

This manuscript is a preprint and has not been reviewed.

It has been submitted to *Journal of Climate*.

Copyright in this work may be transferred without further notice.

1

2 **Decomposing the Drivers of Polar Amplification with a Single Column**

3 **Model**

4 Matthew Henry*

5 *College of Engineering, Mathematics and Physical Sciences, University of Exeter, Exeter, UK*

6 Timothy M. Merlis

7 *Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec,*

8 *Canada*

9 Nicholas J. Lutsko

10 *Scripps Institution of Oceanography, University of California at San Diego, La Jolla, CA, USA*

11 Brian E.J. Rose

12 *Department of Atmospheric and Environmental Sciences, University at Albany (State University*

13 *of New York), USA*

14 *Corresponding author address: Matthew Henry, College of Engineering, Mathematics and Phys-
15 ical Sciences, Harrison Building, Streatham Campus, University of Exeter, North Park Road, Ex-
16 eter, UK, EX4 4QF.

17 E-mail: m.henry@exeter.ac.uk

ABSTRACT

18 The precise mechanisms driving Arctic amplification are still under debate.
19 Previous attribution methods based on top-of-atmosphere energy budgets have
20 assumed all forcings and feedbacks lead to vertically-uniform temperature
21 changes, with any departures from this collected into the lapse-rate feedback.
22 We propose an alternative attribution method using a single column model
23 that accounts for the forcing-dependence of high latitude lapse-rate changes.
24 We test this method in an idealized General Circulation Model (GCM), find-
25 ing that, even though the column-integrated carbon dioxide (CO₂) forcing
26 and water vapor feedback are stronger in the tropics, they contribute to polar-
27 amplified surface warming as they lead to bottom-heavy warming in high lati-
28 tudes. A separation of atmospheric temperature changes into local and remote
29 contributors shows that, in the absence of polar surface forcing (e.g., sea-ice
30 retreat), changes in energy transport are primarily responsible for the polar
31 amplified pattern of warming. The addition of surface forcing substantially
32 increases polar surface warming and reduces the contribution of atmospheric
33 dry static energy transport. This physically-based attribution method can be
34 applied to comprehensive GCMs to provide a clearer view of the mechanisms
35 behind Arctic amplification.

36 **1. Introduction**

37 The Arctic amplification of surface temperature change is a robust feature of observations
38 (Stocker et al. 2013) and comprehensive climate model simulations (Pithan and Mauritsen 2014).
39 A number of mechanisms are thought to contribute to Arctic amplification, including the surface
40 albedo feedback, increased atmospheric energy transport convergence (Hwang and Frierson 2010),
41 and the temperature feedback (Pithan and Mauritsen 2014); however, the precise contribution of
42 each mechanism is still unclear. Clarifying how these different factors contribute to Arctic ampli-
43 fication is essential for reducing the uncertainty in the rate of Arctic warming through improved
44 process-level understanding.

45 The tropics differ from the high latitudes in that they are close to radiative-convective equilib-
46 rium: heating by convection is balanced by radiative cooling, and the vertical temperature profile
47 is mostly determined by surface temperature and humidity. The high latitudes, on the other hand,
48 are close to radiative-advective equilibrium: warming from horizontal atmospheric heat trans-
49 port is balanced by cooling from radiation. This means that different forcings and feedbacks
50 induce different lapse rate responses. For example, an increase in longwave optical depth leads to
51 bottom-heavy warming (Cronin and Jansen 2016; Henry and Merlis 2019), whereas atmospheric
52 energy transport is thought to primarily affect the midtroposphere at high latitudes (Laliberté and
53 Kushner 2013; Feldl et al. 2017a). This implies that the ratio between surface warming and top-
54 of-atmosphere (TOA) net radiation changes at the high latitudes is different for each forcing and
55 feedback. Surface temperature change attributions based on TOA budget analyses (Pithan and
56 Mauritsen 2014) and moist energy balance models (Roe et al. 2015) assume a linear relationship
57 between surface temperature change and TOA net radiation change that is independent of forcing.
58 Hence these attribution methods neglect the dependence of the vertical structure of warming on

59 the perturbation type at high latitudes, and may produce misleading attributions of the drivers of
60 Arctic amplification.

61 Process oriented and mechanism denial experiments are useful tools for studying the mecha-
62 nisms responsible for Arctic amplification. For example, the analysis from Stuecker et al. (2018)
63 suggests that local forcings and feedbacks dominate the polar-amplified pattern of surface temper-
64 ature change in a comprehensive GCM in which CO₂ concentrations are increased in restricted
65 latitudinal bands. Using the conventional TOA budget-based attribution method, they suggest that
66 the lapse rate feedback is a main contributor to this surface temperature change pattern. Screen
67 et al. (2012) attribute near-surface warming to local forcings and feedbacks and warming aloft to
68 atmospheric energy transport increases by prescribing local and remote sea surface temperature
69 (SST) and sea ice concentration (SIC) changes in two comprehensive atmospheric GCMs. How-
70 ever, fixing SST where the model would otherwise warm (or cool) the surface is akin to imposing
71 a surface heat sink (or source), hence the results are not easily interpretable.

72 While these comprehensive GCM studies provide important insights into the mechanisms of
73 Arctic amplification, a hierarchy of models is required for a complete understanding of the drivers
74 of Arctic amplification in climate models and observations. Previous work using single column
75 model representations of the high latitude atmosphere suggested that the high latitude temperature
76 response is sensitive to the forcing type (Abbot and Tziperman 2008; Payne et al. 2015). Cronin
77 and Jansen (2016) have developed a 1-dimensional model of an atmosphere in radiative-advective
78 equilibrium for the high latitudes, which led to the important insight that high latitude lapse rate
79 changes are forcing-dependent. The present work seeks to bridge the gap between their simple
80 radiative-advective column model and complex climate model simulations in order to advance our
81 understanding of the drivers of Arctic amplification.

82 Using an idealized moist atmospheric GCM with aquaplanet surface boundary conditions, no
83 clouds, and no sea ice (hence no surface albedo feedback), we qualitatively reproduce the pattern
84 of surface temperature change from comprehensive GCMs. To simulate the effect of melting sea
85 ice, we impose a polar surface heat source, ranging from 0 to 24 W m^{-2} . Then, we use a single
86 column model (SCM) to emulate the tropics and high latitudes of the idealized GCM. This allows
87 us to calculate the response to each individual forcing and feedback and thus decompose the drivers
88 of tropical and polar temperature change. By accounting for each forcing and feedback's impact
89 on the vertical structure of temperature change, this physically-based attribution method does not
90 assume a universal high latitude lapse rate feedback, and, therefore, does not ignore how the
91 vertical structure of temperature change depends on the perturbation. The idealized GCM acts as a
92 test-case for the attribution method, which could potentially be used to untangle the contributions
93 of the various mechanisms of polar amplification in comprehensive models or in observations.

94 **2. Idealized atmospheric GCM**

95 We use an idealized moist atmospheric GCM based on the Geophysical Fluid Dynamics Lab-
96 oratory (GFDL) spectral dynamical core and the comprehensive radiation scheme of the GFDL
97 AM2 GCM, with no sea ice or clouds. This is similar to the setup in Merlis et al. (2013) and to
98 the Model of an Idealized Moist Atmosphere (MiMA, Jucker and Gerber (2017)). These GCMs
99 follow the moist idealized GCM described in Frierson et al. (2006), but use comprehensive clear-
100 sky radiation instead of grey radiation. In the MiMA setup, the surface albedo is globally uniform
101 and increased to compensate for the cooling effect of clouds. In Merlis et al. (2013), an idealized
102 cloud distribution is prescribed for the radiative transfer calculation. Here, there are no clouds and
103 we set the surface albedo to a hemispherically symmetric analytic distribution similar to Earth's
104 northern hemisphere TOA albedo, as estimated from the Cloud and the Earth's Radiant Energy

105 System data (Loeb et al. (2018), see supplemental figure S1), in order to produce an Earth-like
106 meridional surface temperature gradient. The model uses the comprehensive radiation scheme de-
107 scribed in Anderson et al. (2004) with annual mean solar insolation and a solar constant equal to
108 1365 W m^{-2} .

109 The surface boundary condition is a slab mixed layer ocean aquaplanet with no representation
110 of ocean heat transport and the heat capacity of 1m of water. We use annual-mean insolation and
111 the small mixed layer depth allows the model to run quickly without meaningfully affecting the
112 model's climate, as we only consider annual-mean quantities. The GCM was run at T42 spectral
113 truncation, for a nominal horizontal resolution of $2.8^\circ \times 2.8^\circ$, and with 30 vertical levels. The skin
114 temperature is interactively computed using the surface radiative and turbulent fluxes, which are
115 determined by bulk aerodynamic formulae. A k-profile scheme with a dynamically determined
116 boundary layer height is used to parameterize the boundary layer turbulence. The GCM uses a
117 simplified Betts-Miller convection scheme (Frierson 2007), and large scale condensation is pa-
118 rameterized such that the relative humidity does not exceed one and condensed water is assumed
119 to immediately return to the surface. As there is no representation of sea ice, there is no surface
120 albedo feedback. To mimic the presence of the surface albedo feedback, we run perturbation ex-
121 periments with an added polar surface heat source. All simulations are run for 20 years with time
122 averages over the last 10 years shown, when all climate states have reached a statistical steady
123 state.

124 We perform four simulations: a control run in which the atmospheric CO_2 concentration is set
125 to 300 ppm, a run with quadrupled (1200 ppm) CO_2 concentration, and two runs with quadrupled
126 CO_2 concentrations and constant surface heat sources Q_s of 12 W m^{-2} and 24 W m^{-2} poleward
127 of 80° in both hemispheres. The heat sources simulate surface heating through the surface albedo
128 feedback or increased oceanic energy transport convergence. Given that the polar surface tempera-

129 ture change under $4xCO_2$ is approximately 8K, a 12 (24) $W m^{-2}$ surface heat source is equivalent
130 to a 1.5 (3) $W m^{-2} K^{-1}$ local feedback. This can be compared to the locally defined surface
131 albedo feedback from the models participating in the fifth coupled model intercomparison project
132 (CMIP5) which is approximately $1 W m^{-2} K^{-1}$ in the Arctic and $2 W m^{-2} K^{-1}$ in the Antarctic
133 (Feldl and Bordoni 2016, their figure 1).

134 Figure 1a shows the zonal-mean surface skin temperature differences between the control and
135 three perturbation simulations, in addition to the zonal-mean surface temperature responses of
136 abrupt $4xCO_2$ experiments with 7 models participating in the sixth Coupled Model Intercom-
137 parison Project (CMIP6) listed in the legend of figure 1 (Eyring et al. 2016). Figure 1b shows the
138 surface temperature changes normalized by their global mean. The patterns of surface temperature
139 change from the idealized model experiments (black) approximately span the CMIP6 model re-
140 sponses (colors). The amount of Arctic amplification is underestimated in the $4xCO_2$ experiment,
141 but adding a polar surface heat source brings the idealized GCM closer to CMIP6 in the Arctic,
142 with high latitude warming of 2 to 3 times the global-mean surface temperature change. Note that
143 the CMIP6 temperature changes are not fully equilibrated, and, at equilibrium, the Antarctic is
144 also expected to have amplified warming, but this warming is transiently delayed by ocean heat
145 uptake (Manabe et al. 1991; Rugenstein et al. 2019).

146 **3. Single column model**

147 To emulate the tropical and high-latitude atmosphere of the idealized GCM, we use the single
148 column model (SCM) from the ClimLab python package for process-oriented climate modeling
149 Rose (2018). The temperature tendency budgets for atmospheric and surface temperature are given
150 by the following equations:

$$\frac{\partial T_{atm}(p)}{\partial t} = \frac{\partial T_{atm}(p)}{\partial t} \Big|_{rad} + \frac{\partial T_{atm}(p)}{\partial t} \Big|_{conv} + \frac{\partial T_{atm}(p)}{\partial t} \Big|_{adv} + \frac{\partial T_{atm}(p)}{\partial t} \Big|_{cond} \quad (1)$$

$$\frac{\partial T_s}{\partial t} = \frac{\partial T_s}{\partial t} \Big|_{rad} + \frac{\partial T_s}{\partial t} \Big|_{SH} + \frac{\partial T_s}{\partial t} \Big|_{LH}, \quad (2)$$

151 where t is time and p is pressure (with 40 pressure levels). The subscripts ‘rad’, ‘conv’, ‘adv’, and
 152 ‘cond’, ‘SH’, ‘LH’ refer to radiative, convective, advective, condensation, sensible heat flux, and
 153 latent heat flux temperature tendencies, respectively. The radiative, convective, sensible heat flux,
 154 and latent heat flux temperature tendencies are computed interactively. The RRTMG radiation
 155 scheme is used for the computation of shortwave and longwave radiative temperature tendencies.
 156 The surface albedo and control insolation are set such that the upwelling and downwelling TOA
 157 shortwave radiation match the idealized GCM simulations in the tropics (10°S to 10°N) and pole-
 158 ward of 80°. Convection is implemented as an adjustment of the temperature profile to the moist
 159 adiabat, whereas the idealized GCM uses a simplified Betts-Miller convection scheme (Frierson
 160 2007). Note that at high latitudes, horizontal atmospheric energy transport induces a temperature
 161 structure stable to convection, hence convection has no effect. The surface sensible and latent heat
 162 fluxes are computed using bulk aerodynamic formulae with 5×10^{-2} drag coefficient and 5 ms^{-1}
 163 near surface wind speed (Rose 2018).

164 Values from the idealized GCM experiments averaged in the tropics (10°S to 10°N) and pole-
 165 ward of 80°N are used to prescribe the specific humidity profile, which affects the radiation and
 166 surface latent heat flux. In addition, the time-mean advection and condensation temperature ten-
 167 dency profiles from the idealized GCM simulations are added as external temperature tendency
 168 terms to simulate the dry and moist components of atmospheric energy transport convergence,
 169 respectively (see supplementary figure S2 for the temperature tendency profiles). The advective

170 temperature tendency term is calculated in the GCM as the difference in temperature tendency be-
171 fore and after running the dynamics module, hence it contains the horizontal and vertical advection
172 temperature tendencies.

173 The climatological temperature profiles of the idealized GCM and SCM are similar (figure 2),
174 though the SCM has an overly strong near-surface temperature inversion compared to the GCM.
175 This may be due to the absence of boundary layer scheme in the SCM, which would smooth dif-
176 ferences between the surface and lower atmospheric layers. Similarities between the temperature
177 profiles simulated by the idealized GCM and by the SCM still hold when the latitudinal bounds of
178 the tropics are set to 20°S-20°N and the high latitudes to 60° (see supplementary figure S3).

179 **4. Attribution of idealized GCM tropical and polar lapse rate changes to forcings and feed-** 180 **backs.**

181 As discussed in the introduction, the forcing dependence of the high latitude lapse rate feedback
182 makes a TOA budget approach to attributing the polar surface warming to different forcings and
183 feedbacks ambiguous (see next section). The SCM allows us to attribute the idealized GCM's
184 tropical and polar lapse rate changes to the different forcings and feedbacks. We individually
185 perturb CO₂, water vapor, and atmospheric energy transport (moist and dry components) in the
186 tropics and high latitudes to attribute the total warming to each of these individual components.

187 Figure 3 shows the decomposition of (a) tropical and (b,c,d) polar lapse rate changes of the
188 three idealized GCM perturbation experiments: 4xCO₂ (a,b), 4xCO₂ with $Q_s=12 \text{ W m}^{-2}$ (a,c) and
189 $Q_s=24 \text{ W m}^{-2}$ (a,d); Table 1 summarizes the surface temperature change attributions.

190 The tropical lapse rate changes for the three experiments are similar enough to be plotted in the
191 same figure 3a: the $Q_s = 12 \text{ W m}^{-2}$ and $Q_s = 24 \text{ W m}^{-2}$ experiment changes are shown in dashed
192 and dash-dotted respectively, and fall close to each other. They are decomposed into the tempera-

193 ture change from the CO₂ forcing (red), water vapor feedback (blue), and energy transport (green).
194 For each GCM experiment, the SCM's response to applying all of the perturbations simultaneously
195 (black) is exactly the same as the sum of the responses to the individual perturbations and fits the
196 idealized GCM's response well throughout the troposphere (grey), demonstrating the accuracy of
197 the attribution method. Differences in the stratosphere between the SCM and idealized GCM may
198 be due to the different radiation schemes or ozone distributions. Since convection is triggered in
199 the tropics, the temperature profiles are moist adiabatic and the vertical structure of tropospheric
200 temperature change ($\Delta T/\Delta T_S$) is approximately the same for all SCM experiments. The energy
201 transport is slightly reduced in the experiments with surface heat sources.

202 The polar lapse rate changes (b,c,d) are decomposed into the temperature changes from the CO₂
203 forcing (red), water vapor feedback (blue), the 'local' water vapor feedback (blue dashed, see
204 section 6), the energy transport (dry component in orange and moist component in cyan), and
205 surface heat source (yellow). Again, for each GCM experiment, the SCM's response to applying
206 all of the perturbations simultaneously (black) is exactly the same as the sum of the responses to the
207 individual perturbations, and fits the idealized GCM's response well throughout the troposphere
208 (grey), showing the accuracy of the attribution method. The increase in longwave absorbers (CO₂
209 and water vapor) leads to bottom-heavy warming, the dry component of energy transport leads to
210 top-heavy warming, the moist component of energy transport leads to mid-troposphere enhanced
211 warming, and the surface heat source leads to very bottom-heavy warming.

212 The polar surface temperature change is 3.6K and 7.6K higher in the $Q_s = 12\text{W m}^{-2}$ and $Q_s =$
213 24W m^{-2} cases respectively, which is caused mainly by 4.6K and 8.9K warming, respectively,
214 due to the surface heat source. The reduction in the dry component of energy transport causes
215 a 2.1K and 4.1K cooling respectively versus no warming in the simulation with $Q_s = 0\text{W m}^{-2}$.
216 There are also slight increases in warming due to the water vapor feedback (discussed in section

217 6) and moist component of the energy transport compared to the 4xCO₂ experiment (Table 1).
 218 This is consistent with Hwang et al. (2011), who found that enhanced Arctic warming due to
 219 local feedbacks weakens the equator-to-pole temperature gradient and reduces the dry component
 220 of the atmospheric energy transport, which outweighs the increase in the moist component of
 221 atmospheric energy transport that arises from the enhanced warming.

222 5. Surface temperature change attribution method comparison

223 The conventional surface temperature change attribution method (Pithan and Mauritsen 2014;
 224 Stuecker et al. 2018) assumes forcings and feedbacks lead to vertically uniform temperature
 225 changes. The deviation from vertically uniform temperature change is then accounted for in the
 226 lapse rate feedback. One can decompose the surface temperature change as follows (equation 3 in
 227 the Methods section of Stuecker et al. (2018)):

$$\Delta T_S(\phi) = \left(-\frac{1}{\lambda_P}\right) \{ \Delta T_S(\phi) [\lambda'_P(\phi) + \lambda_{LR}(\phi) + \lambda_{WV}(\phi) + \lambda_{AL}(\phi)] + \mathcal{F}(\phi) + \Delta(\nabla \cdot \vec{F}(\phi)) \} \quad (3)$$

228 where ϕ is the latitude. The surface temperature change attributions are then given by the average
 229 of $\Delta T_S(\phi)$ over the tropics and Arctic. The Planck feedback is decomposed into its global-mean
 230 $\overline{\lambda_P}$ and its deviation λ'_P , λ_{LR} is the lapse rate feedback, λ_{WV} is the water vapor feedback, λ_{AL} is the
 231 surface albedo feedback and λ_{CL} is the cloud feedback.

232 To apply the conventional attribution method to the GCM simulations, we use aquaplanet ker-
 233 nels from Feldl et al. (2017b) (available at <https://github.com/nfeldl/aquakernels>) to calculate the
 234 feedbacks. The CO₂ forcing \mathcal{F} is computed as the change in TOA net radiation between the con-
 235 trol simulation and an idealized GCM simulation where sea surface temperatures (SST) are fixed
 236 to the control SST and CO₂ concentrations are quadrupled (Hansen et al. 2005). The change in

237 atmospheric energy transport convergence $\Delta(\nabla \cdot \vec{F})$ is computed as the change in net TOA radia-
238 tion (minus the surface forcing) between the control and perturbed simulations. This method of
239 attributing surface temperature changes to forcings and feedbacks then tells us how much surface
240 temperature change is required to balance the TOA energy imbalance caused by each forcing or
241 feedback, assuming the atmospheric temperature change is vertically uniform (except for the lapse
242 rate feedback).

243 Figure 4 compares this TOA energy budget surface temperature change attribution method
244 (crosses) with the single column model based attribution method (filled circles) for the $4xCO_2$
245 (a), $4xCO_2$ with $Q_s = 12W m^{-2}$ (b) and $Q_s = 24W m^{-2}$ (c). The tropical (x-axis, $10^\circ S$ to $10^\circ N$)
246 and polar (y-axis, $80^\circ N$ to $90^\circ N$) attributions are plotted against each other. If a point falls above
247 (below) the one-to-one line, the forcing or feedback contributes to polar (tropical) amplification.
248 As in Pithan and Mauritsen (2014), the TOA attribution method suggests that the Planck and lapse
249 rate feedbacks contribute to polar amplification. The lapse rate feedback contributes to more po-
250 lar amplification in the surface heat source experiments. The single column model attribution
251 method, in contrast, has no temperature feedback in its decomposition. Since the TOA energy
252 budget method assumes that the temperature response to a TOA energy imbalance is vertically
253 uniform, it will attribute a larger (smaller) amplitude change in surface temperature than the sin-
254 gle column model if the response to the forcing or feedback is top-heavy (bottom-heavy). In the
255 tropics, all temperature changes are top-heavy as they follow the moist adiabat, hence the SCM
256 attributions are all closer to the y-axis than the corresponding TOA method attributions. In the
257 high latitudes, the SCM temperature changes from increases in CO_2 , water vapor, and surface
258 heat source are bottom-heavy, hence they all contribute a larger surface temperature change than
259 is diagnosed from the TOA method. The energy transport convergence change leads to top-heavy

260 warming, hence the warming attributed to it by the SCM method is smaller than the warming
261 attributed by the TOA method, and even negative in the surface heat source cases.

262 In summary, we underline two main points from this comparison of the single column model
263 and TOA-based surface temperature change attribution methods:

264 • The increase in longwave absorbers (CO₂ and water vapor) go from contributing to tropical
265 amplification in the TOA attribution method to contributing to polar amplification in the SCM
266 attribution method. The forcing from CO₂ and the water vapor feedback are stronger in
267 the tropics than the high latitudes, but since the tropical SCM attribution includes the effect
268 of convection, the warming maximum is pushed to the upper-troposphere and there is less
269 surface warming. In the high latitudes however, an increase in longwave absorbers leads to
270 bottom-heavy warming (Cronin and Jansen 2016; Henry and Merlis 2019). Russotto and
271 Biasutti (2020) analyze the response of atmospheric GCMs using a moist energy balance
272 model, and similarly find that a tropically amplified CO₂ forcing and water vapor feedback
273 lead to a polar amplified temperature response.

274 • Since the increase in atmospheric energy transport convergence preferentially affects the mid-
275 troposphere, it leads to less surface warming at high latitudes, and even surface cooling in the
276 surface heat source experiments.

277 **6. Local and remote drivers of temperature change.**

278 The SCM attribution method can also be used to decompose the drivers of polar amplification
279 into local and remote drivers. The CO₂ and surface heat source perturbations are local drivers,
280 while the energy transport can be considered as a remote driver. The water vapor feedback includes
281 both local and remote contributions. First, the change in specific humidity can be decomposed

282 into a temperature-dependent change and a change due to relative humidity: $\Delta q = \Delta q|_{fixedRH} +$
 283 $\Delta RH \times q^*|_{clim}$ where $q^*|_{clim}$ is the climatological saturation specific humidity. Since the relative
 284 humidity in the idealized GCM stays relatively constant (supplementary figure S4), we ignore
 285 the second term of this equation. Using fixed relative humidity (RH) SCM experiments, we can
 286 decompose the temperature-dependent changes in specific humidity into the ‘local’ changes in
 287 response to the temperature changes forced by increased CO₂ and the surface heat source, and
 288 the ‘remote’ changes in response to the temperature change forced by altered energy transports:
 289 $\Delta q \approx \Delta q|_{fixedRH} = \Delta q|_{fixedRH, \Delta CO_2, \Delta Q_s} + \Delta q|_{fixedRH, \Delta ET}$.

290 This local versus remote decomposition of the water vapor concentration increase is not perfect,
 291 as it assumes the energy transport simply affects the humidity of the high latitudes by changing
 292 its temperature and activating the local water vapor feedback, whereas the general circulation can
 293 directly advect water vapor. The energy transport term also contains vertical advection, which can
 294 change as a result of local diabatic forcings (shown in magenta in supplementary figure S2). More-
 295 over, GCM experiments where the forcing from a CO₂ increase is constrained to the high latitudes
 296 show changes in energy transport, which would also affect the water vapor feedback (Stuecker
 297 et al. 2018). Since energy transport is affected by both temperature and humidity gradients, it is
 298 not clear that any perfect local / remote decomposition exists. Nevertheless, our definition of ‘lo-
 299 cal’ recovers traditional SCM treatments of fixed relative humidity water vapor feedback (Manabe
 300 and Wetherald 1967) in the limit of no energy transport change.

301 The fixed-RH SCM simulations have the same modules and parameters as the standard SCM
 302 simulations, but instead of prescribing the idealized GCM’s specific humidity, they have fixed rel-
 303 ative humidity and the specific humidity is free to evolve with the temperature. The climatological
 304 temperature of the fixed RH SCMs have a warm bias (supplementary figure S5) and the climato-
 305 logical specific humidity is biased high (supplementary figures S6). We do two sets of fixed-RH

306 SCM experiments: the first ('local') experiment is forced with the increase in CO₂ concentration
307 (and surface heat source), and the second is forced with increased CO₂ concentration (and surface
308 heat source) and perturbed energy transport. The latter has less tropical warming and similar polar
309 warming compared to the idealized GCM (red lines in supplementary figure S7 for the 4xCO₂
310 experiment), and similar changes in specific humidity in the tropics and a higher increase in high
311 latitudes compared to the idealized GCM (red lines in supplementary figures S8 for the 4xCO₂
312 experiment). The 'local' increase in water vapor, $\Delta q|_{fixedRH,\Delta CO_2,\Delta Q_s}$, is taken to be the change in
313 water vapor from the first set of fixed-RH SCM experiments (blue lines in figure S8 for the 4xCO₂
314 experiment), and the 'remote' increase in water vapor, $\Delta q|_{fixedRH,\Delta ET}$, is taken to be the residual
315 between the total change in water vapor and the 'local' change in water vapor. We then force the
316 original SCM with the 'local' and 'remote' specific humidity changes to deduce the 'q (local)' and
317 'q (remote)' temperature changes (shown in table 2). The 'q (local)' experiments are comparable
318 to the fixed RH experiments in Payne et al. (2015). The temperature changes from the high latitude
319 'q (local)' experiments are shown in figure 3 (blue dashed).

320 Table 2 summarizes the result of this local / remote decomposition of surface temperature
321 change. In the three perturbation experiments, the warming from CO₂ alone is 1.8K in the tropics
322 and 3.1K at high latitudes, hence increasing CO₂ leads to polar amplification in the absence of any
323 feedbacks. The addition of the 'local' water vapor feedback increases the tropical surface warm-
324 ing to 7.5K and the polar surface warming to 4.0K in the 4xCO₂ experiment, and thus cancels
325 the polar amplification from CO₂ alone. Payne et al. (2015) also found a tropical amplification of
326 surface temperature change in their fixed-RH SCM simulations, though with somewhat different
327 magnitude. Finally, adding the atmospheric energy transport and its implied water vapor change
328 decreases the tropical surface warming to 3.7K, and increases the polar surface warming to 9.0K
329 in the 4xCO₂ experiment, thus leading to polar amplification. The polar surface heat source gen-

330 erally increases the amount of polar amplification despite the partial compensation by a reduction
331 in dry energy transport. For the $4\times\text{CO}_2$ experiment, approximately half of the polar warming is
332 due to local sources (4.0K out of 9K of total warming), but the polar amplified pattern of warming
333 is primarily caused by the increase in atmospheric energy transport which cools the tropics and
334 warms the high latitudes. The high latitude warming is then strongly enhanced by the increased
335 water vapor from remote sources. When a polar surface heat source is added, almost all of the
336 polar surface warming is due to local sources because of the surface heat source and the compen-
337 sating reduction in the dry component of energy transport: 10.2K and 16.1K from local sources
338 for a total warming of 12.6K and 16.6K for the $Q_s = 12\text{W m}^{-2}$ and $Q_s = 24\text{W m}^{-2}$ experiments,
339 respectively.

340 **7. Summary and discussion**

341 Unlike the tropics which are close to radiative-convective equilibrium, the high latitudes are
342 in radiative-advective equilibrium: different forcings and feedbacks induce different lapse rate
343 responses. Previous surface temperature attributions have assumed that different forcings and
344 feedbacks induce vertically homogeneous warming, and attributed the deviation from vertically
345 uniform warming to the lapse rate feedback. In these attributions, the lapse rate feedback functions
346 as a residual that cannot be clearly ascribed to any particular physical process.

347 We introduce a surface temperature change attribution method based on a single column model,
348 which accounts for the vertically inhomogeneous temperature change contributions of each forc-
349 ing and feedback. We find that the warming from increased longwave absorbers (CO_2 and water
350 vapor) is bottom-heavy and accounts for most of the surface warming in the absence of a surface
351 heat source. By contrast, the warming from atmospheric heat transport preferentially warms the
352 mid and upper troposphere. When a polar surface heat source is added, there is a reduction in the

353 dry component of atmospheric energy transport which partially compensates for the extra surface
354 warming from the polar surface heat source. Compared to the conventional surface temperature
355 change attribution method, the increase in longwave absorbers (CO₂ and water vapor) goes from
356 contributing to tropical amplification to polar amplification. In addition, the polar warming contri-
357 bution from the increase in atmospheric energy transport convergence is reduced as it preferentially
358 warms the mid and upper troposphere. Finally, we separated the drivers of atmospheric tempera-
359 ture change into local and remote contributors and found that, in the absence of a polar surface heat
360 source, the change in energy transport was primarily responsible for the polar amplified pattern of
361 warming. The addition of a polar surface heat source increases the contribution of local drivers to
362 polar warming at the expense of remote drivers, as the dry energy transport is reduced.

363 It is important to note that clouds and sea ice were ignored in this analysis (aside from the surface
364 heat source that mimics the effects of shortwave cloud feedbacks and sea ice), though they may
365 play an important role in explaining the pattern of surface temperature change in comprehensive
366 climate model simulations. Arctic amplification also has seasonality — it is strong in winter and
367 suppressed in summer — which has been suggested to result from the increased polar ocean heat
368 uptake in summer and ocean heat release in winter from the melting sea ice (Manabe and Stouffer
369 1980; Bintanja and Van der Linden 2013; Dai et al. 2019). Nevertheless, we believe that the single-
370 column model can be a stepping stone for connecting simple physical models with comprehensive
371 climate models: clouds and seasonality can be prescribed in the SCM, which would be a valuable
372 extension of the present work. This would allow us to understand the basic mechanisms driving
373 Arctic amplification and reduce the uncertainty in the rate of Arctic warming.

374 *Acknowledgments.* The code and data needed to reproduce all figures, tables and supplementary
375 figures will be made available at https://github.com/matthewjhenry/HMLR19_SCM. Documenta-
376 tion for the python ClimLab package can be found at <https://climlab.readthedocs.io/>. The top-of-
377 atmosphere albedo data from the Cloud and the Earth’s Radiant Energy System (CERES) can be
378 found at <https://ceres.larc.nasa.gov/>. The CMIP6 data is available on the Earth System Grid Fed-
379 eration database. This work was supported by a Natural Sciences and Research Council (NSERC)
380 Discovery grant and Canada Research Chair, as well as a Compute Canada allocation. B.E.J.R
381 was supported by NSF grant AGS-1455071.

382 **References**

- 383 Abbot, D. S., and E. Tziperman, 2008: Sea ice, high-latitude convection, and equable climates.
384 *Geophysical Research Letters*, **35** (3).
- 385 Anderson, J. L., and Coauthors, 2004: The new GFDL global atmosphere and land model AM2–
386 LM2: Evaluation with prescribed SST simulations. *Journal of Climate*, **17** (24), 4641–4673.
- 387 Bintanja, R., and E. Van der Linden, 2013: The changing seasonal climate in the arctic. *Scientific*
388 *Reports*, **3**, 1556.
- 389 Cronin, T. W., and M. F. Jansen, 2016: Analytic radiative-advective equilibrium as a model for
390 high-latitude climate. *Geophysical Research Letters*, **43** (1), 449–457.
- 391 Dai, A., D. Luo, M. Song, and J. Liu, 2019: Arctic amplification is caused by sea-ice loss under
392 increasing co 2. *Nature communications*, **10** (1), 121.
- 393 Eyring, V., S. Bony, G. A. Meehl, C. A. Senior, B. Stevens, R. J. Stouffer, and K. E. Taylor, 2016:
394 Overview of the coupled model intercomparison project phase 6 (CMIP6) experimental design
395 and organization. *Geoscientific Model Development (Online)*, **9** (LLNL-JRNL-736881).

396 Feldl, N., B. T. Anderson, and S. Bordoni, 2017a: Atmospheric eddies mediate lapse rate feedback
397 and arctic amplification. *Journal of Climate*, **30** (22), 9213–9224.

398 Feldl, N., and S. Bordoni, 2016: Characterizing the Hadley circulation response through regional
399 climate feedbacks. *J. Climate*, **29** (2), 613–622.

400 Feldl, N., S. Bordoni, and T. M. Merlis, 2017b: Coupled high-latitude climate feedbacks and their
401 impact on atmospheric heat transport. *Journal of Climate*, **30** (1), 189–201.

402 Frierson, D. M., 2007: The dynamics of idealized convection schemes and their effect on the
403 zonally averaged tropical circulation. *J. Atmos. Sci.*, **64** (6), 1959–1976.

404 Frierson, D. M., I. M. Held, and P. Zurita-Gotor, 2006: A gray-radiation aquaplanet moist GCM.
405 part I: Static stability and eddy scale. *Journal of the Atmospheric Sciences*, **63** (10), 2548–2566.

406 Hansen, J., and Coauthors, 2005: Efficacy of climate forcings. *J. Geophys. Res.*, **110**, D18 104.

407 Henry, M., and T. M. Merlis, 2019: Forcing dependence of atmospheric lapse rate changes domi-
408 nates residual polar warming in solar radiation management scenarios. (*In review*).

409 Hwang, Y.-T., and D. M. Frierson, 2010: Increasing atmospheric poleward energy transport with
410 global warming. *Geophysical Research Letters*, **37** (24).

411 Hwang, Y.-T., D. M. Frierson, and J. E. Kay, 2011: Coupling between Arctic feedbacks and
412 changes in poleward energy transport. *Geophysical Research Letters*, **38** (17).

413 Jucker, M., and E. Gerber, 2017: Untangling the annual cycle of the tropical tropopause layer with
414 an idealized moist model. *Journal of Climate*, **30** (18), 7339–7358.

415 Laliberté, F., and P. Kushner, 2013: Isentropic constraints by midlatitude surface warming on the
416 arctic midtroposphere. *Geophysical Research Letters*, **40** (3), 606–611.

417 Loeb, N. G., and Coauthors, 2018: Clouds and the earth's radiant energy system (CERES) en-
418 ergy balanced and filled (EBAF) top-of-atmosphere (TOA) edition-4.0 data product. *Journal of*
419 *Climate*, **31** (2), 895–918.

420 Manabe, S., R. Stouffer, M. Spelman, and K. Bryan, 1991: Transient responses of a coupled
421 ocean–atmosphere model to gradual changes of atmospheric CO₂. Part 1. annual mean response.
422 *J. Climate*, **4** (8), 785–818.

423 Manabe, S., and R. J. Stouffer, 1980: Sensitivity of a global climate model to an increase of CO₂
424 concentration in the atmosphere. *Journal of Geophysical Research: Oceans*, **85** (C10), 5529–
425 5554.

426 Manabe, S., and R. T. Wetherald, 1967: Thermal equilibrium of the atmosphere with a given
427 distribution of relative humidity. *Journal of the Atmospheric Sciences*, **24** (3), 241–259.

428 Merlis, T. M., T. Schneider, S. Bordoni, and I. Eisenman, 2013: Hadley circulation response to
429 orbital precession. part I: Aquaplanets. *Journal of Climate*, **26** (3), 740–753.

430 Payne, A. E., M. F. Jansen, and T. W. Cronin, 2015: Conceptual model analysis of the influence
431 of temperature feedbacks on polar amplification. *Geophys. Res. Lett.*, 9561–9570.

432 Pithan, F., and T. Mauritsen, 2014: Arctic amplification dominated by temperature feedbacks in
433 contemporary climate models. *Nat. Geosci.*, **7**, 181–184.

434 Roe, G. H., N. Feldl, K. C. Armour, Y.-T. Hwang, and D. M. Frierson, 2015: The remote impacts
435 of climate feedbacks on regional climate predictability. *Nature Geoscience*, **8** (2), 135.

436 Rose, B. E., 2018: Climlab: A python toolkit for interactive, process oriented climate modeling.
437 *J. Open Source Software*, **3** (24), 659.

- 438 Rugestein, M., and Coauthors, 2019: LongRunMIP—motivation and design for a large collection
439 of millennial-length AO-GCM simulations. *Bulletin of the American Meteorological Society*,
440 **(2019)**.
- 441 Russotto, R. D., and M. Biasutti, 2020: Polar amplification as an inherent response of a circu-
442 lating atmosphere: results from the TRACMIP aquaplanets. *Geophysical Research Letters*, **(In**
443 **Review)**.
- 444 Screen, J. A., C. Deser, and I. Simmonds, 2012: Local and remote controls on observed arctic
445 warming. *Geophysical Research Letters*, **39 (10)**.
- 446 Stocker, T. F., and Coauthors, Eds., 2013: *Climate Change 2013: The Physical Science Basis*.
447 Cambridge University Press, Cambridge and New York.
- 448 Stuecker, M. F., and Coauthors, 2018: Polar amplification dominated by local forcing and feed-
449 backs. *Nature Climate Change*, **8 (12)**, 1076.

450 **LIST OF TABLES**

451 **Table 1.** Surface temperature change attribution based on the single column model de-
452 composition for the three perturbation experiments. ‘CO₂’ and ‘Water Vapor’
453 denote the radiative effect of their increase on surface temperature, whereas
454 ‘ET’ denotes the effect of the change in energy transport on surface tempera-
455 ture and is decomposed into its dry and moist components in the pole. ‘Q_s’
456 denotes the effect of the surface heat source on the surface temperature change. . . . 24

457 **Table 2.** Surface temperature change attribution based on the single column model de-
458 composition for the three perturbation experiments. The tropical surface tem-
459 perature change attributions are sufficiently similar to be in a single column.
460 The three successive values separated by a comma refer to the the 4xCO₂,
461 Q_s = 12W m⁻², and Q_s = 24W m⁻² experiments respectively. Slight discrep-
462 ancies between the total and the sum of local and remote totals occur as the total
463 is the surface temperature change from the experiment with all perturbations. . . . 25

Forcing (W m^{-2}) / Feedback ($\text{W m}^{-2} \text{K}^{-1}$)	4xCO ₂	4xCO ₂ + 12 W m ⁻²	4xCO ₂ + 24 W m ⁻²
Tropics			
CO ₂	1.8	1.8	1.8
Water Vapor	2.8	2.9	2.9
ET	-0.8	-0.5	-0.5
Tropics total	3.7	4.0	4.1
Pole			
CO ₂	3.1	3.1	3.1
Water Vapor	4.3	4.8	5.5
ET (dry)	0	-2.1	-4.1
ET (moist)	1.4	2.0	2.8
Q_s	0	4.6	8.9
Pole total	9.0	12.6	16.6

464 TABLE 1. Surface temperature change attribution based on the single column model decomposition for the
465 three perturbation experiments. ‘CO₂’ and ‘Water Vapor’ denote the radiative effect of their increase on surface
466 temperature, whereas ‘ET’ denotes the effect of the change in energy transport on surface temperature and is
467 decomposed into its dry and moist components in the pole. ‘ Q_s ’ denotes the effect of the surface heat source on
468 the surface temperature change.

Forcing / feedback	Tropics	Pole (4xCO ₂)	Pole (4xCO ₂ +12)	Pole (4xCO ₂ +24)
CO ₂	1.8	3.1	3.1	3.1
q (local)	5.7	0.9	2.5	4.1
Q_s	0	0	4.6	8.9
Local total	7.5	4.0	10.2	16.1
q (remote)	-2.9,-2.8,-2.8	3.4	2.3	1.4
ET	-0.8,-0.5,-0.5	1.5	-0.1	-1.4
Remote total	-3.8,-3.4,-3.4	4.9	2.2	0
Total	3.7,4.0,4.1	9.0	12.6	16.6

469 TABLE 2. Surface temperature change attribution based on the single column model decomposition for the
470 three perturbation experiments. The tropical surface temperature change attributions are sufficiently similar to
471 be in a single column. The three successive values separated by a comma refer to the the 4xCO₂, $Q_s = 12\text{W m}^{-2}$,
472 and $Q_s = 24\text{W m}^{-2}$ experiments respectively. Slight discrepancies between the total and the sum of local and
473 remote totals occur as the total is the surface temperature change from the experiment with all perturbations.

474
475
476
477
478
479
480
481
482

483
484
485

486
487
488
489
490
491
492
493
494
495
496
497

498
499
500
501
502
503
504
505

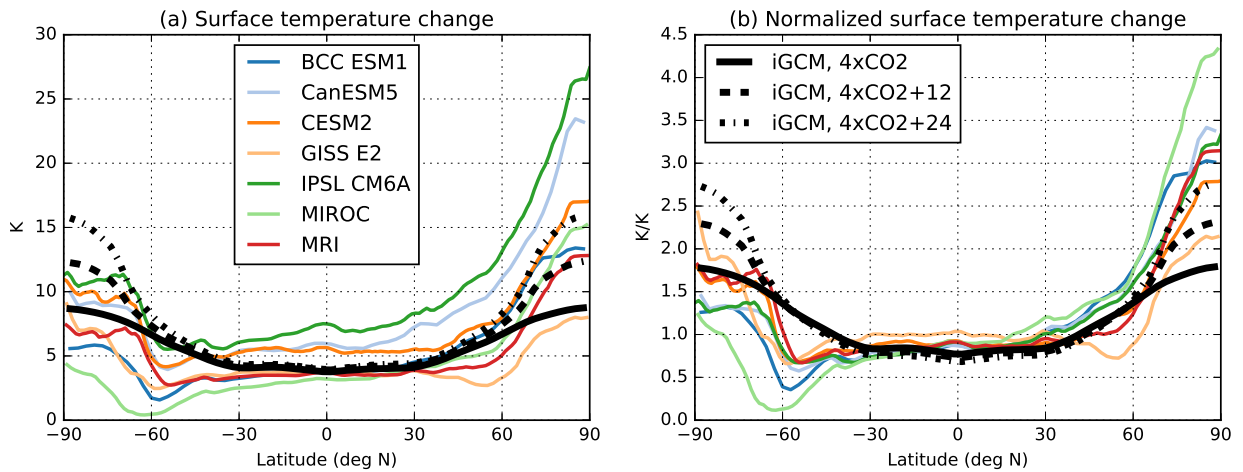
LIST OF FIGURES

Fig. 1. (a) Surface temperature difference between the control experiment (300ppm CO₂ concentration) and increased CO₂ experiment (1200ppm) (black) and increased CO₂ experiment (1200ppm) with a 12 W m⁻² (black dashed) and 24 W m⁻² (black dash-dot) surface heat source poleward of 80° using an idealized moist atmospheric GCM with no clouds or sea ice. These are compared to CMIP6 abrupt 4xCO₂ surface temperature changes in simulations with the following models: BCC-ESM1, CanESM5, CESM2, GISS-E2, IPSL CM6A, MIROC, MRI. (b) Same as (a), but the temperature changes are normalized by global mean surface temperature change. 27

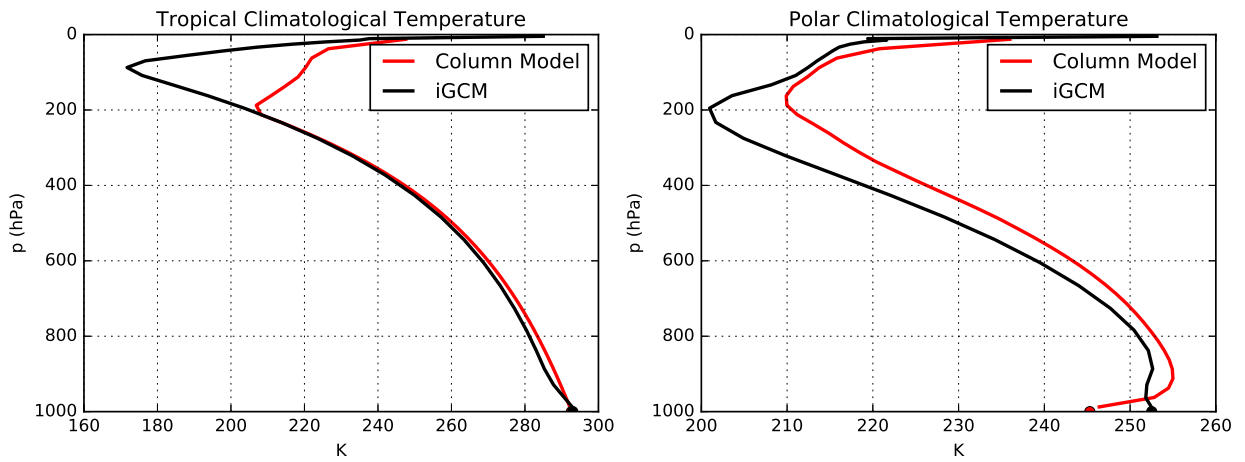
Fig. 2. Comparison between the single column model (red) and idealized GCM (black) for the (a) tropical (|lat|<10°) climatological temperature and (b) the polar (lat>80° North) climatological temperature. 28

Fig. 3. Tropical (a) and polar (b,c,d) temperature change for the idealized GCM (grey) and three perturbation experiments using the single column model: 4xCO₂ (a,b), 4xCO₂ with 12 W m⁻² surface heat source poleward of 80° (a,c), and 4xCO₂ with 24 W m⁻² surface heat source poleward of 80° (a,d). The SCM experiments with all changes (black) are exactly the same as the sum of individual changes and fit the idealized GCM (grey) well. The individual forcing and feedback contributions are calculated by individually perturbing them in the single column model (colors). They include the CO₂ increase (red), the water vapor feedback (blue), the ‘local’ water vapor feedback (blue dashed, see section 6), the energy transport (green in tropics, separated into dry (orange) and moist (cyan) in high latitudes), and the surface heat source (yellow). The tropical temperature changes of the three experiments (a) are similar enough to be plotted together (12 W m⁻² in dashed lines and 24 W m⁻² in dash-dotted lines). Surface temperature change attributions are summarized in table 1. 29

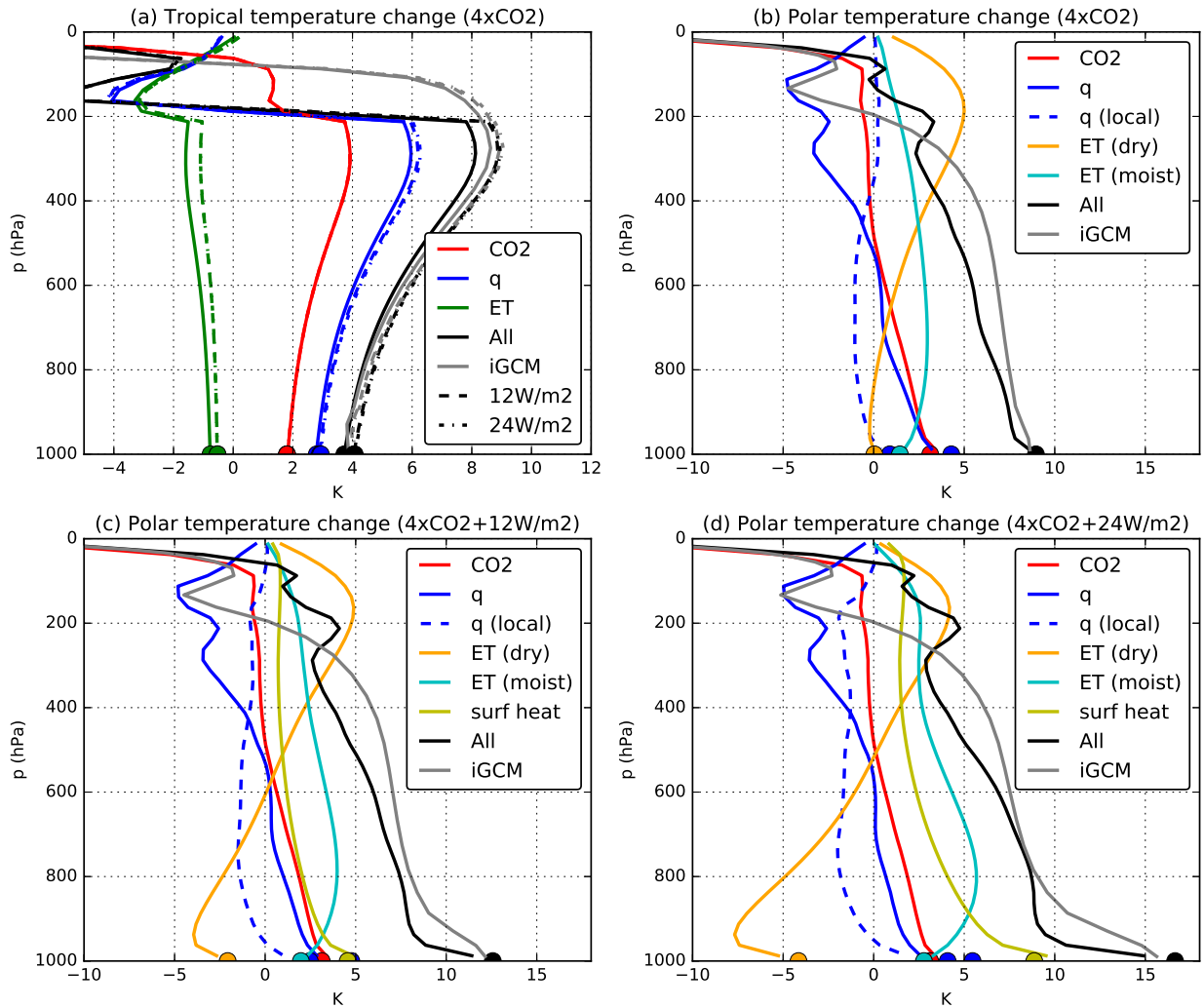
Fig. 4. Surface temperature change attributions for the 4xCO₂ (a) and 4xCO₂ with $Q_s = 12\text{W m}^{-2}$ (b) and $Q_s = 24\text{W m}^{-2}$ (c) using the TOA energy budget method (crosses) and the SCM method presented in this paper (dots). Presented are the surface temperature change attributions to the increase in CO₂ (red), water vapor (blue), energy transport (green), surface heat source (yellow) for both methods. The Planck (magenta) and lapse rate (cyan) feedback contributions are also shown for the TOA energy budget method. When the point is above (below) the one-to-one line, the forcing or feedback contributes to polar (tropical) amplification. 30



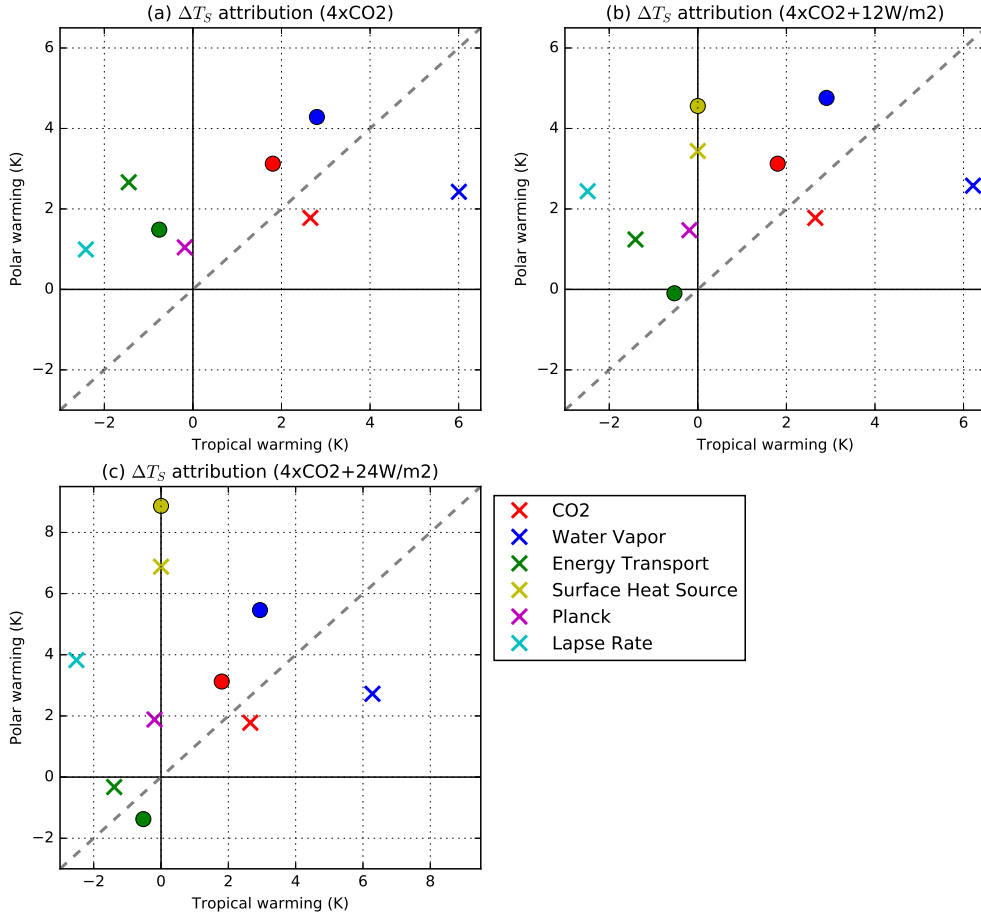
506 FIG. 1. (a) Surface temperature difference between the control experiment (300ppm CO₂ concentration)
 507 and increased CO₂ experiment (1200ppm) (black) and increased CO₂ experiment (1200ppm) with a 12 W m⁻²
 508 (black dashed) and 24 W m⁻² (black dash-dot) surface heat source poleward of 80° using an idealized moist
 509 atmospheric GCM with no clouds or sea ice. These are compared to CMIP6 abrupt 4xCO₂ surface temperature
 510 changes in simulations with the following models: BCC-ESM1, CanESM5, CESM2, GISS-E2, IPSL CM6A,
 511 MIROC, MRI. (b) Same as (a), but the temperature changes are normalized by global mean surface temperature
 512 change.



513 FIG. 2. Comparison between the single column model (red) and idealized GCM (black) for the (a) tropical
 514 ($|\text{lat}| < 10^\circ$) climatological temperature and (b) the polar ($\text{lat} > 80^\circ$ North) climatological temperature.



515 FIG. 3. Tropical (a) and polar (b,c,d) temperature change for the idealized GCM (grey) and three perturbation
 516 experiments using the single column model: $4xCO_2$ (a,b), $4xCO_2$ with 12 W m^{-2} surface heat source poleward
 517 of 80° (a,c), and $4xCO_2$ with 24 W m^{-2} surface heat source poleward of 80° (a,d). The SCM experiments with
 518 all changes (black) are exactly the same as the sum of individual changes and fit the idealized GCM (grey) well.
 519 The individual forcing and feedback contributions are calculated by individually perturbing them in the single
 520 column model (colors). They include the CO_2 increase (red), the water vapor feedback (blue), the ‘local’ water
 521 vapor feedback (blue dashed, see section 6), the energy transport (green in tropics, separated into dry (orange)
 522 and moist (cyan) in high latitudes), and the surface heat source (yellow). The tropical temperature changes of
 523 the three experiments (a) are similar enough to be plotted together (12 W m^{-2} in dashed lines and 24 W m^{-2} in
 524 dash-dotted lines). Surface temperature change attributions are summarized in table 1.



525 FIG. 4. Surface temperature change attributions for the 4xCO₂ (a) and 4xCO₂ with $Q_s = 12\text{W m}^{-2}$ (b) and
 526 $Q_s = 24\text{W m}^{-2}$ (c) using the TOA energy budget method (crosses) and the SCM method presented in this paper
 527 (dots). Presented are the surface temperature change attributions to the increase in CO₂ (red), water vapor (blue),
 528 energy transport (green), surface heat source (yellow) for both methods. The Planck (magenta) and lapse rate
 529 (cyan) feedback contributions are also shown for the TOA energy budget method. When the point is above
 530 (below) the one-to-one line, the forcing or feedback contributes to polar (tropical) amplification.