. This is only a
578 hypothesis, because these experiments do not provide a way to attribute the observed SST changes
579 to causes.

580 The result is that the spread in feedbacks over the full historical record are only weakly correlated
581 with $\lambda_{4\times\text{CO}_2}$ ($r=0.51$, Figure 3), because of the strong pattern effect post 1980. Hence, we can say little
582 about future $\lambda_{4\times\text{CO}_2}$ directly from climate change post 1980 or even the full historical record without
583 adjusting for a pattern effect. In contrast, the feedbacks operating over the earlier 1871-1980 time-
584 period are correlated with $\lambda_{4\times\text{CO}_2}$ ($r=0.69$, Figure 5a), but here the climate change signal is smaller and
585 the observations poorer which limits the utility of this time-period to act as an observational
586 constraint.

587 That recent decadal feedbacks are the most unrepresentative of the long-term climate sensitivity is
588 unfortunate, not just because it coincides with the advent of the satellite record and so is extremely
589 well observed, but also because climate change since ~1980 ought to provide the best constraint on
590 ECS (e.g. Jiménez-de-la-Cuesta and Mauritsen, 2019). This is because it offers a strong global
591 warming signal, which AOGCMs attribute to greenhouse gas increases, while avoiding the
592 uncertainty due to aerosol radiative forcing, which has only changed slowly over this period (at least
593 globally, strong regional changes may have impacted on SST patterns, e.g. Smith et al. 2015;

594 Takahashi & Watanabe, 2016; Moseid et al., 2020). Although feedbacks operating over the earlier
595 1871-1980 part of the historical record are correlated with long-term CO₂ induced feedbacks, a
596 reliable observational constraint is harder because the climate change signal is smaller and the
597 observations poorer. We discuss this further in the Discussion section.

598 Up to now we have only considered a comparison of *amip-piForcing* feedbacks to a single definition
599 of *abrupt-4xCO2* feedbacks (i.e. feedbacks diagnosed over years 1-150 in *abrupt-4xCO2*). Here we
600 briefly consider separating λ_{4xCO_2} into the two principal timescales of the *abrupt-4xCO2* response
601 following Andrews et al. (2015) by calculating λ_{4xCO_2} over years 1-20 (a fast timescale) and 21-150 (a
602 slow timescale) (Table 2). The rationale is that 20 years is approximately the timescale required for
603 the mixed-layer to equilibrate in response to step forcing, and any subsequent climate response
604 scaling with the slower deep-ocean timescale, as approximated by two-layer models (Held et al.,
605 2010; Geoffroy et al., 2013; Gregory et al., 2015).

606 Figure 5c shows λ_{hist} from 1871-1980 is largely scattered about the one-to-one line with λ_{4xCO_2} from
607 years 1-20, suggesting little to no pattern effect between these two. This is potentially consistent
608 with the historical record largely being the result of the faster timescale responses (Held et al. 2010;
609 Proistosescu & Huybers, 2017). In contrast, post-1980 λ_{hist} is far from the one-to-one line (i.e. large
610 pattern effect to years 1-20 of *abrupt-4xCO2*, Figure 5c) but is marginally correlated ($r=0.53$),
611 suggesting recent decades do contain some information relevant to the feedback seen in the fast
612 timescale response to CO₂. However, the longer-term feedbacks associated with the slow timescale
613 response to CO₂ (years 21-150 of *abrupt-4xCO2*, Figure 5d) have no correlation with λ_{hist} post-1980
614 ($r=-0.06$, Figure 5d). This is not surprising given that the eastern tropical Pacific and Southern Ocean
615 have largely cooled in the post-1980 period, while they warm substantially over years 21-150 of
616 *abrupt-4xCO2*.

617 618 3.4 Decadal variability in feedbacks and the pattern effect 619

620 In this final section of GCM results we present how λ_{hist} and the pattern effect varies on decadal
621 timescales in the *amip-piForcing* and *hadSST-piForcing* simulations.

622 Following Gregory and Andrews (2016) we calculate $\lambda_{hist} = dN/dT$ over a moving 30 year window in
623 the *amip-piForcing* and *hadSST-piForcing* simulations (Figure 6a and b). For example λ_{hist} calculated
624 over the 30 year period 1925 to 1954 is presented at year 1939.5 in Figure 6. In Figures 6c-h the LW
625 and SW clear-sky and cloud radiative effect of the feedback are also shown. The correlation
626 coefficient between the *amip-piForcing* and *hadSST-piForcing* multi-model-mean λ_{hist} timeseries is
627 0.84, suggesting the broad features of the decadal λ_{hist} variations are robust to the SST datasets. In
628 particular λ_{hist} peaks (least negative, smallest pattern effect) around 1940 while generally being large
629 in magnitude (large pattern effect) over recent decades (see also Gregory and Andrews, 2016; Zhou
630 et al. 2016; Andrews et al. 2018; Gregory et al. 2020). The clear sky feedbacks (Figures 6c-f) are
631 largely stable, while the variation in λ_{hist} is almost entirely explained by variation in cloud feedback
632 (Figures 6g-h), consistent with previous findings (e.g. Zhou et al. 2016; Andrews et al. 2018).

633 In Section 5, we discuss further the reasons for the decadal variations in SST patterns and λ_{hist} , i.e.
634 whether they are the result of spatiotemporal changes in forcings such as aerosols or volcanic
635 forcing or due to unforced variability.

636

637

638 4. Observed climate change

639 We next consider whether the radiative feedback and pattern effects simulated by the GCMs are
640 consistent with observed variations in the Earth’s energy budget. Gregory et al. (2020) asked a
641 similar question for the post 1980 period and suggested they are (see their Figure 5c), but here we
642 go a few steps further. Specifically, not only do we consider the post 1980 period, but also assess
643 changes in the Earth’s energy budget back to the 1800s. Furthermore we investigate the implications
644 of a strongly negatively feedback parameter (large pattern effect) since 1985 on the observed rate of
645 global warming.

646 The observations also provide an opportunity to bring our λ_{hist} and pattern effect estimate up to date
647 with the most recently observed data (up to and including 2019), whereas our GCM analysis
648 generally finished in 2014. The observations post 2014 period are of particular interest given they
649 include the major El-Nino event of 2015/2016 that was associated with eastern-pacific warming and
650 marked changes in the observed radiation budget (Loeb et al. 2020; 2021). We expect these post
651 2014 years to have an impact λ_{hist} and the pattern effect, given the process understanding discussed
652 previously (e.g. Zhou et al., 2016, Andrews and Webb, 2018, Ceppi and Gregory, 2017; Dong et al.
653 2019).

654 4.1 Comparison of AGCM results to observed estimates

655 We first validate the AGCM λ_{hist} estimates over recent decades. To do this we use a merged satellite
656 dataset (ERBE WFOV + CERES) (Allan et al. 2014) that provides an observational estimate of dN
657 variations from 1985 to 2019. For dT we use the HadCRUT5 analysis dataset (Morice et al. 2021). For
658 dF we use the IPCC AR6 (Forster et al. 2021) best estimate historical ERF changes. These datasets are
659 described in further detail in Section 2.4. We first consider the 30-year period 1985 to 2014,
660 consistent with many of the AGCMs.

661 Figure 7a and 7b show the dT , dN and dF timeseries over this period. The 1985-2014 ‘observed’ $-\lambda_{\text{hist}}$
662 $= d(F - N)/dT \sim 2.0 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$ relationship is shown in Figure 7d. Note the stated 5-95%
663 uncertainty is $\pm 1.645\sigma$ from the standard error of the linear fit, with no allowance for systematic
664 uncertainties. As discussed in Section 2.4, observed multi-decadal changes in dN are subject to a
665 substantial uncertainty (up to 0.5 W m^{-2}) primarily related to the breaks in the record prior to 2000,
666 though are considerably smaller afterwards (Liu et al. 2020). Note also that years 1991-2 are
667 excluded from the calculation as these years are identified as being strongly impacted by the
668 volcanic forcing from the Pinatubo eruption (Figure 7b). Whilst λ_{hist} is robust to this (we get just the
669 same $\lambda_{\text{hist}} \sim -2.0 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$ if we include these years), including these years has an impact on the
670 ocean heat uptake efficiency estimate (see Section 4.3). The observed 1985-2014 λ_{hist} estimate is
671 shown on Figure 6a and 6b (red line) as an illustration in comparison to the AGCM decadal variations
672 in λ_{hist} . The observed λ_{hist} best estimate agrees exceptionally well with the AGCM multi-model mean,
673 and nearly all models are within the 5-95% uncertainty estimate as they approach the 1985-2014
674 value (Figure 6a and 6b).

675 A more rigorous comparison of individual AGCM results to the observed estimate is shown in Figure
676 8. Here the AGCM λ_{hist} estimates from *amip-piForcing* and *hadSST-piForcing* have been calculated in
677 the same way as the observations, i.e. over 1985-2014 excluding 1991-2. The overlap between the
678 model and observed estimates points to broad consistency between the models and observations in
679 the recent decadal value of λ_{hist} (Figure 8). The large uncertainties (which are likely underestimated
680 since we have not accounted for structural errors) inhibit a more precise validation of individual
681 models against the observed estimate.

682 For the full the historical record we estimate λ_{hist} from IPCC AR6 assessed changes in T , N and F .
683 Forster et al. (2021) give these as $\Delta T = 1.03 \pm 0.20$ K, $\Delta N = 0.59 \pm 0.35$ W m⁻² and $\Delta F = 2.20$ [1.53 to
684 2.91] W m⁻² for the time-period 1850-1900 to 2006-2019. For simplicity we assume $\Delta F = 2.20 \pm 0.7$ W
685 m⁻², where we have approximated the uncertainty in ΔF as a Gaussian. Randomly sampling (with
686 replacement) from the Gaussian distributions in ΔN , ΔF and ΔT gives $\lambda_{\text{hist}} = (\Delta N - \Delta F)/\Delta T = -1.6 \pm 0.8$
687 W m⁻² K⁻¹. This is again in agreement with the *amip-piForcing* ($\lambda_{\text{hist}} = -1.65 \pm 0.46$ W m⁻² K⁻¹, Table 2)
688 and *hadSST-piForcing* ($\lambda_{\text{hist}} = -1.43 \pm 0.43$ W m⁻² K⁻¹, Table 2) 1871-2010 ensembles, though an exact
689 match is not expected given the slightly different time-periods and methods (e.g. finite differences
690 versus regression) used. Still, the agreement provides further confidence in the GCM's simulated
691 radiative response to observed SST and sea-ice variations over the historical record, and strengthens
692 the conclusion that λ_{hist} has become more negative over recent decades compared to the longer
693 1871-2010 time-period.

694 Finally, IPCC AR6 assessed the long-term ECS relevant feedback parameter (analogous to our $\lambda_{4\times\text{CO}_2}$)
695 to be -1.16 ± 0.65 W m⁻² K⁻¹ (Forster et al., 2021) by combining lines of evidence from observations,
696 theory, process models and GCMs on individual climate feedback processes. Combining this with our
697 observed λ_{hist} estimates above gives an estimate of the pattern effect independently of our GCM
698 ensemble. This gives an estimated pattern effect of $\sim 0.8 \pm 1.0$ W m⁻² K⁻¹ for 1985-2015 and $\sim 0.4 \pm 1.1$
699 W m⁻² K⁻¹ for the full historical record (the 1850-1900 to 2006-2019 changes). While the uncertainties
700 are substantial, there is again agreement with our GCM results.

701 4.2 Recent observed trends and the efficiency of ocean heat uptake

702 We have seen that both models and observed variations in the Earth's energy budget agree on the
703 Earth having had strongly stabilizing feedbacks over recent decades relative to AOGCM feedbacks
704 under long-term CO₂ forced climate change. Quantifying this in a different way, a feedback
705 parameter of ~ -2.0 W m⁻² K⁻¹ suggests an EffCS = $-F_{2\times}/\lambda_{\text{hist}}$ as low as $\sim 4.0/2.0 \sim 2.0$ K operating over
706 1985-2014, assuming $F_{2\times} = 4.0$ W m⁻² (Sherwood et al. 2020). From this it seems possible that the
707 rate of global warming over this period (~ 0.19 K dec⁻¹, Tokarska et al., 2020) might have been larger
708 had the Earth warmed over this period with a pattern of SST associated with more positive
709 feedbacks, as found in earlier parts of the historical record (Section 3). However, we also investigate
710 the possibility that changes in ocean heat uptake efficiency may have compensated the changes in
711 feedbacks and low EffCS to maintain the higher warming rate over this period.

712 To do this we turn to the 'climate resistance' (ρ , units W m⁻² K⁻¹) "zero-layer" model of Gregory and
713 Forster (2008) to analyse the ocean heat uptake efficiency (κ , units W m⁻² K⁻¹). This is expressed as
714 $dF = \rho dT$, where $\rho = \kappa - \lambda$, and κ is defined as $\kappa = dN/dT$ and is found to be strongly related to the
715 thermal coupling constant (γ , units W m⁻² K⁻¹) between the upper and lower ocean in the two-layer
716 model (Gregory et al. 2015; see their Figure 8). While the zero-layer model is a gross simplification of
717 the climate system (we discuss potential limitations below), $dF = \rho dT$ is found to be an excellent
718 approximation ($r=0.86$) over 1985 – 2014 (excluding the 1991-2 Pinatubo years, see below) in our
719 data (Figure 7c). From this relationship we deduce $\rho = dF/dT \sim 2.4 \pm 0.5$ W m⁻² K⁻¹ over 1985-2014
720 (Figure 7c) and similarly $\kappa = dN/dT \sim 0.4 \pm 0.8$ W m⁻² K⁻¹. In contrast, AOGCM simulations of steady
721 increasing CO₂ generally have a larger ocean heat uptake efficiency ($\kappa = 0.73 \pm 0.18$ W m⁻² K⁻¹ for
722 years 61-80 of CMIP5 1%CO₂ AOGCM simulations, Gregory et al., 2015).

723 Another effect on surface temperature to consider is the possibility that the pattern of surface
724 warming and/or atmospheric circulation may change the efficiency of global heat uptake, thus not
725 only is λ inconstant, but κ may also vary too. Using passive ocean uptake experiments wherein
726 ocean circulation cannot change, Newsom et al. (2020) find that ocean heat uptake efficiency can be

727 expected to be smaller when warming is enhanced in the tropics (where deep ocean ventilation is
728 small) and larger when warming is enhanced in the high latitudes (where deep ocean ventilation is
729 large). With relatively small warming in the southern high latitudes, this suggests that the
730 surface/ocean-mixed layer might have been less efficient at fluxing heat into the deep ocean over
731 the same period as the large pattern effect, potentially enhancing global surface warming and
732 muting some of the impact of feedback changes. However, stronger trade winds, as have been
733 observed over 1981-2010, can also be expected to accelerate subtropical cells, enhancing ocean
734 heat uptake efficiency and slowing global surface warming (England et al. 2014), an effect not
735 accounted for in the passive ocean heat uptake experiments of Newsom et al. (2020). Thus,
736 variations in both radiative feedbacks and ocean heat uptake appear to be physically linked through
737 SST patterns and may even to some extent co-vary (Newsom et al. 2020).

738 As our dN timeseries does not predate 1985 we cannot investigate whether κ has varied in a way
739 that would counter changes in λ_{hist} prior to 1985. Instead, we go forward in time exploiting the
740 datasets up to and including 2019. This includes the major El-Nino event of 2015/2016 and marked
741 changes in the observed radiation budget (Loeb et al. 2020; 2021). Figure 9 illustrates the impact of
742 this event on the pattern of decadal surface warming trends. Over 1985-2014 there is marked
743 cooling over the eastern Pacific (Figure 9a) which is much reduced when the pattern is calculated
744 over 1987-2016 (Figure 9b) to include the peak 2015-16 El-Nino years. The difference (Figure 9c)
745 shows the warming event of the 2015-16 El-Nino on the eastern Pacific, while cooling in the western
746 Pacific, as well as a slight reduction in Southern Ocean cooling. This is precisely the pattern of SST
747 change we'd expect to have an impact on λ .

748 Table 4 shows the impact on 30-year derived ρ , λ and κ values moving forward in time from 2014, up
749 to and including 1990-2019. Figure 7 (red crosses) shows these additional 5 years in comparison to
750 the 1985-2014 ρ and λ relationships. Post 2014, λ reduces in magnitude (Table 4) and all the red
751 crosses fall below the 1985-2014 λ relationship in Figure 7d. This is consistent with process based
752 arguments that the shift to eastern Pacific warming post 2014 ought to drive more positive
753 feedbacks and consequently a reduction of the pattern effect over these years. If the AGCM
754 simulations were extended to cover this time-period we ought to expect them to simulate a similar
755 response. This would be worthwhile future work complementary to Loeb et al. (2020).

756 In contrast to λ , ρ is relatively stable to these additional years (Table 4) and the 1985-2014 ρ
757 relationship is found to be an excellent predictor for 2015-2019 (red crosses fall on or close to the
758 line, Figure 7c). A consequence of ρ being well approximated as constant but λ not, is that κ (equal to
759 $\rho + \lambda$) must compensate for the change in λ . Thus beyond 2014, the pattern effect declines but its
760 impact on surface temperature is buffered by a change in ocean heat uptake efficiency. This is
761 consistent with the original hypothesis that variations in SST patterns affect both heat loss to space
762 (radiative feedbacks) and the efficiency of heat uptake into the deep-ocean in a way that might co-
763 vary (Newsom et al., 2020). However, the extent of any anti-correlation is unclear, it may simply
764 apply to short term variability. It clearly does not apply to longer-term forced changes, given that
765 Gregory et al. (2015) found substantial variations in ρ , which would not occur if κ and λ were
766 strongly anti-correlated.

767 While the zero-layer model appears to work well on this short timescale (Figure 7c) we caution
768 against assuming all changes in ocean heat content are driven by global T , as assumed by the $dN =$
769 κdT relationship. This is because, especially on short timescales, other influences that do not
770 correlate with global T , such as wind-driven ocean circulation changes perhaps, will also alter ocean
771 heat content (England et al., 2014). In such a situation, it would be reasonable to write $N = \kappa T + U$
772 where U is an additional term to the heat balance, not related to global T . This implies $\kappa = N/T - U/T$,

773 and including this term in the forced heat balance, $N = F + \lambda T + U$, gives $\lambda = (N-F)/T - U/T$. Thus U/T
774 would perturb the estimate of κ (a positive number) and λ (a negative number) in opposite
775 directions, as we see in our data. Hence our results are potentially evidence for variation in ocean
776 heat content not driven by global T , but we cannot say exactly what it is – other than it does not
777 scale with global T .

778 We caution that structural errors could impact on our diagnosis. Specifically, both κ and λ are related
779 to dN and so any bias or error in the observed dN trend would bias κ and λ in opposite directions.
780 Moreover $\rho = dF/dT$ would be unaffected by any bias or error in dN , and so the anti-correlation would
781 compensate to leave $\rho = \kappa - \lambda$ unaffected. We illustrate this in Table 4, which shows these quantities
782 calculated over 1985-2014 using 5 available different versions of the DEEP-C dN datasets (see
783 Section 2.4). Differences in the results emerge (λ reduces in magnitude from $\sim -2.2 \text{ W m}^{-2} \text{ K}^{-1}$ to ~ -2.0
784 $\text{W m}^{-2} \text{ K}^{-1}$, with a compensating increase in κ) as the DEEP-C datasets transition from v3 to v4 (i.e. v2
785 and v3 give the same results, as do v4 and v5), highlighting the impact of potential structural errors
786 in these results. We do not pursue the cause of the difference in the results, but it is likely due to
787 changes between v3 and v4 in how the DEEP-C method bridges the gap between satellite products in
788 the 1990s (a longer adjustment period and a different modelling ensemble is used) (Liu et al., 2020).
789 However it is also important to note that the observational record since 2000, applying the CERES
790 dataset, is subject to much smaller structural uncertainty than the earlier record implying a greater
791 confidence in our analysis of the anomalous N variations post 2014.

792 4.3 Effect of the Pinatubo volcanic eruption

793 Finally, we comment on the effect of the Pinatubo volcanic eruption on these results. There is a large
794 negative spike in dF and dN around 1991 and 1992 (Figure 7b). While we found no impact of these
795 years on our estimate of 1985 – 2014 λ_{hist} , they have a strong impact on ρ and κ . Including these
796 years in the regression analysis, we find $\rho = dF/dT \sim 2.9 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$ and $\kappa = dN/dT \sim 0.8 \pm 0.9 \text{ W}$
797 $\text{m}^{-2} \text{ K}^{-1}$, much larger than when these years are excluded from the analysis as above. This is
798 consistent with Gregory et al. (2015) who found the ‘transient climate response parameter’ (equal to
799 $1/\rho$, units $\text{K W}^{-1} \text{ m}^2$) to explosive eruptions to be smaller (ρ larger) than that evaluated in AOGCMs
800 under steadily increasing CO_2 , principally because the surface/mixed-layer readily gives up heat (κ
801 larger) in response to a short-lived forcing like an explosive volcanic eruption. Hence if the time-
802 period under consideration contains large volcanic eruptions then the “zero-layer” model ($dF = \rho dT$)
803 is found to be a poor approximation (i.e. ρ not constant) over the entire time-period because it
804 neglects the importance of the upper-ocean heat capacity on short timescales (Gregory and Forster,
805 2008; Held et al. 2010; Gregory et al., 2016). This manifests itself as a sensitivity of ρ and κ to the
806 inclusion or exclusion of volcanic years, as we have found here.

807

808 5. Summary, Discussion and Conclusions

809 5.1 Historical feedbacks and the pattern effect

810 The dependence of radiative feedback on the pattern of SST change was investigated in fourteen
811 Atmospheric General Circulation Models (AGCMs) forced with observed variations in sea-surface-
812 temperature (SST) and sea-ice over the historical record from 1871 to near-present (*amip-piForcing*
813 experiment). We found that the pattern effect identified in a previous model intercomparison
814 (Andrews et al, 2018) is largely robust to a wider set of new generation AGCMs with a broader range
815 of atmospheric physics and climate sensitivities. Our qualitative conclusions were not strongly
816 dependent on the AMIP II SST dataset used to force the AGCMs; indeed, the feedbacks in eight

817 AGCMs using SSTs from HadISST1 (*hadSST-piForcing*) were found to be strongly correlated with
818 feedbacks in *amip-piForcing*, though the magnitude of the pattern effect post 1980 was found to be
819 smaller under HadISST1 SSTs (see also Andrews et al., 2018; Lewis and Mauritsen, 2021; Zhou et al.,
820 2021; Fueglistaler and Silvers, 2021).

821 Separating the historical record at 1980, we found that over 1871-1980 the Earth warmed with a
822 relatively uniform warming pattern and feedbacks largely consistent and strongly correlated with
823 long-term *abrupt-4xCO2* feedbacks (i.e. with relatively small pattern effect - Figures 2 and 5). In
824 contrast, post 1980 the Earth warmed with a strong tropical Pacific SST gradient (Figure 4) where
825 regions of deep convection warm substantially more than the tropical mean (Fueglistaler and Silvers,
826 2021). This drove large negative feedbacks and pattern effects in both our *amip-piForcing* and
827 *hadSST-piForcing* simulations, consistent with the physical understanding of how lapse-rate and
828 cloud feedbacks depend on tropical Pacific SST patterns (Zhou et al., 2016; Andrews and Webb,
829 2018; Ceppi and Gregory, 2017; Dong et al., 2019).

830 As well as a large pattern effect, feedbacks post 1980 were found to be uncorrelated with long term
831 CO₂ driven feedbacks (Figure 5). This is unfortunate, because the feedback inferred from this period
832 therefore does not constrain the CO₂ feedback or ECS. It is also surprising, because the period since
833 ~1980 contains a well observed large global temperature response, which AOGCMs attribute to
834 increasing greenhouse gases, and it avoids the aerosol forcing uncertainty issue (Jiménez-de-la-
835 Cuesta and Mauritsen, 2019). Despite this, it turns out to be the worst period for inferring the
836 Earth's long-term CO₂ climate sensitivity from the observed global energy balance. Conversely,
837 feedbacks acting earlier in the record (1871-1980) are representative of the long-term response (i.e.
838 smaller pattern effect) and do correlate with λ_{4xCO_2} across models, yet this period has a smaller
839 climate change signal and is not as well observed, containing much larger uncertainties relative to
840 the climate change signal (e.g. Otto et al., 2013), as well as a large forcing uncertainty. Hence the
841 usefulness of this time-period is limited for setting a constraint on λ_{hist} .

842 Considering the historical record as a whole is useful for informing studies that use the entire
843 observed record to estimate ECS via energy budget constraints (e.g. Sherwood et al. 2020). We
844 found that the pattern effect over 1871-2010 to be $\Delta\lambda = 0.70 \pm 0.47 \text{ W m}^{-2} \text{ K}^{-1}$ in our *amip-piForcing*
845 ensemble and $\Delta\lambda = 0.44 \pm 0.31 \text{ W m}^{-2} \text{ K}^{-1}$ in *hadSST-piForcing*, where the smaller uncertainty in
846 *hadSST-piForcing* likely reflects the narrower set of model physics in this smaller ensemble (we do
847 not have *hadSST-piForcing* experiments for the models with either the largest (CESM2) or smallest
848 (MIROC6) pattern effects in *amip-piForcing*). The question therefore arises as to which of these
849 estimates ought to be used for adjusting historical energy budget constraints on ECS for pattern
850 effects.

851 Both Lewis and Mauritsen (2021) and Fueglistaler and Silvers (2021) showed that the AMIP II dataset
852 had the largest warm pool trends relative to the tropical-mean of all SST reconstructions they
853 considered. Hence one interpretation of our results is that the pattern effect in *amip-piForcing* might
854 usefully be regarded as an upper bound on the structural uncertainty of the experimental design to
855 observational uncertainty in SST reconstructions. A best estimate might place more weight on the
856 *hadSST-piForcing* pattern effects, which have warm pool trends (relative to the tropical-mean) closer
857 to the middle of the range of SST reconstructions (Fueglistaler and Silvers, 2021; Lewis and
858 Mauritsen, 2021). In that case, we recommend a best estimate of the historical pattern effect of
859 $0.44 \pm 0.47 \text{ W m}^{-2} \text{ K}^{-1}$ for the time-period 1871-2010, which represents the pattern effect from
860 *hadSST-piForcing* but retaining the larger uncertainty from the (larger ensemble) *amip-piForcing*
861 results. If calculated over 1871-2014 the pattern effect increases by $0.05 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$ according
862 to the *hadSST-piForcing* ensemble. This best estimate of the historical pattern effect is close to that

863 used in Sherwood et al. (2020), who assumed a value of $0.5 \pm 0.5 \text{ W m}^{-2} \text{ K}^{-1}$ (they were informed by
864 Andrews et al. (2018) but allowed for a potentially smaller pattern effect than that study based on
865 expert judgement). In the future, a model intercomparison of the pattern effect to a broader range
866 of SST datasets would be useful to address any outstanding structural uncertainty to SST
867 reconstructions.

868 To provide independent evidence for the historical pattern effect, we used IPCC AR6 assessed
869 changes in T , N and F between 1850-1900 to 2006-2019 (Forster et al. 2021) to estimate a historical
870 feedback parameter of $\lambda_{\text{hist}} = (\Delta N - \Delta F)/\Delta T = -1.6 \pm 0.8 \text{ W m}^{-2} \text{ K}^{-1}$. This was found to be in agreement
871 with the *amip-piForcing* and *hadSST-piForcing* ensembles. IPCC AR6 also assessed the long-term ECS
872 relevant feedback parameter ($-1.16 \pm 0.65 \text{ W m}^{-2} \text{ K}^{-1}$, Forster et al., 2021) from combining lines of
873 evidence from observations, theory, process models and GCMs on individual climate feedback
874 processes. Contrasting this with the λ_{hist} estimate above gives an estimate of the pattern effect of 0.4
875 $\pm 1.1 \text{ W m}^{-2} \text{ K}^{-1}$ for historical changes between 1850-1900 to 2006-2019. While the uncertainties are
876 substantial, this is in agreement with our GCM based estimate of the historical pattern effect.

877 5.2 Observed climate change since 1985 and ocean heat uptake efficiency

878 Satellite based reconstructions of the Earth's energy balance over 1985 to 2014 suggest a feedback
879 parameter of $\sim -2.0 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$, in agreement with our *amip-piForcing* and *hadSST-piForcing*
880 ensembles. Evidence is also emerging from satellite records in support of the physical processes and
881 mechanisms of the pattern effect between surface temperature, atmospheric stability, cloudiness
882 and radiative fluxes over recent decades (e.g. Zhou et al., 2016; Ceppi and Gregory, 2017; Loeb et al.,
883 2020; Fueglistaler and Silvers, 2021; Ceppi and Fueglistaler, 2021).

884 Extending our analysis post 2014 included the major El-Nino event of 2015/2016 that was associated
885 with eastern-pacific warming and marked changes in the observed radiation budget (Loeb et al.
886 2020; 2021). Including these post 2014 years (up to and including 2019) reduced the magnitude of
887 the observed λ estimate, consistent with eastern Pacific warming driving more positive feedbacks (as
888 also suggested in Loeb et al., 2020). This suggests the pattern effect that has existed over recent
889 decades may be waning if a shift from western to eastern Pacific warming is maintained in the
890 longer term, as might be expected from a change in the PDO index identified by Loeb et al. (2020).

891 Given the substantial rate of global warming since 1985, what does the presence of a large pattern
892 effect imply for ocean heat uptake efficiency (κ)? We estimated $\kappa = dN/dT \sim 0.4 \pm 0.8 \text{ W m}^{-2} \text{ K}^{-1}$ over
893 1985-2014, which is smaller (but not necessarily inconsistent) with AOGCM simulations of steady
894 increasing CO_2 ($\kappa = 0.73 \pm 0.18 \text{ W m}^{-2} \text{ K}^{-1}$ for years 61-80 of CMIP5 1% CO_2 AOGCM simulations,
895 Gregory et al. 2015). It raises the possibility that the pattern of surface warming and/or atmospheric
896 circulation may also change the efficiency of global heat uptake, thus both λ and κ might vary and to
897 some extent be related (Newsom et al., 2020). If an anti-correlation existed, it could buffer the
898 impact of a large pattern-effect on transient climate change.

899 We found that despite the change in radiative feedback post 2014 when the eastern Pacific warmed,
900 the climate resistance $\rho = dF/dT = \kappa - \lambda$ remained approximately constant, suggesting that κ and λ
901 co-varied. We showed that this result is potential evidence for a change in ocean heat content not
902 driven by global T . While this result is suggestive, the extent of this compensation and timescales it
903 applies to remains unclear. It may simply apply to short term variability and clearly does not apply to
904 longer-term forced changes (e.g. Gregory et al., 2015). Future research investigating how ocean
905 uptake efficiency and atmospheric radiative feedbacks are linked through patterns of SST change
906 would be useful.

907 5.4 Outlook and Implications for AOGCMs

908 Our results raise important questions for studies that have used emergent relationships from
909 AOGCMs to constrain ECS from recently observed decadal warming since ~1980 (e.g. Jiménez-de-la-
910 Cuesta and Mauritsen, 2019; Tokarska et al., 2020; Nijse et al., 2020).

911 Firstly, how is it possible that AOGCMs produce an emergent relationship between their recent
912 decadal warming trends and their ECS, while our results suggest that recent decadal feedbacks
913 ought to be unrelated to ECS? One solution to this conundrum is provided by Fueglistaler and Silvers
914 (2021), who showed that AOGCMs typically do not simulate the recent configuration of tropical
915 Pacific SST patterns that gave rise to the recent pattern effect (though some models do have broad
916 agreements, e.g. Olonscheck et al. 2021, Watanabe et al. 2021). Instead, the pattern of warming in
917 AOGCMs (and thus feedbacks) over recent decades is more similar to that seen in their *abrupt-*
918 *4xCO2* simulations (Gregory et al., 2020; Dong et al. 2021). Hence AOGCMs are generally biased in
919 their simulation of the recent decadal feedbacks and the pattern effect, compared to their
920 equivalent AGCMs forced with observed SST variations, as shown in Gregory et al. (2020) and Dong
921 et al. (2021).

922 If AOGCMs are biased in their simulation of recent decadal feedbacks and the pattern effect, it
923 suggests they may be biased toward simulating recent decadal temperature trends that are too high;
924 in turn, this would bias emergent constraints that use them toward values of ECS that are too low.
925 Alternatively, those models that do match the observed warming trend may do so via a
926 compensation of processes: too small a pattern effect balanced against too large a heat uptake into
927 the deep-ocean. Some evidence for the potential of this compensating behaviour is provided by
928 Hedemenn et al. (2017). Analysing the origins of decadal temperature variability in models, they
929 demonstrated an anti-correlation between the TOA radiative flux and deep-ocean (defined as below
930 100m) flux contributions to the model's surface layer and decadal temperature trends (see their
931 Figure 3). In other words, when the TOA radiative flux is in such a configuration to reduce its
932 contribution to the surface layer, then the surface/mixed-layer taps into the deep-ocean to
933 compensate for this loss, and vice versa. We speculate that such a configuration of TOA radiative flux
934 is potentially consistent with a large negative feedback, since in this configuration of atmospheric
935 feedbacks the surface efficiently radiates heat back to space. This again suggests a potential anti-
936 correlation between the ocean heat uptake efficiency and λ during unforced decadal variability
937 timescales as discussed previously.

938 Going forward, a critical question for future research is to understand what caused the particular
939 configuration of SST patterns over recent decades (e.g. strong warming in the western Pacific while
940 cooling in the eastern Pacific and Southern Ocean, despite temperature increasing in the global-
941 mean; Figure 4 and 9), and how might this pattern evolve in the future. For example, various
942 hypotheses have been put forward:

- 943 1. It could represent a mode of unforced coupled atmosphere-ocean variability (e.g.
944 Xie et al., 2016; Watanabe et al. 2021), albeit an unusual one is that is rarely
945 simulated by AOGCMs (Fueglistaler and Silvers, 2021). In this scenario, we might
946 expect the pattern effect to reduce in the near-future as the configuration of
947 tropical SST patterns shift to more warming in the east than the west. There is some
948 evidence (Loeb et al. 2020; 2021) this has already begun to happen in the most
949 recent years, as we have also shown. We might therefore, expect an acceleration of
950 warming trends, unless the additional heat at the surface from the reduced pattern

951 effect is tempered by compensating heat exchanges with the deep-ocean
952 (Hedemann et al. 2017).

- 953 2. Spatiotemporal variations in anthropogenic forcings such as aerosols (e.g., Smith et
954 al., 2015; Takahashi & Watanabe, 2016; Moseid et al., 2020; Heede and Fedorov,
955 2021) or explosive volcanic eruptions (Smith et al. 2015; Gregory et al. 2020) have
956 been implicated in driving tropical Pacific SST patterns. In these scenarios, the
957 pattern effect may decline with the reduction in aerosol emissions in the future, or
958 continue to have decadal variations associated with future volcanism. Whether
959 changes in deep-ocean fluxes will be accompanied with such forced changes in the
960 pattern effect is unclear.
- 961 3. While not explaining the eastern Pacific cooling per se, a delayed warming in the
962 eastern Pacific relative to the west is an expected transient response to forcing due
963 to the upwelling of (as yet) unperturbed waters from below (Clement et al., 1993;
964 Held et al. 2010; Heede and Fedorov, 2021). The implication of this is that
965 eventually the eastern Pacific will warm, and hence we might expect the pattern
966 effect to reduce and the Earth to warm with stronger (positive) cloud feedbacks.
- 967 4. In contrast, AOGCMs may overstate the expected warming in the eastern Pacific
968 (e.g. Seager et al., 2020). Under this scenario, we might expect the pattern effect to
969 reduce after the eastern Pacific stops cooling, but the full pattern effect according
970 to AOGCMs may never materialise if they incorrectly simulate a strong 'ENSO-like'
971 pattern in their long-term response to CO₂. However, a lack of eastern Pacific
972 warming in the long-term seems unlikely according to paleoclimate records (Tierney
973 et al. 2019; 2020).
- 974 5. Teleconnections from either the Atlantic Ocean (McGregor et al. 2018) or Southern
975 Ocean (Hwang et al. 2017) have potentially driven the tropical Pacific SST patterns.
976 Under the scenario of an Atlantic influence, we might expect the pattern effect to
977 reduce as Atlantic SST trends evolve over the next few decades. Under the scenario
978 of a Southern Ocean influence, we might expect the pattern effect to reduce as the
979 Southern Ocean surface warms; this could take years to decades if the Southern
980 Ocean temperature trends have been largely mediated by internal variability (e.g.,
981 Zhang et al. 2019) but could take centuries or longer if Southern Ocean cooling
982 continues due, for instance, to freshwater input from ongoing Antarctic ice shelf
983 melt (e.g., Sadai et al. 2020).

984 These are merely some of the proposed hypotheses, and not meant to be an exhaustive list. But
985 whatever the reason, the fact that AOGCMs rarely simulate this pattern (e.g. Watanbe et al., 2021;
986 Fueglistaler and Silvers, 2021; Dong et al., 2021) is a concern, suggesting either that their unforced
987 decadal variability is deficient, or that their forced response is biased, and in either case there is a
988 serious systematic error which affects all AOGCMs. Moreover, each of the above interpretations
989 imply different futures, and therefore untangling them is critical for informing both near-term and
990 long-term climate projections. This is time critical because satellite evidence suggests the Pacific SST
991 pattern that has dominated recent decades is currently shifting (Loeb et al., 2020) and indeed the
992 Earth's energy balance is rapidly changing with it (Loeb et al. 2021; Raghuraman et al., 2021).
993 Predicting the near future therefore depends on maintaining the continuity of the satellite record
994 and untangling the above mechanisms.

995

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1015

1016 **Data Availability**

1017 Global-annual-ensemble-mean dT and dN from all *amip-piForcing*, *hadSST-piForcing* and *abrupt-*
1018 *4xCO2* simulations are provided here <https://github.com/timothyandrews/amip-hadSST-piForcing>.
1019 Raw data from CMIP6 *amip-piForcing* simulations (indicated in Table 1) are available at
1020 <https://pcmdi.llnl.gov/CMIP6/>. *abrupt-4xCO2* raw data for most models is available at CMIP5
1021 (<https://esgf-node.llnl.gov/projects/cmip5/>) or CMIP6 (<https://pcmdi.llnl.gov/CMIP6/>). The
1022 HadCRUT5 analysis dataset is available at <https://www.metoffice.gov.uk/hadobs/hadcrut5/>. IPCC
1023 AR6 ERF timeseries is available at <https://github.com/IPCC-WG1/Chapter-7> (see
1024 https://github.com/IPCC-WG1/Chapter-7/blob/main/data_output/AR6_ERF_1750-2019.csv). DEEP-C
1025 v5 dN radiative fluxes can be obtained from <https://researchdata.reading.ac.uk/347/> and previous
1026 versions described at <http://www.met.reading.ac.uk/~sgs02rpa/research/DEEP-C/GRL/>. The
1027 HadISST1 SSTs used to force the *hadSST-piForcing* simulations are available at
1028 <https://www.metoffice.gov.uk/hadobs/hadisst/>, see
1029 https://www.metoffice.gov.uk/hadobs/hadisst/data/HadISST_sst.nc.gz.

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1352 **Table1: Summary of the Atmospheric General Circulation Model (AGM) simulations used in this study.** *amip-piForcing* refers to an AGCM simulation
 1353 forced with time-varying observed monthly SSTs and sea-ice using the AMIP II boundary condition SST and sea-ice dataset, forcing agents such greenhouse
 1354 gases, aerosol emission etc. are kept at pre-industrial levels. *hadSST-piForcing* is identical in all aspects except SSTs are taken from the HadISST1 database
 1355 (sea-ice remains the same as *amip-piForcing*). The ensemble size and time-periods covered for each experiment and AGCM is indicated. *amip-piForcing*
 1356 simulations included in the CFMIP3 (Webb et al. 2017) contribution to CMIP6 are indicated by a y/n. The corresponding name of each AGCMs parent
 1357 AOGCM is indicated. Global-annual-ensemble-mean dT and dN timeseries data are available for all *amip-piForcing* and *hadSST-piForcing* AGCM simulations
 1358 (see Data Availability Statement).

AGCM	Corresponding AOGCM name	Model description	<i>amip-piForcing</i>			<i>hadSST-piForcing</i>	
			CMIP6? (y/n)	Ensemble size	Time-period covered	Ensemble size	Time-period covered
CAM4	CCSM4	Neale et al. (2013)	n	3	1870 – 2014	3	1870 – 2014
CESM2	unchanged	Danabasoglu et al. (2020)	y	1	1870 – 2014	-	-
CNRM-CM6-1	unchanged	Voltaire et al. (2019)	y	1	1870 – 2014	-	-
CanESM5	unchanged	Swart et al. (2019)	y	3	1870 – 2014	-	-
ECHAM6.3	MPI-ESM1.1	Mauritsen et al. (2019)	n	5	1871 – 2010	5	1871 – 2015
GFDL-AM3	GFDL-CM3	Donner et al. (2011)	n	1	1870 – 2014	1	1870 – 2014
GFDL-AM4	GFDL-CM4	Held et al. (2019)	n	1	1870 – 2016	1	1870 – 2016
HadAM3	HadCM3	Pope et al. (2000)	n	4	1871 – 2012	4	1871 – 2012
HadGEM2	HadGEM2-ES	Martin et al. (2011)	n	4	1871 – 2012	1	1871 – 2012
HadGEM3-GC31-LL	unchanged	Williams et al. (2017)	y	1	1870 – 2014	1	1871 – 2016
IPSL-CM6A-LR	unchanged	Boucher et al. (2020)	y	1	1870 – 2014	-	-
MIROC6	unchanged	Tatebe et al. (2019)	y	1	1870 – 2014	-	-
MRI-ESM2-0	unchanged	Yukimoto et al. (2019), Kawai et al. (2019)	y	1	1870 – 2014	-	-
MPI-ESM1-2-LR	unchanged	Mauritsen et al. (2019)	n	3	1871 – 2017	3	1871 – 2017

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1361 **Table 2: Feedback parameter in *amip-piForcing* and *hadSST-piForcing* simulations over various historical time-periods, as well as *abrupt-4xCO2***
1362 **sensitivity parameters.** λ values from *amip-piForcing* and *hadSST-piForcing* are calculated from OLS regression ($\lambda = dN/dT$) over the relevant time-periods
1363 using global-annual-mean timeseries data. F_{2xCO_2} is calculated as $F_{4xCO_2}/2$ and $ECS = -F_{2x}/\lambda_{4xCO_2}$ from 150 years of *abrupt-4xCO2* experiments (λ_{4xCO_2} calculated
1364 over years 1-20 and 21-150 is also shown) (see Andrews et al., 2012; 2015).

	abrupt-4xCO2					$\lambda_{1871-2010}$ ($W m^{-2} K^{-1}$)		$\lambda_{1871-1980}$ ($W m^{-2} K^{-1}$)		$\lambda_{1981-2010}$ ($W m^{-2} K^{-1}$)	
	ECS (K)	F_{2x} ($W m^{-2}$)	λ_{4xCO_2} ($W m^{-2} K^{-1}$)	$\lambda_{4xCO_2, 1-20}$ ($W m^{-2} K^{-1}$)	$\lambda_{4xCO_2, 21-150}$ ($W m^{-2} K^{-1}$)	AMIP	HadISST1	AMIP	HadISST1	AMIP	HadISST1
CAM4	2.95	3.64	-1.23	-1.52	-0.94	-2.14	-1.77	-1.22	-1.45	-2.84	-2.70
CESM2	5.16	3.39	-0.66	-1.17	-0.49	-1.93	-	-0.87	-	-3.08	-
CNRM-CM6-1	4.88	3.66	-0.75	-0.93	-0.87	-1.23	-	-1.10	-	-1.64	-
CanESM5	5.61	3.64	-0.65	-0.70	-0.59	-1.44	-	-0.93	-	-1.83	-
ECHAM6_3	3.01	4.10	-1.36	-1.47	-1.08	-1.92	-1.57	-1.43	-1.38	-2.69	-2.42
GFDL-AM3	3.99	2.97	-0.74	-1.13	-0.61	-1.44	-1.35	-0.72	-0.99	-1.90	-1.41
GFDL-AM4	3.84	3.32	-0.86	-1.54	-0.60	-1.84	-1.66	-1.33	-1.40	-2.57	-2.93
HadAM3	3.37	3.52	-1.04	-1.25	-0.75	-1.65	-1.44	-1.35	-1.40	-2.19	-1.86
HadGEM2	4.62	2.90	-0.63	-0.81	-0.33	-1.39	-1.04	-1.12	-1.08	-2.26	-1.54
HadGEM3-GC31-LL	5.54	3.49	-0.63	-0.81	-0.60	-1.28	-1.01	-0.95	-0.84	-1.87	-1.55
IPSL-CM6A-LR	4.56	3.41	-0.75	-0.98	-0.61	-1.59	-	-1.17	-	-2.50	-
MIROC6	2.58	3.72	-1.44	-1.61	-1.60	-1.42	-	-1.21	-	-1.87	-
MRI-ESM2-0	3.13	3.44	-1.10	-1.68	-0.78	-1.93	-	-1.23	-	-2.79	-
MPI-ESM1-2-LR	3.02	4.21	-1.39	-1.61	-1.34	-1.88	-1.58	-1.30	-1.45	-2.55	-2.42
MEAN	4.02	3.53	-0.95	-1.23	-0.80	-1.65	-1.43	-1.14	-1.25	-2.33	-2.10
1.645 σ	1.64	0.57	0.49	0.54	0.55	0.46	0.43	0.33	0.37	0.72	0.90

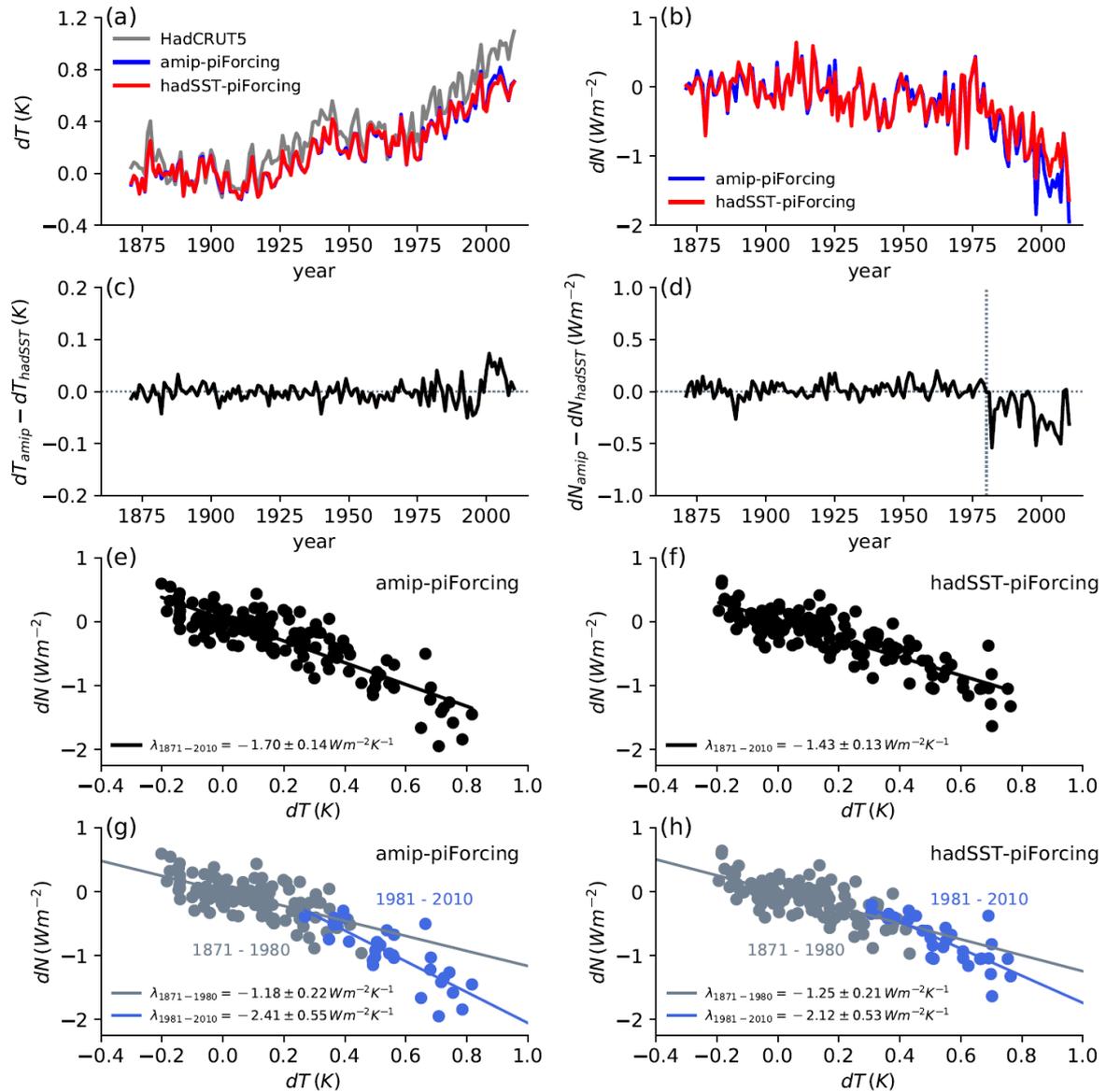
1365 **Table 3: The pattern effect ($\Delta\lambda = \lambda_{4xCO_2} - \lambda_{hist}$, with λ_{4xCO_2} from years 1-150 of *abrupt-4xCO2*)**
1366 **between *abrupt-4xCO2* radiative feedback and radiative feedback calculated over different**
1367 **historical periods (i.e. λ_{hist} from 1871-2010, and its separation into 1871-1980 and 1981-2010) in**
1368 ***amip-piForcing* and *hadSST-piForcing*, as well as their difference.**

	1871 – 2010 (W m ⁻² K ⁻¹)			1871 – 1980 (W m ⁻² K ⁻¹)			1981 – 2010 (W m ⁻² K ⁻¹)		
	AMIP	HadSST	Diff	AMIP	HadSST	Diff	AMIP	HadSST	Diff
CAM4	0.90	0.53	0.37	-0.01	0.22	-0.23	1.60	1.47	0.13
CESM2	1.27			0.21			2.43		
CNRM-CM6-1	0.48			0.35			0.89		
CanESM5	0.80			0.28			1.19		
ECHAM6_3	0.56	0.21	0.35	0.07	0.02	0.05	1.32	1.06	0.26
GFDL-AM3	0.69	0.61	0.08	-0.03	0.24	-0.27	1.15	0.67	0.48
GFDL-AM4	0.97	0.80	0.17	0.47	0.53	-0.06	1.70	2.07	-0.37
HadAM3	0.61	0.40	0.21	0.31	0.35	-0.04	1.15	0.82	0.33
HadGEM2	0.76	0.41	0.35	0.49	0.45	0.04	1.63	0.91	0.72
HadGEM3-GC31-LL	0.65	0.38	0.27	0.32	0.21	0.11	1.24	0.92	0.32
IPSL-CM6A-LR	0.84			0.43			1.76		
MIROC6	-0.02			-0.23			0.42		
MRI-ESM2-0	0.83			0.14			1.69		
MPI-ESM1-2-LR	0.49	0.19	0.30	-0.09	0.06	-0.15	1.16	1.03	0.13
MEAN	0.70	0.44	0.26	0.19	0.26	-0.07	1.38	1.12	0.26
1.645σ	0.47	0.31	0.16	0.35	0.28	0.07	0.75	0.69	0.06

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1370 **Table 4: Comparison of the 1985-2014 climate resistance ($\rho = dF/dT$), feedback parameter ($-\lambda = -$
1371 $d(N - F)/dT$ and ocean heat uptake efficiency ($\kappa = dN/dT$) using different versions of the DEEP-C
1372 (Allan et al., 2014) satellite based reconstruction of dN (see Section 2.4). The lower half of the
1373 table shows how ρ , λ and κ estimates change as the 30 year moving window advances to 1990-
1374 2019. In all calculations HadCRUT5 analysis dT (Morice et al. 2021) and IPCC AR6 dF (Forster et al.,
1375 2021) are used. Years 1991-2 are excluded from the calculation as these years are identified as
1376 being strongly impacted by the volcanic forcing from the Pinatubo eruption (Section 4).**

dN dataset version	Start year	End year	ρ ($W m^{-2} K^{-1}$)	$-\lambda$ ($W m^{-2} K^{-1}$)	κ ($W m^{-2} K^{-1}$)
DEEP-C v2G			2.38	2.24	0.14
DEEP-C v3			2.38	2.24	0.14
DEEP-C v3G	1985	2014	2.38	2.24	0.14
DEEP-C v4			2.38	1.98	0.41
DEEP-C v5			2.38	1.98	0.41
DEEP-C v5	1986	2015	2.38	1.75	0.63
DEEP-C v5	1987	2016	2.25	1.55	0.70
DEEP-C v5	1988	2017	2.21	1.62	0.59
DEEP-C v5	1989	2018	2.23	1.66	0.57
DEEP-C v5	1990	2019	2.30	1.44	0.86



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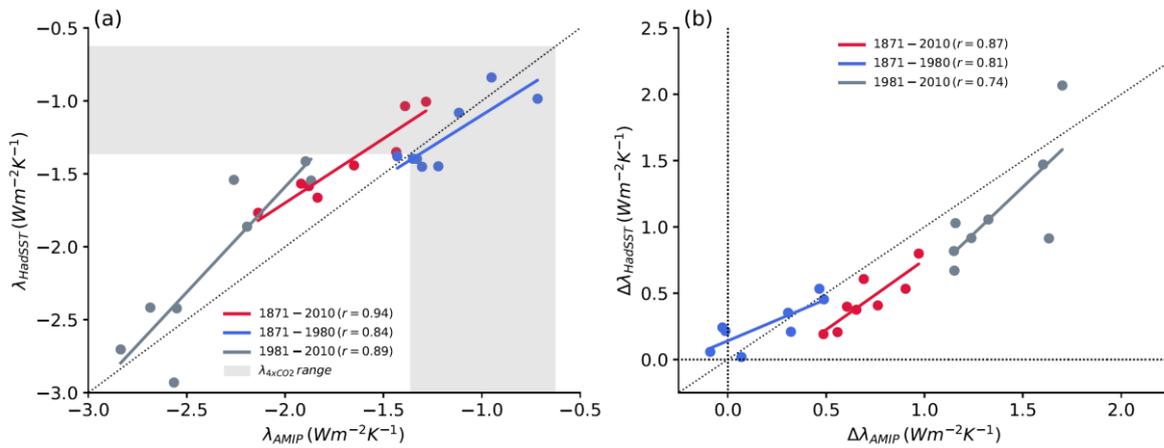
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Figure 1: Comparison of multi-model ensemble-annual-mean (a) dT and (b) dN in the amip-piForcing and hadSST-piForcing simulations. (c) and (d) shows the difference in dT and dN respectively, highlighting 1980 as a key year where the dN response diverges according to the SST dataset. In (a) the HadCRUT5 observed dT evolution is shown for comparison. (e) and (f) show the relationship between global-annual-mean dT and dN in amip-piForcing and hadSST-piForcing respectively, where $\lambda=dN/dT$ is calculated from OLS regression on the global-annual-mean data points. The stated 5-95% uncertainty is $\pm 1.645\sigma$ from the standard error of the linear fit. (g) and (h) show the dT and dN relationship separated into two time-periods: years 1871-1980 (grey) and years 1981-2010 (blue).

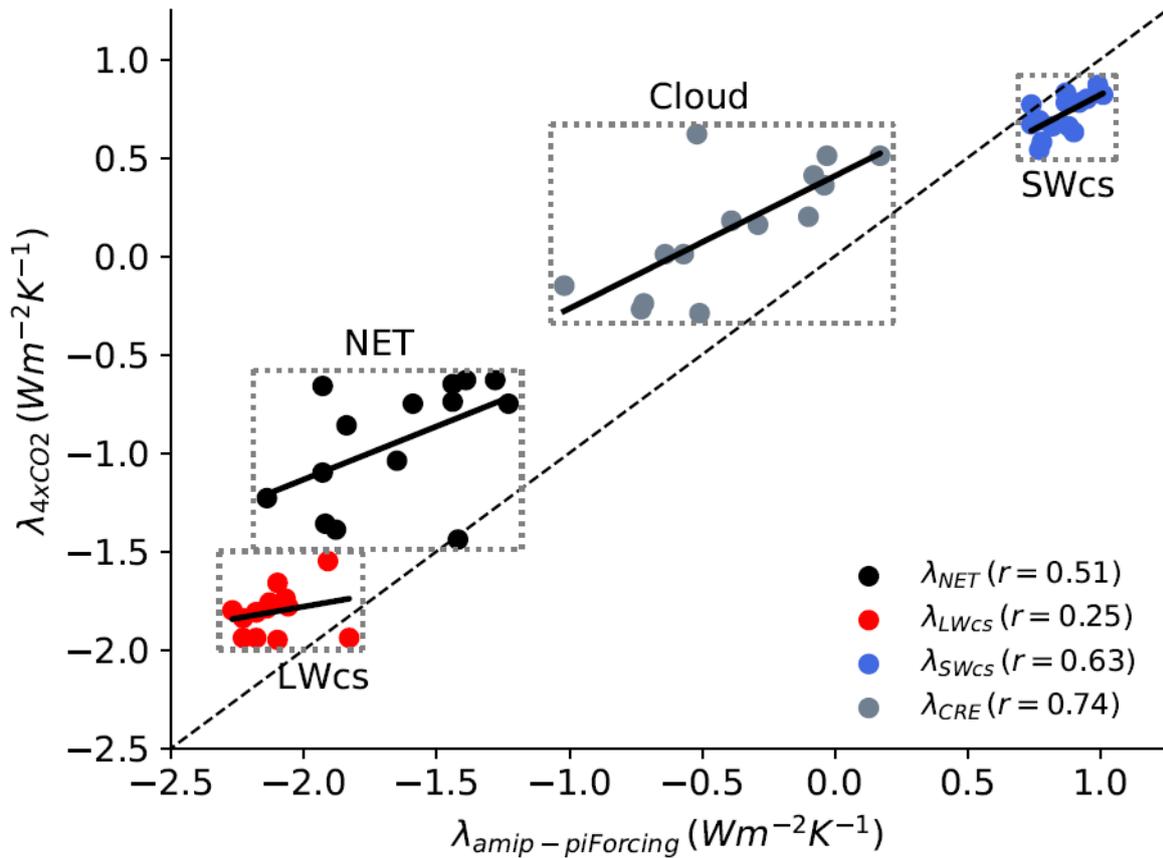
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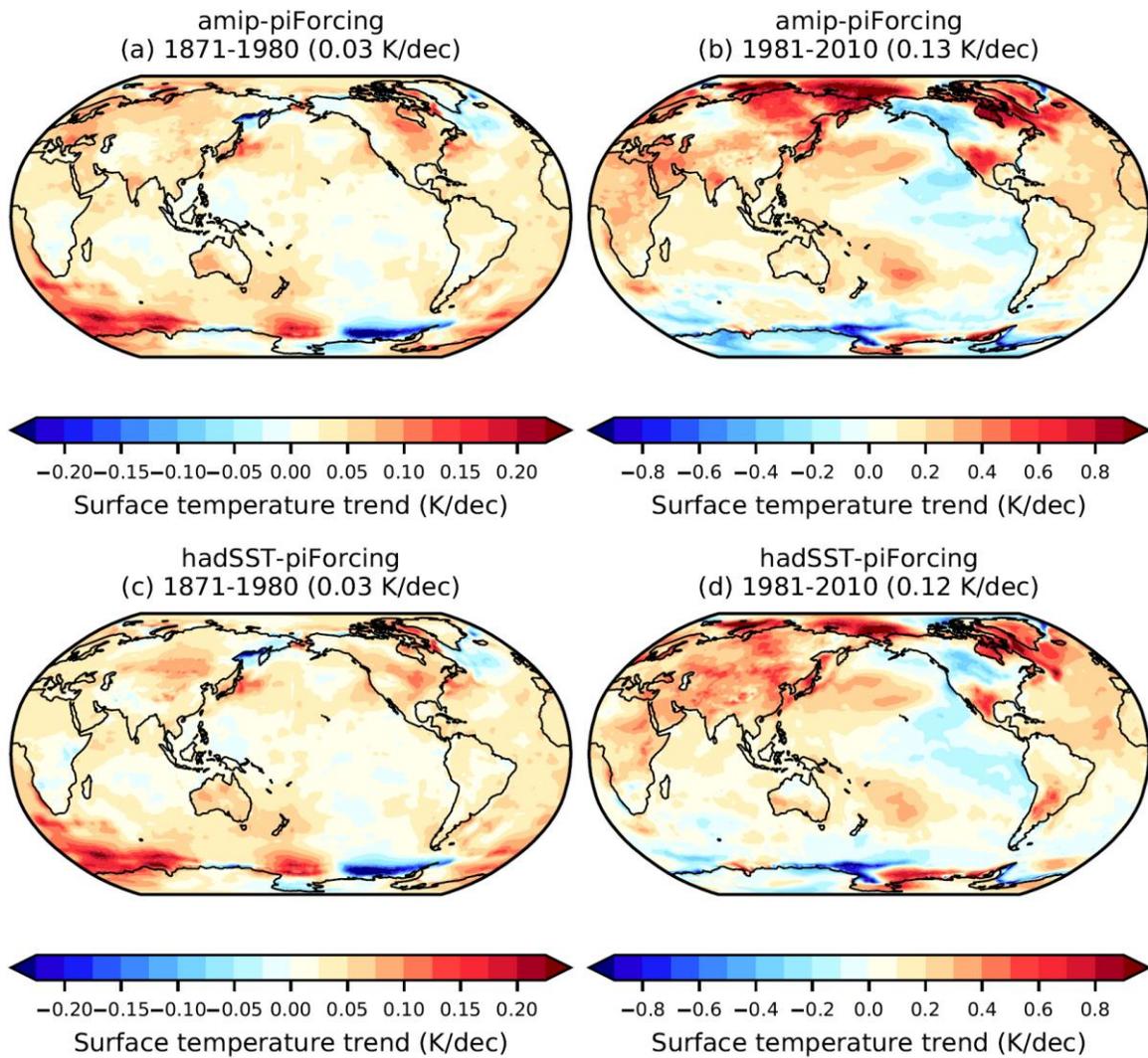
1389 **Figure 2: (a) Relationship between the feedback parameter, λ , in the *amip-piForcing* and *hadSST-***
1390 ***piForcing* simulations over various historical time-periods. Each point is a single AGCM. The**
1391 **shaded grey region shows the range of λ_{4xCO2} from the AGCMs corresponding parent AOGCM**
1392 ***abrupt-4xCO2* simulation. The one-to-one line (dotted) is shown. (b) Relationship between the**
1393 **pattern effect, $\Delta\lambda = \lambda_{4xCO2} - \lambda_{hist}$, diagnosed from the *amip-piForcing* and *hadSST-piForcing***
1394 **simulations over various historical time-periods.**

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1397 **Figure 3: Relationship across models (dots) between the feedback parameter in *amip-piForcing***
 1398 **(calculated over years 1871-2010) and *abrupt-4xCO2* simulation (calculated over years 1-150). The**
 1399 **net feedback parameter is decomposed into its longwave clear-sky, SW clear-sky and cloud**
 1400 **radiative effect components.**



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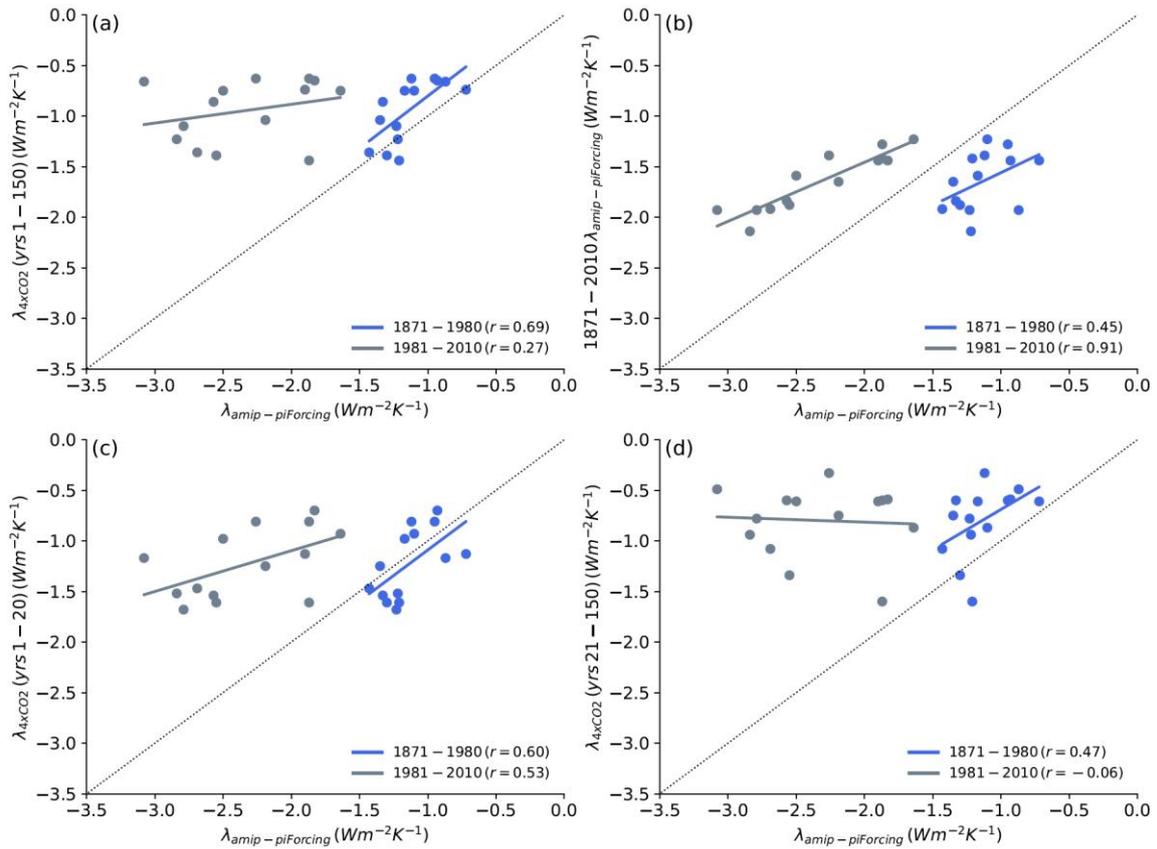
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Figure 4: Decadal surface temperature trends over 1870-1980 and 1981-2010 in (a) and (b) *amip-piForcing* and (c) and (d) *hadSST-piForcing*. Trends are calculated from the linear regression of dT against time over the corresponding time-periods, on annual-mean data. Data from HadGEM3-GC31-LL simulations have been used for this illustration.

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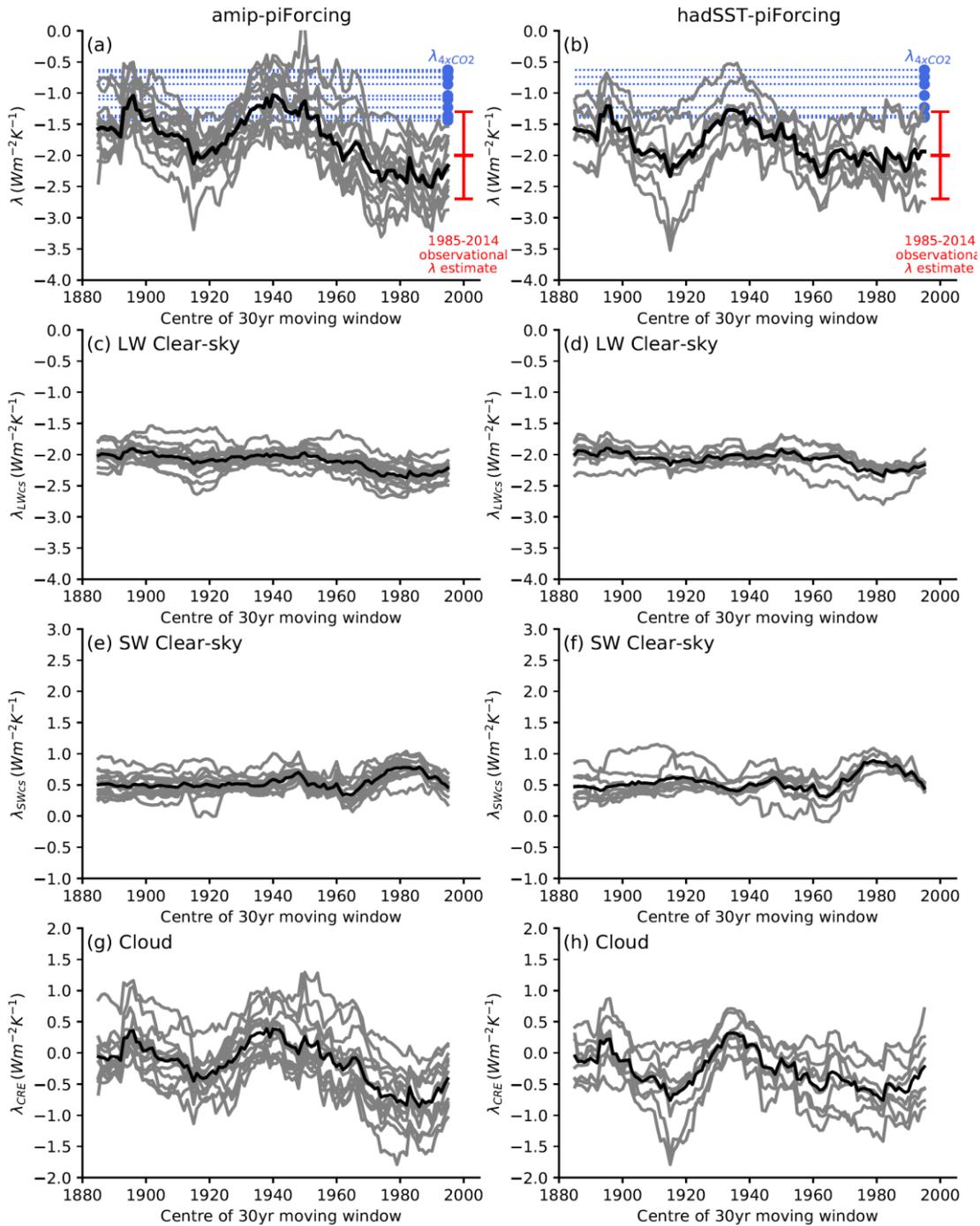
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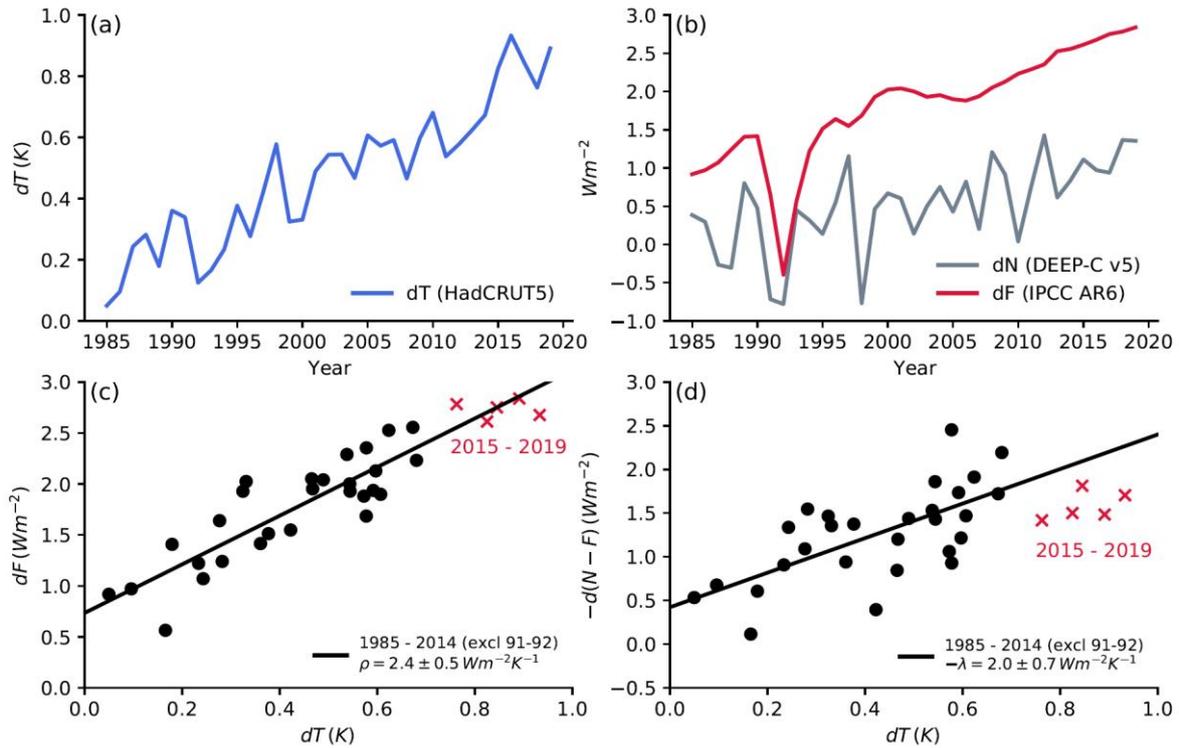
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Figure 5: Relationships between model simulated feedbacks in *amip-piForcing* over years 1871-1980 (blue) or 1981-2010 (grey) and (a) λ_{4xCO2} from *abrupt-4xCO2*, (b) λ_{hist} over the entire historical record (1871-2010), (c) λ_{4xCO2} from *abrupt-4xCO2* over years 1-20 and (d) years 21-150.



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1413 **Figure 6: Decadal variation in the feedback parameter λ from 1871 to 2010. Left column shows**
 1414 **results from *amip-piForcing* and right column shows results from *hadSST-piForcing*. Each grey line**
 1415 **represents a single AGCM (see Table 1). Thick black is the ensemble-mean of the results. X-axis**
 1416 **represents the centre of a 30 year moving window in which $\lambda=dN/dT$ is calculated from OLS**
 1417 **regression on annual-mean data, i.e. λ at 1980.5 represents the feedback parameter over years**
 1418 **1966 to 1995. Shown in (a) and (b) is the net feedback parameter. Blue dots and lines represent**
 1419 **the corresponding λ_{4xCO2} values from AOGCM *abrupt-4xCO2* simulations (Table 2). Red shows an**
 1420 **observational estimate and 5-95% uncertainty of $\lambda=d(N - F)/dT \sim -2.0 \pm 0.7 \text{ W m}^{-2} \text{ K}^{-1}$ over years**
 1421 **1985-2014 (see Section 4). (c) – (h) shows the corresponding LW clear-sky, SW clear-sky and cloud**
 1422 **radiative effect (CRE) components of λ .**

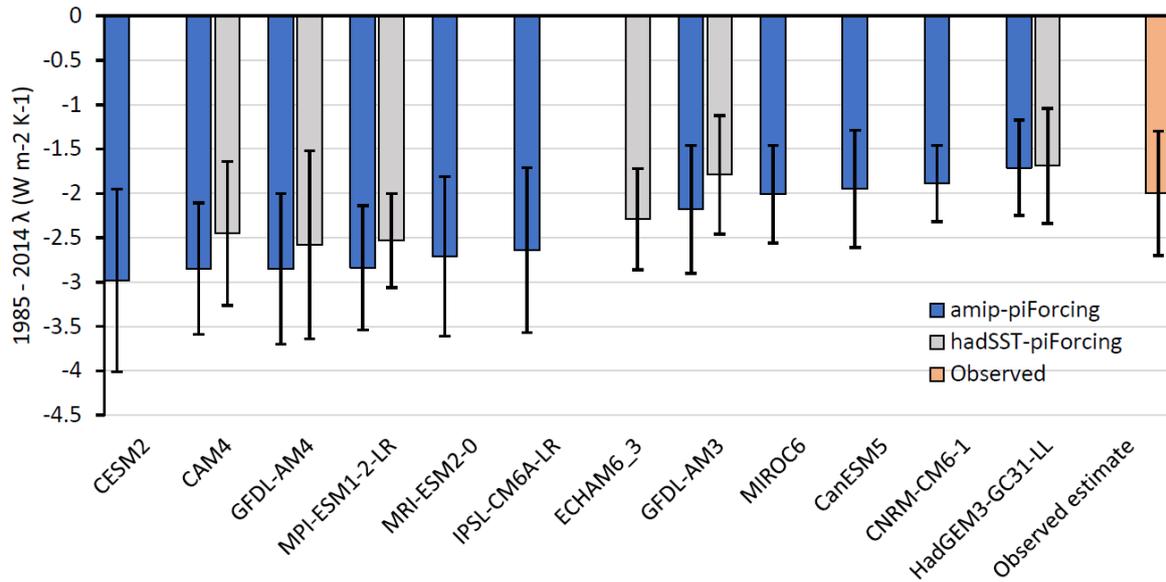


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1424 **Figure 7: Observational estimate of the Earth's 1985-2019 energy balance. All points are global-**
 1425 **annual-means. (a) dT (HadCRUT5 analysis dataset; Morice et al., 2021), (b) dN (DEEP-C v5; Allan et**
 1426 **al., 2014; Liu and Allan, 2022) and dF (IPCC AR6; Forster et al., 2021). (c) $\rho = dF/dT$ relationship and**
 1427 **(d) $-\lambda_{\text{hist}} = -d(N - F)/dT$ relationship over years 1985-2014. Black dots are global-annual means over**
 1428 **years 1985-2014 excluding years 1991-2 which are strongly influenced by the Pinatubo explosive**
 1429 **volcanic eruption (see red line panel b). Red points in (c) and (d) are years 2015-2019. The stated**
 1430 **5-95% uncertainties are $\pm 1.645\sigma$ from the standard error of the linear fit.**

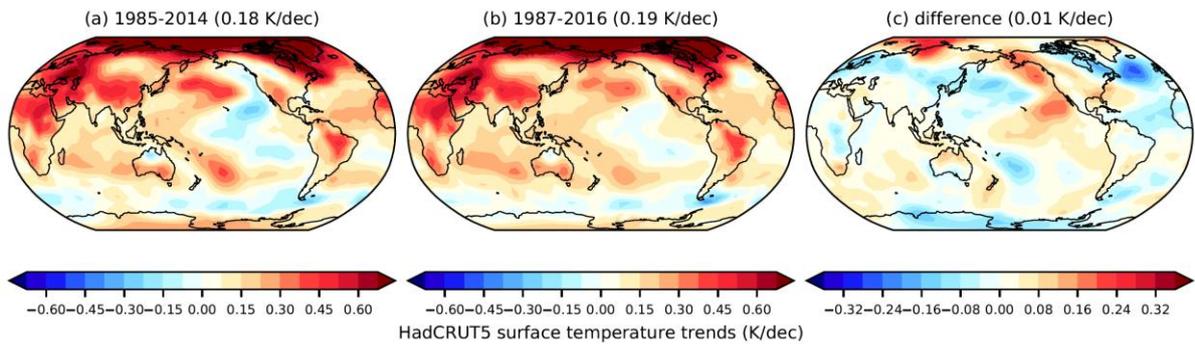
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1434 **Figure 8: Comparison of the 1985-2014 feedback parameter, $\lambda_{\text{hist}} = d(N - F)/dT$, in *amip-piForcing***
 1435 **and *hadSST-piForcing* simulations to an observed estimate based on DEEP-C V5 dN (Allan et al.,**
 1436 **2014; Liu and Allan, 2022), HadCRUT5 analysis dT (Morice et al. 2021) and IPCC AR6 dF (Forster et**
 1437 **al., 2021). The 5-95% uncertainty is simply 1.645σ from the standard error of the linear fit, with no**
 1438 **allowance for systematic uncertainties. Note also that years 1991-2 are excluded from the**
 1439 **calculation as these years are identified as being strongly impacted by the volcanic forcing from**
 1440 **the Pinatubo eruption (Figure 7b).**



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1442 **Figure 9: Decadal trend in near-surface temperature change over (a) 1985-2014 and (b) 1987-2016,**
 1443 **(c) shows the difference (b minus a). Data is the HadCRUT5 analysis dataset (Morice et al. 2021).**

1444 **Trends are calculated from linear regression on annual-mean data points at each grid box.**