

Clouds and the atmospheric circulation response to warming

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1 **Clouds and the atmospheric circulation response to warming**

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ABSTRACT

7 We study the effect of clouds on the atmospheric circulation response to
8 CO₂ quadrupling in an aquaplanet model with a slab-ocean lower bound-
9 ary. The cloud effect is isolated by locking the clouds to either the control
10 or 4xCO₂ state in the shortwave (SW) or longwave (LW) radiation schemes.
11 In our model, cloud-radiative changes explain more than half of the total pole-
12 ward expansion of the Hadley cells, midlatitude jets, and storm tracks under
13 CO₂ quadrupling, even though they cause only one-fourth of the total global-
14 mean surface warming. The effect of clouds on circulation results mainly
15 from the SW cloud-radiative changes, which strongly enhance the Equator-
16 to-pole temperature gradient at all levels in the troposphere, favoring stronger
17 and poleward-shifted midlatitude eddies. By contrast, quadrupling CO₂ while
18 holding the clouds fixed causes strong polar amplification and weakened mid-
19 latitude baroclinicity at lower levels, yielding only a small poleward expan-
20 sion of the circulation. Our results show that (a) the atmospheric circulation
21 responds sensitively to cloud-driven changes in meridional and vertical tem-
22 perature distribution, and (b) the spatial structure of cloud feedbacks likely
23 plays a dominant role in the circulation response to greenhouse gas forcing.
24 While the magnitude and spatial structure of the cloud feedback are expected
25 to be highly model-dependent, an analysis of 4xCO₂ simulations of CMIP5
26 models shows that the SW cloud feedback likely forces a poleward expansion
27 of the tropospheric circulation in most climate models.

28 **1. Introduction**

29 Clouds exert a very substantial effect on the energy balance of the Earth's atmosphere through
30 their effects on shortwave (SW) and longwave (LW) radiation, with an approximate global-mean
31 effect of -20 W m^{-2} (Boucher et al. 2013). With increasing greenhouse gas forcing, the SW and
32 LW radiative effects of clouds are expected to change, and while the magnitude of this change is
33 highly uncertain, most climate models predict a positive global-mean forcing from cloud changes
34 – a positive cloud feedback (Soden et al. 2008; Vial et al. 2013). Previous research has mainly
35 focused on the impact of cloud feedbacks on the global energy balance and climate sensitivity
36 (e.g., Soden et al. 2008; Zelinka and Hartmann 2010; Zelinka et al. 2012; Vial et al. 2013). How-
37 ever, cloud feedbacks also possess rich spatial structures, and hence they affect spatial patterns
38 of warming (Roe et al. 2015), meridional energy transport by atmospheric motions (Hwang and
39 Frierson 2010; Zelinka and Hartmann 2012), and likely also the atmospheric circulation (Ceppi
40 et al. 2014; Voigt and Shaw 2015).

41 While quantitative aspects of the circulation response to greenhouse gas forcing remain highly
42 uncertain, robust qualitative aspects of the response include a weakening of the Hadley circulation
43 (Held and Soden 2006; Vecchi and Soden 2007), a rise of the tropopause and upward expansion of
44 the circulation (e.g., Lorenz and DeWeaver 2007), and a poleward expansion of the Hadley cells,
45 midlatitude jets, and storm tracks (Kushner et al. 2001; Yin 2005; Lu et al. 2007; Frierson et al.
46 2007; Chang et al. 2012; Barnes and Polvani 2013). How clouds contribute to shaping such circu-
47 lation changes is presently not well understood. It is also unclear to what extent the uncertainty in
48 the cloud feedbacks affects the inter-model spread in atmospheric circulation changes; it has been
49 suggested that this effect could be substantial in the case of the midlatitude jet response (Ceppi
50 et al. 2014).

51 The purpose of this paper is to quantitatively assess the effect of cloud-radiative changes on
52 the atmospheric circulation response to CO₂ increase in a climate model. Here, we use an aqua-
53 planet model with interactive sea surface temperature to demonstrate that clouds can cause a very
54 substantial enhancement of the circulation response to CO₂ quadrupling. Overall, clouds explain
55 more than half of the total poleward expansion of the circulation in our model. This occurs mainly
56 through the SW effect of clouds, which acts to strongly increase the Equator-to-pole temperature
57 gradient and make the midlatitudes more baroclinically unstable. Remarkably, CO₂ quadrupling
58 only yields a weak poleward expansion of the circulation if the clouds are held fixed, indicating
59 that the cloud response is a key influence on the circulation changes predicted by our model. Be-
60 cause clouds have such a strong effect, the results presented here suggest that cloud feedbacks
61 could significantly contribute to the uncertainty in the atmospheric circulation response to global
62 warming, highlighting the need for better constraints on the cloud response in climate models.

63 We begin by presenting the methodology used to isolate the effect of cloud-radiative changes on
64 atmospheric circulation in our climate model in section 2. In section 3, we then present the key
65 results of our experiments, followed by a discussion in section 4, and a summary and concluding
66 remarks in section 5.

67 **2. Methods**

68 The atmospheric model used in this study is the Geophysical Fluid Dynamics Laboratory
69 (GFDL) AM2.1 (The GFDL Global Atmospheric Model Development Team 2004). It is run in
70 aquaplanet configuration, coupled to a slab-ocean lower boundary representing a mixed layer of
71 50 m depth. While there is no seasonal cycle, insolation is set to its annual-mean value at every
72 latitude. The model also has no sea ice, but the sea surface temperature can be below freezing.
73 We study the effects of cloud feedbacks on atmospheric circulation by comparing two model cli-

74 matologies with identical boundary conditions except for CO₂ forcing. These two climates, which
75 we describe as CTL and 4xCO₂, have CO₂ mixing ratios of 348 and 1392 ppm, respectively.

76 We use the cloud-locking method to assess the effect of cloud-radiative changes on the atmo-
77 spheric circulation response. This method involves prescribing clouds from two different climate
78 states in the climate model's radiation code, to obtain the effect of cloud changes in isolation. In
79 our case, the two climate states that the clouds are "locked" to are CTL and 4xCO₂. Note that only
80 the radiation code experiences the locked clouds, which override the cloud-radiative properties
81 simulated by the model interactively; all other model components (e.g. the cloud microphysics,
82 or the large-scale condensation scheme) use the model's internally simulated clouds. Locking
83 of model fields such as clouds and water vapor as a method to quantify feedback processes has
84 been successfully implemented in many studies (e.g., Wetherald and Manabe 1980, 1988; Hall and
85 Manabe 1999; Schneider et al. 1999; Langen et al. 2012; Mauritsen et al. 2013; Voigt and Shaw
86 2015). Unlike previous studies, however, we discriminate between SW and LW cloud effects by
87 separately prescribing cloud-radiative properties in the SW and LW radiation schemes.

88 When locking clouds, it is necessary to use the full time-varying cloud-radiative properties,
89 rather than time-averaged values. This is because cloud-radiative properties (e.g. cloud optical
90 depth) and cloud-radiative effects are generally not linearly related, so that using time-mean cloud
91 properties would yield a large climate bias. We therefore prescribe instantaneous cloud-radiative
92 properties taken from every call of the radiation code. As discussed in previous studies (Schneider
93 et al. 1999; Mauritsen et al. 2013; Voigt and Shaw 2015), prescribing cloud properties at every time
94 step results in the loss of the spatio-temporal correlation between cloud, moisture, and temperature
95 anomalies, which may cause a bias in the mean climate. For example, the radiation code could
96 experience cloud-free conditions in a gridbox in which ascent and condensation are occurring,
97 because the prescribed cloud-radiative properties are decorrelated from the weather. We will show

98 in the next section that this climate bias is small, however, and is unlikely to affect our conclusions.
99 To ensure that variables are similarly decorrelated in all experiments, the prescribed cloud fields
100 are offset by one year relative to the model’s simulated climate.

101 The cloud-locked experiments are performed as follows. We first run the CTL and $4\times\text{CO}_2$
102 experiments with interactive clouds for twenty years (after discarding two years of model spin-up),
103 and save all cloud variables used in the model’s radiation scheme at every call of the radiation code
104 (every 6 h). We then use the cloud-radiative properties output by the interactive CTL and $4\times\text{CO}_2$
105 simulations to run a total of eight cloud-locked simulations, involving all possible combinations
106 of CO_2 concentration G, SW cloud-radiative properties S, and LW cloud-radiative properties L.
107 Denoting the CTL and $4\times\text{CO}_2$ states by numbers 1 and 2, respectively, the eight experiments
108 are G1S1L1, G2S1L2, G1S2L1, G1S1L2, G2S2L1, G2S1L2, G1S2L2, and G2S2L2. In each of
109 these cloud-locked simulations, the time-varying cloud properties from either the CTL or $4\times\text{CO}_2$
110 simulation are read in at every time step, and override the cloud properties calculated by the model.
111 Separately locking SW and LW cloud-radiative properties is possible because the AM2.1 radiation
112 scheme uses different cloud properties in the SW and LW schemes.

113 Locking the model clouds allows us to calculate the separate effects of changing clouds while
114 keeping CO_2 levels fixed, and increasing CO_2 while keeping the clouds fixed. For simplicity, here-
115 after we refer to these components as the “effect of cloud-radiative changes,” and the “effect of
116 CO_2 increase,” but it must be kept in mind that each of these effects includes additional contribu-
117 tions from other climate feedbacks (see discussion below). We calculate the effects of clouds and
118 CO_2 increase using a method similar to Voigt and Shaw (2015), and follow their notation in the
119 discussion below. Consider a variable X , which is a function of G, S, and L. The total response of

120 X to changes in all of these variables can be written as

$$\delta X = X_{G2S2L2} - X_{G1S1L1}, \quad (1)$$

121 where the subscripts 1 and 2 refer to the control and perturbed states, respectively. The individual
 122 contributions of greenhouse gas forcing and cloud SW and LW effects can then be expressed as

$$\delta X_G = \frac{1}{2}[(X_{G2S1L1} - X_{G1S1L1}) + (X_{G2S2L2} - X_{G1S2L2})], \quad (2)$$

123

$$\delta X_S = \frac{1}{4}[(X_{G1S2L1} - X_{G1S1L1}) + (X_{G2S2L1} - X_{G2S1L1}) + (X_{G1S2L2} - X_{G1S1L2}) + (X_{G2S2L2} - X_{G2S1L2})], \quad (3)$$

124

$$\delta X_L = \frac{1}{4}[(X_{G1S1L2} - X_{G1S1L1}) + (X_{G2S1L2} - X_{G2S1L1}) + (X_{G1S2L2} - X_{G1S2L1}) + (X_{G2S2L2} - X_{G2S2L1})], \quad (4)$$

125 Equations 2–4 represent averages over the various pairs of experiments that involve changes in
 126 each of the three variables of interest. It can easily be shown that the right-hand sides of Eqs. 2–4
 127 add up to the right-hand side of Eq. 1, so that $\delta X = \delta X_G + \delta X_S + \delta X_L$ by construction. In the
 128 remainder of this paper, for additional clarity, the terms δX_G , δX_S , and δX_L are referred to as
 129 δX_{CO_2} , $\delta X_{SW\text{ cloud}}$, and $\delta X_{LW\text{ cloud}}$, respectively. We additionally define the change in X due to
 130 the net cloud-radiative change as the sum of the SW and LW effects, $\delta X_{\text{net cloud}} = \delta X_{SW\text{ cloud}} +$
 131 $\delta X_{LW\text{ cloud}}$.

132 It is important to note that the cloud and CO_2 responses in our experiments are affected by
 133 other feedbacks. In our model, this includes the temperature feedbacks (Planck and lapse rate), as
 134 well as the water vapor feedback; surface albedo values are kept constant between experiments.
 135 Unlike other studies (Langen et al. 2012; Mauritsen et al. 2013; Voigt and Shaw 2015), we do
 136 not separately account for the positive water vapor feedback, which likely amplifies the anomalies
 137 caused by the CO_2 and cloud perturbations in our experiments. Thus, the “effect of cloud-radiative
 138 changes” as defined in this paper encompasses all effects of replacing the clouds from the CTL

139 climate with 4xCO₂ clouds, including subsequent temperature and water vapor feedbacks. The
140 same applies to the component of the response that we ascribe to the CO₂ increase. This should
141 be kept in mind in the interpretation of our results, since the water vapor feedback in isolation
142 has been shown to have a non-negligible effect on the atmospheric circulation response (Voigt and
143 Shaw 2015).

144 **3. Results**

145 *a. Climate response to CO₂ and cloud changes*

146 We begin by describing the total response to CO₂ quadrupling, including the effects of cloud
147 feedbacks, in the experiment with locked clouds (left column of Fig. 1); this is equivalent to the
148 change described by Eq. 1. CO₂ quadrupling produces a large increase in sea surface temperature
149 (SST), with a global-mean increase of 4.4 K and amplified warming at high latitudes (Fig. 1a,
150 left). The surface warming is smallest near the edge of the tropics, so that the meridional SST
151 gradient increases within the tropics, but decreases in the extratropics. The vertical structure of the
152 temperature response (Fig. 1b) features the familiar maximum in the upper tropical troposphere (as
153 expected if the tropical troposphere remains close to neutral stability relative to the moist adiabat),
154 and stratospheric cooling, a direct consequence of the CO₂ increase. The temperature changes
155 result in a large zonal wind response (Fig. 1c) with a poleward shift of the tropospheric jet and
156 a vertical expansion of the upper-level westerlies. The upper tropical troposphere also features
157 a transition from easterly to superrotating winds at the Equator, a feature previously reported in
158 warmed aquaplanet climates (Caballero and Huber 2010), with westerly winds peaking near 8 m
159 s⁻¹ around 100 hPa. Finally, the response of the mean meridional circulation reflects the combined
160 effects of a Hadley cell weakening, and upward and poleward expansion of the circulation, all of

161 which are typical features of global warming experiments (e.g., Frierson et al. 2006; Lorenz and
162 DeWeaver 2007; Langen et al. 2012). Differences between hemispheres appear to be minimal,
163 suggesting that the responses are very robust and unaffected by sampling variability.

164 Before we study the individual effects of cloud feedbacks and CO₂ increase on the circulation
165 response, we need to ensure that the total response in the cloud-locked experiment is similar to
166 the response in the case with interactive clouds. As mentioned in the previous section, the mean
167 CTL and 4xCO₂ climates may be different owing to the decorrelation between cloud, temperature,
168 and moisture anomalies in the cloud-locked case. The differences in the responses to CO₂ quadru-
169 pling, shown in the right column of Fig. 1, are relatively small overall. The case with interactive
170 clouds has very slightly larger surface warming (0.05 K global-mean difference), with the largest
171 temperature differences in the stratosphere and in the subtropics of the Northern Hemisphere. (Re-
172 call that since the model is hemispherically and zonally symmetric, any differences between the
173 hemispheres are solely due to sampling error.) The slightly enhanced warming results in a modest
174 enhancement of the poleward shift of the eddy-driven jet, particularly in the Northern Hemisphere,
175 combined with a slight weakening of the subtropical jet core and an enhancement of the tropical
176 superrotation. Differences in the mean meridional circulation response appear to be very small.
177 We conclude that overall, the experiment with locked clouds provides a meaningful representation
178 of the total climate response to CO₂ quadrupling in our model.

179 *b. Surface temperature and cloud response*

180 We next consider the breakdown of the SST response into cloud and CO₂ effects (Fig. 2a).
181 Quadrupling CO₂ while holding the clouds fixed (Eq. 2) causes a global-mean SST increase of
182 3.4 K, with the temperature change smoothly increasing with latitude from the tropics to the poles
183 (green curve in Fig. 2a). As discussed in section 2, note that this response includes the effects

184 of the water vapor and lapse rate feedbacks. While the ice-albedo feedback is not active in our
185 simulations because of the absence of sea ice, amplified warming at high latitudes is still expected
186 for several reasons. Temperature (Planck and lapse-rate) feedbacks have been shown to drive polar
187 amplification in CMIP5 models (Pithan and Mauritsen 2014), although the lapse-rate feedback is
188 likely much weaker in our aquaplanet model given the lack of sea ice and associated low-level
189 temperature inversions. But more importantly, even in the absence of local positive feedbacks,
190 an increase in poleward energy transport by the atmosphere is to be expected in response to an
191 increasing meridional moist static energy (MSE) gradient with warming, yielding enhanced energy
192 convergence in polar regions (Hwang et al. 2011; Roe et al. 2015). The MSE gradient increase
193 results from the larger increase in specific humidity at low latitudes, consistent with the Clausius-
194 Clapeyron relationship under the assumption of near-constant relative humidity.

195 The SW cloud effect (Fig. 2a, purple curve; Eq. 3) causes a negligible change in global-mean
196 SST (-0.2 K), but features a strong latitude dependence, with a weak temperature increase in the
197 tropics and lower midlatitudes, and strong cooling at high latitudes. The temperature response is
198 in close agreement with the SW cloud feedback, shown in Fig. 2b (purple curve)¹. The negative
199 SW cloud feedback at high latitudes results from increases in cloud water and optical depth rather
200 than total cloud amount (Figs. 2c–d), consistent with most climate models (Zelinka et al. 2012,
201 Fig. 8b). The high-latitude cloud water increase is thought to be related to the effect of phase
202 changes in mixed-phase clouds: warming favors a transition from ice to liquid water, reducing
203 the overall precipitation efficiency and yielding an enhanced reservoir of cloud water (Senior and
204 Mitchell 1993; Tsushima et al. 2006; McCoy et al. 2014a; Ceppi et al. 2015). In addition to the

¹The SW and LW cloud feedbacks were calculated in separate partial radiative perturbation (PRP) experiments, where the difference in radiative fluxes between instantaneous CTL and 4xCO₂ clouds was calculated at each time step. The radiative effect of cloud changes is the average of two PRP experiments, one with control CO₂ and one with quadrupled atmospheric CO₂ concentration, equivalent to a two-sided PRP (Colman and McAvaney 1997; Soden et al. 2008).

205 phase change effect, changes in the vertical derivative of the moist adiabat could also favor an
206 increase in cloud water with warming, and this effect is most pronounced at lower temperatures
207 (e.g., Betts and Harshvardhan 1987; Tselioudis et al. 1992). The resulting high-latitude cloud
208 optical depth feedback is a very robust feature of global warming simulations in CMIP5 models
209 (Zelinka et al. 2012; McCoy et al. 2014b; Ceppi et al. 2015). Most climate models also predict
210 a positive SW cloud feedback in the tropics owing to cloud amount decreases (e.g., Bony and
211 Dufresne 2005; Zelinka et al. 2012), although our physical understanding of these changes is more
212 limited (Boucher et al. 2013). Thus, the overall structure of the SW cloud feedback in our model
213 is consistent with the mean behavior of climate models, even though the strongly negative high-
214 latitude feedback in our model causes a more negative global-mean SW cloud feedback compared
215 to most models (Soden et al. 2008; Zelinka et al. 2012; Vial et al. 2013). As will be shown later
216 in the paper, the increase in the meridional SST gradient caused by the SW cloud effect is a key
217 component of the total response to CO₂ increase.

218 The temperature response due to the LW cloud effect (orange curve in Fig. 2a; Eq. 4) mirrors
219 the response to the SW effect, so that both effects partly cancel each other out. The LW cloud
220 feedback largely reflects the high cloud amount response (Fig. 2b–c) and is positive in the global-
221 mean, consistent with the rise of cloud tops under the Fixed Anvil Temperature (FAT) hypothesis
222 (Hartmann and Larson 2002; Zelinka and Hartmann 2010). The high cloud decreases in parts
223 of the tropics are sufficiently large to offset the effect of rising cloud tops, yielding a negative
224 feedback locally. The particularly strong positive LW cloud feedback at high latitudes is associated
225 with very high cloud fraction in the control climate, especially at mid and upper levels (not shown),
226 yielding a strong LW effect of rising cloud tops. Despite the partial cancellation of SW and
227 LW cloud-radiative changes, the SST response to both cloud effects combined (grey curve) is
228 still dominated by the SW effect in terms of the meridional structure, with peak warming at the

229 equator and an overall increased Equator-to-pole temperature gradient, while the global-mean SST
230 increase results entirely from the LW effect of clouds.

231 *c. Atmospheric circulation changes*

232 We now study the vertical structure of changes in temperature and atmospheric circulation in
233 our experiments. We begin by considering the zonal wind response and its relationship with tem-
234 perature changes, shown in Fig. .3.

235 The CO₂ increase causes the expected tropospheric warming and stratospheric cooling, with
236 warming maxima at upper levels in the tropics and in the lower polar troposphere (Fig. 3, top
237 left). An interesting result is that increasing CO₂ while holding the clouds fixed causes very little
238 change in the tropospheric jet (Fig. 3, top right). This result is surprising, since a poleward shift
239 of the tropospheric eddy-driven jet is often regarded as one of the most fundamental circulation
240 responses to greenhouse gas forcing, especially in idealized models (Kushner et al. 2001; Yin
241 2005; Brayshaw et al. 2008; Lu et al. 2010). The zonal wind response mainly consists of an
242 upward shift of the jet stream, consistent with the troposphere becoming warmer and deeper. A
243 slight weakening of the tropospheric jet is seen on the equatorward flank of the jet at the lowest
244 levels, resulting in a poleward jet shift of 0.9° (based on the latitude of peak zonal-mean zonal wind
245 at the surface, cubically interpolated onto a 0.1° grid). The relatively modest poleward jet shift in
246 the troposphere appears consistent with the structure of the temperature response: while at upper
247 levels the warming peaks in the tropics, in the lower troposphere it maximizes at high latitudes, a
248 result consistent with previous modeling evidence (e.g., Held 1993). Upper-level tropical warming
249 and lower-level polar warming have been shown to have opposing influences on the eddy-driven
250 jet response (Butler et al. 2010).

251 By contrast, the relatively modest temperature response caused by the SW cloud feedback pro-
252 duces a substantial zonal wind response in the troposphere, with a clear strengthening and pole-
253 ward shift of the eddy-driven jet (Fig. 3, second row). As will be shown below, the large eddy-
254 driven jet response is related to the spatial structure of the thermal forcing associated with the SW
255 cloud feedback, which causes an enhancement of the meridional temperature gradient at all levels
256 in the troposphere. The fact that an increased midlatitude temperature gradient tends to favor a
257 poleward jet shift has been noted in several previous studies (Brayshaw et al. 2008; Chen et al.
258 2010; Ceppi et al. 2012; Lorenz 2014). While the mechanisms of the eddy-driven jet response
259 to thermal forcing are still a topic of active research, our results appear consistent with several
260 existing theories. Lorenz (2014) proposed that stronger upper-level westerlies near the jet result
261 in changes in Rossby wave propagation, favoring a poleward shift of the region of eddy momen-
262 tum flux convergence. Chen and Held (2007) argued that increasing eddy phase speeds could
263 cause a poleward shift of the eddy-driven circulation; an eddy phase speed increase could occur in
264 response to a strengthened meridional temperature gradient and upper-level westerly winds. Be-
265 sides the poleward jet shift, we also note a transition to more westerly winds in the upper tropical
266 troposphere, which are sustained by enhanced eddy momentum flux convergence associated with
267 tropical waves (not shown).

268 Unlike the effect of SW cloud-radiative changes, the LW effect yields a tropospheric temperature
269 response qualitatively similar to that of CO₂, but weaker overall and with a higher degree of polar
270 amplification at low levels (Fig. 3, third row). Like CO₂, this forcing also mainly causes an upward
271 shift of the jet streams, with a relatively weak tropospheric response that occurs mostly above 500
272 hPa and resembles a narrowing of the westerly jet.

273 Adding the SW and LW cloud responses together yields the net effect of cloud-radiative changes
274 (fourth row of Fig. 3), consisting of generalized tropospheric warming peaking in the tropical

275 upper troposphere. It is noteworthy that the net cloud effect results in a warming pattern quite
276 different from CO₂ forcing, with an increase in Equator-to-pole temperature gradient at all tropo-
277 spheric levels. The temperature change due to clouds yields a clear poleward and upward shift
278 of the tropospheric jet. Finally, the total response to CO₂ quadrupling, including the effects of
279 cloud changes, is shown in the bottom row of Fig. 3; recall that this response is identical to the
280 sum of rows 1–3, by construction. The tropospheric zonal wind response most resembles the ef-
281 fect of clouds (compare rows 4 and 5). The large contribution of cloud-radiative changes to the
282 tropospheric circulation response will be confirmed later in this paper, using various metrics to
283 objectively quantify circulation shifts.

284 The very distinct effects of cloud-radiative changes and CO₂ forcing on the thermal structure of
285 the troposphere are summarized in Fig. 4. To quantify the overall change in tropospheric thermal
286 structure at various levels, we define the mean upper- and lower-tropospheric temperature as the
287 vertically-averaged values from 100 to 500 hPa and 500 to 1000 hPa, respectively, which we
288 denote as $\langle T \rangle_{\text{upper}}$ and $\langle T \rangle_{\text{lower}}$ (Fig. 4a,c). In the upper troposphere, both clouds and CO₂ forcing
289 cause enhanced tropical warming, yielding an enhanced thermal gradient between the tropics (30°
290 S to 30° N) and the extratropics (Fig. 4a,b). Both the SW and LW cloud changes contribute
291 to the enhanced upper-tropospheric temperature gradient, even though the LW effect is almost
292 twice as large. In the lower troposphere, however, only the SW cloud-radiative changes act to
293 enhance the meridional temperature gradient, while both the LW cloud effect and CO₂ forcing
294 cause polar-amplified warming (Fig. 4c,d). Thus, *in a tropospheric-mean sense* the SW cloud-
295 radiative change is the main contributor to the amplified temperature gradient; while CO₂ forcing
296 and LW cloud-radiative changes yield substantial warming, they cause negligible change in the
297 gradient of tropospheric temperature in the vertical mean (Fig. 4e,f). This result strongly suggests
298 that the change in temperature *gradient* at all tropospheric levels is much more relevant to the

299 atmospheric circulation response than the change in mean temperature, at least in terms of the
300 poleward expansion of the circulation.

301 We next assess changes in eddy activity, measured by the eddy kinetic energy as $EKE = (\overline{u'^2} +$
302 $\overline{v'^2})/2$, where primes denote deviations from the zonal and time mean, and overbars indicate zonal
303 and time averages (left column of Fig. 5). Around the midlatitudes, EKE provides a measure
304 of the location and intensity of the storm track, which modulates important climate properties in
305 the extratropics such as cloudiness and precipitation. Comparing with the temperature changes in
306 Fig. 3, we find that the tropospheric EKE response is strongly tied to changes in the meridional
307 temperature gradient, consistent with the idea that baroclinicity is the dominant control on eddy
308 activity. The largest tropospheric response is an increase and poleward shift of EKE caused by
309 the SW cloud feedback in midlatitudes, but it is opposed by weaker EKE decreases by the LW
310 cloud feedback and CO₂ forcing with clouds fixed, resulting in a near-zero total response below
311 200 hPa (Fig. 5, bottom left). The total EKE response mainly consists of an upward expansion
312 in midlatitudes (consistent with the deepening of the troposphere with warming), as well as a
313 strengthening of eddy activity around the equatorial tropopause, which results mainly from the
314 CO₂ and SW cloud effects.

315 Finally, we discuss the response of the meridional mass streamfunction (calculated as $\Psi =$
316 $2\pi a g^{-1} \int_0^{p_0} \bar{v} \cos \phi dp$, where a is the radius of the Earth, g is gravitational acceleration, \bar{v} is zonal-
317 mean meridional wind, ϕ is latitude, p is pressure, and p_0 is surface pressure). The mass stream-
318 function reflects the Hadley circulation climatology, which is an important control on the moisture
319 budget in the intertropical convergence zone (ITCZ) and in subtropical dry regions (Hartmann
320 1994). Overall the mass streamfunction response consists of a weakening of the Hadley circu-
321 lation, except in for the response to SW cloud-radiative changes (right column of Fig. 5). The
322 Hadley cell response to various forcings appears consistent with the competing effects of increas-

323 ing meridional SST gradient and increasing static stability. While the SW effect tends to enhance
324 the meridional SST gradient within the tropics, favoring a strengthening of the circulation, cloud
325 changes also yield a stabilization of the tropics, especially through the LW effect, which favors a
326 Hadley cell weakening (Knutson and Manabe 1995; Gastineau et al. 2008). This results in a very
327 small overall change in Hadley cell strength in response to the net cloud-radiative changes. In the
328 case of CO₂ quadrupling with fixed clouds, tropical SST gradients change little (Fig. 2a) and the
329 stability increase dominates, resulting in a marked weakening of the Hadley circulation.

330 A modest poleward expansion of the Hadley cell edge also occurs in response to each of the
331 forcings; while this response is too weak to be visible in the responses to individual forcings, it
332 appears clearly in the total streamfunction response (Fig. 5, bottom right). The poleward shift of
333 the Hadley cell edge may result from the combined influences of the stabilization of the tropical
334 troposphere, which shifts the latitudes of baroclinic instability poleward (Frierson et al. 2007; Lu
335 et al. 2007), and from changes in the wave driving of the circulation. For example, increases in
336 Rossby wave phase speeds with global warming (Chen and Held 2007) could cause a poleward
337 shift of eddy momentum flux divergence and associated subtropical wave breaking, driving an
338 anomalous meridional circulation consistent with a Hadley cell expansion (Ceppi and Hartmann
339 2013; Vallis et al. 2014). The Hadley cell weakening and poleward expansion are robust features
340 of the atmospheric circulation response to warming (Frierson et al. 2007; Lu et al. 2007; Gastineau
341 et al. 2008; Ceppi and Hartmann 2013; Vallis et al. 2014).

342 *d. Poleward expansion of the atmospheric circulation*

343 We have shown that cloud feedbacks with global warming produce thermal forcings that are
344 particularly effective at inducing a poleward expansion of the tropospheric circulation in our aqua-
345 planet model, particularly through the impact of SW cloud-radiative changes on meridional tem-

346 perature gradients. To objectively quantify the contribution of clouds to the expansion of the
347 circulation, we calculate changes in four circulation metrics: the poleward edge of the Hadley
348 circulation based on the meridional mass streamfunction at 500 hPa; the edge of the subtropi-
349 cal dry zones, calculated as the latitude where precipitation equals evaporation in the subtropics
350 ($P - E = 0$); the jet latitude measured as the peak surface zonal-mean zonal wind; and the lat-
351 itude of the storm tracks, measured as the peak in sea-level pressure (SLP) variance. For each
352 of these metrics, the fields of interest are cubically interpolated onto a 0.1° grid before locating
353 the latitudes. For storm-track latitude, we use SLP variance rather than EKE for consistency with
354 previous studies (e.g., Chang et al. 2012; Harvey et al. 2014); however, note that the results are
355 similar if surface EKE is used instead. As in Harvey et al. (2014), we use 2–6 day band-pass
356 filtered SLP data to quantify the variability associated with transient synoptic eddies.

357 The changes in each of the metrics relative to the control climate are shown in Fig. 6. Both clouds
358 and CO_2 forcing alone contribute to the expansion of the tropics, as measured by the edge of the
359 Hadley cells and of the subtropical dry zones. However, their impacts on the jet and storm-track
360 position are very different, with SW cloud-radiative changes having the largest positive effect. The
361 strong SW cloud effect on jet and storm-track latitude is consistent with the zonal wind and EKE
362 responses shown in Figs. 3 and 5. It is noteworthy that the storm-track latitude is much more sen-
363 sitive to SW and LW cloud effects than is the jet position; this may be related to the much higher
364 climatological latitude of the storm track compared to the jet, as defined here (52.4° versus 38.9°),
365 making the storm track more responsive to high-latitude temperature changes. Remarkably, in
366 our model the SW radiative response associated with clouds is the only factor contributing to the
367 poleward shift of the storm track. The net effect of cloud feedbacks is to force a poleward expan-
368 sion of the circulation that strongly enhances the effect of CO_2 forcing, while the CO_2 increase
369 only yields only a modest circulation shift if the clouds are held fixed. This result becomes clear

370 by comparing the grey and black crosses in Fig. 6, which show that the cloud-radiative changes
371 explain more than half of the total expansion of the circulation.

372 As described in Eqs. 2–4, the responses to each of the forcings result from averages over several
373 experiments. Comparing the responses to a particular forcing across experiments provides a mea-
374 sure of the sensitivity of the response to the reference climate. The shifts in each of the circulation
375 metrics shown in Fig. 6 are listed in Table 1 for all experiments. For each individual forcing, there
376 are clear differences in the magnitude of the shift in each of the metrics between experiments.
377 Part of these differences may result from random internal variability, but we believe most of the
378 differences reflect a sensitivity to the initial climate. Despite this nonlinear behavior, the effect of
379 each forcing on atmospheric circulation remains qualitatively consistent across experiments. For
380 example, for each metric and forcing, the sign of the shift is identical across all experiments; the
381 only exception is the eddy-driven jet response to LW cloud changes, which is generally close to
382 zero.

383 **4. Discussion**

384 The main purpose of this paper is to show that cloud feedbacks produce thermal forcings which
385 can substantially alter the large-scale circulation response to CO₂ increase. Our results support
386 the finding of Ceppi et al. (2014), of a strong relationship between the meridional structure of SW
387 feedbacks and the austral jet stream response in CMIP5 models under RCP8.5 forcing. They are
388 also consistent with the large effect of clouds on the mean circulation shown by Li et al. (2015).
389 Recently, Voigt and Shaw (2015) demonstrated the importance of cloud and water vapor feedbacks
390 on the circulation response in two aquaplanet models forced with a uniform SST increase. Because
391 the SSTs are prescribed, however, it is likely that their results mainly reflect the effect of LW cloud
392 feedbacks, since SW radiation is mostly absorbed at the surface. A novel aspect of our study is

393 the separate consideration of SW and LW cloud feedbacks, which highlights the important but
394 different roles of SW and LW cloud effects when SSTs are allowed to interact with radiation.

395 *a. Cloud feedbacks in contemporary climate models*

396 Care must be taken in generalizing our results to other models, for at least two reasons. First
397 and foremost, cloud feedbacks are highly uncertain and model-dependent, and so is their effect
398 on atmospheric circulation. To quantify their contribution to the mean and spread in atmospheric
399 circulation changes with warming, it is therefore necessary to test the effects of cloud changes in
400 a wider set of models. Despite this uncertainty, we will argue below that the meridional structures
401 of the SW and LW cloud feedbacks produced by our model are fairly representative of the mean
402 behavior of state-of-the-art climate models. Second, our experiment design is highly idealized.
403 The low surface albedo associated with the aquaplanet configuration may lead to an overestima-
404 tion of the SW effect of clouds, particularly compared with Northern Hemisphere conditions. The
405 sensitivity of the atmospheric circulation to external forcings may be overestimated given the low
406 climatological jet latitude in our model (38.9°), especially compared to the Southern Hemisphere
407 (Kidston and Gerber 2010). Also, the zonally symmetric boundary conditions mean that station-
408 ary waves play no role in the atmospheric circulation response to CO_2 forcing, unlike the real
409 world (Simpson et al. 2014). However, the idealized experimental design also allows for an easier
410 interpretation of the basic effects of cloud feedbacks on circulation.

411 Cloud feedbacks play a special role in the atmospheric circulation response to warming for
412 two reasons: (a) they tend to enhance the Equator-to-pole temperature gradient and midlatitude
413 baroclinicity, and (b) they are highly uncertain and cause inter-model spread in circulation changes.
414 Figure 7, showing the cloud feedback components in the abrupt4xCO2 simulations of 28 CMIP5
415 models, illustrates these two points. As in our idealized model, the mean SW cloud feedback in

416 CMIP5 models leads to an overall enhanced meridional gradient of absorbed SW radiation around
417 the midlatitudes, with a positive mean feedback in the tropics and a negative feedback at high
418 latitudes. By contrast, the LW cloud feedback tends to be positive at all latitudes. Because the LW
419 cloud feedback has less spatial structure than the SW feedback, the net feedback is dominated by
420 the SW component (Fig. 7c), tending to enhance the meridional gradient of absorbed SW radiation;
421 this is also in agreement with our model results (see Fig. 2b). Comparing the grey curves in Fig. 7
422 provides an idea of the uncertainty in the magnitude and spatial distribution of the cloud feedbacks,
423 which is particularly large for the SW component.

424 *b. Relationship between feedback and temperature response*

425 Inter-model differences in cloud feedbacks motivate a discussion of the relationship between the
426 meridional structure of the feedbacks and the structure of the resulting temperature response. It
427 is important to recognize that changes in top-of-atmosphere radiation associated with feedbacks
428 do not necessarily predict the meridional structure of the associated temperature change, owing to
429 the role of meridional energy transport (Langen et al. 2012; Merlis 2014), consistent with climate
430 feedbacks being fundamentally nonlocal in nature (Feldl and Roe 2013). With this complication
431 in mind, how robust are our results to variations in the spatial pattern of the SW and LW cloud
432 feedbacks?

433 The strong poleward circulation shift induced by the SW cloud feedback relies on an overall
434 enhancement of the tropospheric meridional temperature gradient. If the tropical SW cloud feed-
435 back is positive as most models predict, the resulting increase in tropical MSE will induce an
436 enhancement of the poleward energy transport by the atmosphere, causing polar-amplified warm-
437 ing unless the high-latitude SW cloud feedback is sufficiently negative. In other words, the SW
438 cloud feedback could produce polar amplification at low levels even if the Equator-to-pole gradi-

439 ent of absorbed SW radiation is enhanced. The remote effects of tropical climate feedbacks on
440 the high-latitude temperature response are clearly illustrated in Fig. 2 of Roe et al. (2015). The
441 circulation impacts of the SW cloud feedback would likely also depend on the degree of tropical
442 upper-tropospheric warming, which we expect to be directly linked to the amount of tropical SST
443 increase caused by SW cloud-radiative changes, since surface and upper-tropospheric tempera-
444 tures are tightly coupled in the tropics through the effects of convection.

445 Thus, the net effect of the SW cloud feedback on circulation is determined by the relative mag-
446 nitudes of the positive tropical forcing and negative high-latitude forcing; for example, we would
447 expect to find a much weaker poleward expansion of the circulation by the SW cloud feedback in a
448 model in which this feedback is much less negative at high latitudes. While the negative SW cloud
449 feedback at high latitudes is a robust feature of CMIP5 global warming experiments (Fig. 7a), and
450 is supported by a robust physical mechanism (phase changes in mixed-phase clouds, section 3b),
451 the magnitude of this negative high-latitude feedback – both in absolute terms and relative to the
452 generally positive SW cloud feedback in the tropics – is highly model-dependent.

453 We believe the temperature and circulation impacts of the LW cloud feedback are somewhat
454 more robust. In presence of a positive LW cloud feedback at most latitudes, the low-level temper-
455 ature response to LW cloud-radiative changes is very likely to be amplified at high latitudes owing
456 to the effect of increasing meridional energy transport and positive temperature feedbacks (Pithan
457 and Mauritsen 2014; Roe et al. 2015). An overall positive LW cloud feedback is expected as cloud
458 tops rise with warming, consistent with the Fixed Anvil Temperature hypothesis (Hartmann and
459 Larson 2002); models agree on this effect, and there is no physical argument to expect a negative
460 LW cloud feedback at high latitudes. However, the degree of polar amplification at low levels will
461 still be affected by the magnitude of the local LW cloud feedback. In our model, the high-latitude
462 LW cloud feedback appears too positive, which we ascribe to an unrealistically high climatolog-

463 ical cloud fraction in our aquaplanet configuration in high latitudes (section 3b). It is therefore
464 possible that our model overestimates the amount of polar amplification associated with the LW
465 cloud feedback, and therefore underestimates the contribution of LW cloud-radiative changes on
466 the poleward expansion of the circulation, compared to more realistic models.

467 Despite the complex relationship between feedback patterns and temperature responses, Ceppi
468 et al. (2014) showed that the meridional structure of SW feedbacks (mainly from clouds and sea
469 ice) explains the changes in SST gradient very well in RCP8.5 simulations around the Southern
470 midlatitudes. From the perspective of the atmospheric circulation response, the results in the
471 present paper suggest that the spatial distribution of the thermal forcing, both at lower and upper
472 tropospheric levels, is more important than the global-mean effect, as discussed in section 4a.
473 Hence, the results in Fig. 7 support the idea that the cloud feedback likely enhances the poleward
474 expansion of atmospheric circulation in most climate models.

475 *c. Effects of other climate feedbacks*

476 While the focus of this paper has been on the effects of clouds, other feedbacks will also affect
477 the temperature and circulation responses to greenhouse gas forcing in climate models. For exam-
478 ple, the large-scale effects of the water vapor feedback have been demonstrated in previous studies
479 (Schneider et al. 1999; Hall and Manabe 1999; Mauritsen et al. 2013; Voigt and Shaw 2015). Al-
480 though Voigt and Shaw (2015) found an equatorward contraction of the atmospheric circulation
481 in response to radiative changes of water vapor, it is not obvious that a similar response would
482 be obtained in a coupled atmosphere-ocean climate model like ours. This is because water vapor
483 changes cause a very different temperature response when SSTs are allowed to respond to the ra-
484 diative forcing, with substantial warming in the tropical upper troposphere (compare e.g. Fig. 6d
485 in Langen et al. 2012 with Fig. 3c in Voigt and Shaw 2015). Furthermore, since the water vapor

486 content is so strongly tied to temperature through the Clausius-Clapeyron relationship, we spec-
487 ulate that the uncertainty in the circulation response associated with the water vapor feedback is
488 much smaller than that caused by cloud changes.

489 By contrast, we believe that the temperature and surface albedo feedbacks could contribute sig-
490 nificant uncertainty to the spatial pattern of the temperature increase and the associated circulation
491 response in climate models. Temperature feedbacks (including the Planck and lapse rate feed-
492 backs) have been shown to contribute to polar warming (Pithan and Mauritsen 2014). The lapse
493 rate feedback, which is the strongest contribution to Arctic warming in CMIP5 models (Pithan and
494 Mauritsen 2014), is positive at high latitudes because of the existence of strong low-level inver-
495 sions that trap warming near the surface. It is therefore plausible that the lapse-rate feedback in
496 high latitudes could depend on the strength of the polar low-level inversions in the control climate.
497 Finally, the surface albedo feedback is dominated by fairly uncertain changes in sea ice extent and
498 snow cover, and while its effect on global-mean temperature is much smaller than that of cloud
499 feedbacks (Vial et al. 2013), it has a strong effect on polar amplification in CMIP5 models (Pithan
500 and Mauritsen 2014).

501 **5. Summary and conclusions**

502 This paper investigates the effect of cloud feedbacks on the atmospheric circulation response
503 to CO₂ quadrupling in an aquaplanet model with a slab-ocean lower boundary. We use a cloud-
504 locking technique to break down the circulation response into two main components: the response
505 to CO₂ increase while clouds are fixed, and the response to cloud changes while CO₂ is fixed. The
506 response to cloud changes is further decomposed into SW and LW cloud effects. We find that cloud
507 changes cause a very substantial atmospheric circulation response, inducing a poleward expansion
508 of the Hadley cells, midlatitude jet streams, and storm tracks. This response is dominated by the

509 SW effect of clouds, while LW cloud-radiative changes alone force a modest tropical expansion,
510 no jet shift, and an equatorward shift of the storm tracks.

511 While quadrupling CO₂ with fixed clouds also forces an expansion of the circulation, this effect
512 is smaller than the net effect of cloud changes, despite the fact that CO₂ quadrupling causes three
513 times as much surface warming than cloud changes in the global mean (3.4 versus 1.1 K). We
514 explain this surprising result in terms of the spatial structures of the thermal forcings associated
515 with CO₂ and cloud-radiative changes. The SW effect of cloud changes is to strongly enhance the
516 Equator-to-pole temperature gradient at all tropospheric levels, increasing midlatitude baroclinic-
517 ity. Previous research has associated this type of forcing with a clear strengthening and poleward
518 shift of the jet streams and storm tracks. By contrast, the CO₂ increase (and to a lesser extent the
519 LW cloud-radiative changes) cause global warming with peak warming in low-level polar regions
520 and in the upper tropical troposphere. We believe that the different changes in meridional tempera-
521 ture gradient at upper and lower levels have opposing effects on atmospheric circulation, reducing
522 the impact of these forcings on the expansion of the circulation.

523 Our results highlight the importance of the spatial structure of the temperature response as
524 opposed to the global-mean response, since the SW cloud-radiative changes cause the smallest
525 global-mean surface temperature change (-0.2 K), but the largest midlatitude circulation response
526 in our model. Thus, it is important to note that clouds could enhance the atmospheric circulation
527 response to CO₂ forcing even in a hypothetical case where the global-mean cloud feedback is
528 near-zero or negative. This suggests that in terms of large-scale circulation impacts, changes
529 in meridional temperature gradients may be at least as important as the amount of global-mean
530 warming.

531 We caution that the results presented in this paper are based on a single model, and are not neces-
532 sarily representative of the atmospheric circulation impacts of cloud feedbacks in other models or

533 in the real world. However, an analysis of the cloud feedbacks in CMIP5 model experiments with
534 quadrupled CO₂ concentrations reveals that the key basic features of the cloud-radiative response
535 are similar to our model – particularly the tendency of cloud feedbacks to enhance the Equator-
536 to-pole temperature gradient through the SW effect. We therefore argue that cloud changes likely
537 enhance the poleward expansion of the circulation with global warming in most state-of-the-art
538 climate models. Because of the large uncertainty in the cloud response, it is also likely that clouds
539 significantly contribute to inter-model differences in the atmospheric circulation response, as sug-
540 gested by previous research (Ceppi et al. 2014; Voigt and Shaw 2015).

541 This study has focused on the atmospheric circulation response mainly from the perspective of
542 the poleward expansion of the Hadley cells, jet streams, and storm tracks, in an idealized, zonally-
543 and hemispherically-symmetric setting. In a more realistic configuration, cloud feedbacks would
544 likely also have an important effect on the asymmetric component of the circulation, impacting
545 the amplitude and location of stationary waves (Donner and Kuo 1984; Slingo and Slingo 1988)
546 as well as inter-hemispheric asymmetries and the latitude of the intertropical convergence zone
547 (Frierson and Hwang 2012). This further underlines the fact that constraining cloud feedbacks is
548 essential not only for an accurate estimation of climate sensitivity, but also for a realistic represen-
549 tation of the atmospheric circulation response to greenhouse gas forcing.

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780 **LIST OF TABLES**

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784 text. The symbols used for the experiments are described in section 2. The
785 mean CO₂, SW cloud, and LW cloud effects are calculated as in Eqs. 2–4. . . . 38

786 TABLE 1. Hemispherically-averaged latitudinal shift in various atmospheric circulation metrics, with pole-
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788 of the metrics, see Fig. 6 and text. The symbols used for the experiments are described in section 2. The mean
789 CO₂, SW cloud, and LW cloud effects are calculated as in Eqs. 2–4.

Experiment	$\Psi_{500} = 0$	$P - E = 0$	ϕ_{jet}	$\phi_{\sigma^2(\text{SLP})}$
CTL	26.7	35.2	38.9	52.4
G2S1L1 – G1S1L1	0.8	1.1	1.4	0.1
G2S2L2 – G1S2L2	0.5	0.9	0.5	0.1
mean CO ₂ effect	0.6	1.0	0.9	0.1
G1S2L1 – G1S1L1	0.9	1.5	2.5	1.9
G1S2L2 – G1S1L2	0.2	0.8	1.0	3.1
G2S2L1 – G2S1L1	0.6	1.2	1.9	4.6
G2S2L1 – G2S1L2	0.2	1.0	1.1	1.4
mean SW cloud effect	0.5	1.1	1.6	2.7
G1S1L2 – G1S1L1	0.8	0.8	0.9	-2.4
G1S2L2 – G1S2L1	0.1	0.3	-0.2	-1.2
G2S1L2 – G2S1L1	0.5	0.4	0.1	-2.0
G2S2L2 – G2S2L1	0.1	0.3	-0.3	-1.5
mean LW cloud effect	0.4	0.4	0.1	-1.8

790 **LIST OF FIGURES**

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792 meridional mass streamfunction after CO₂ quadrupling. The left column shows the changes
793 between the CTL and 4xCO₂ experiments, with clouds locked to CTL and 4xCO₂ climates,
794 respectively (Eq. 1). The right column shows the difference between the response in cases
795 with interactive and locked clouds. In panel (d), 1 Sv = 10⁹ kg s⁻¹. 40

796 **Fig. 2.** (a) SST response broken down into effects of SW and LW cloud-radiative changes and CO₂
797 forcing. (b) SW and LW cloud feedback. (c) High ($p < 440$ hPa), low ($p > 680$ hPa), and
798 total cloud amount response. (d) Liquid and ice water path response. The cloud feedback
799 in (b) is normalized by the total global-mean surface warming in the 4xCO₂ experiment
800 including cloud changes (4.4 K). 41

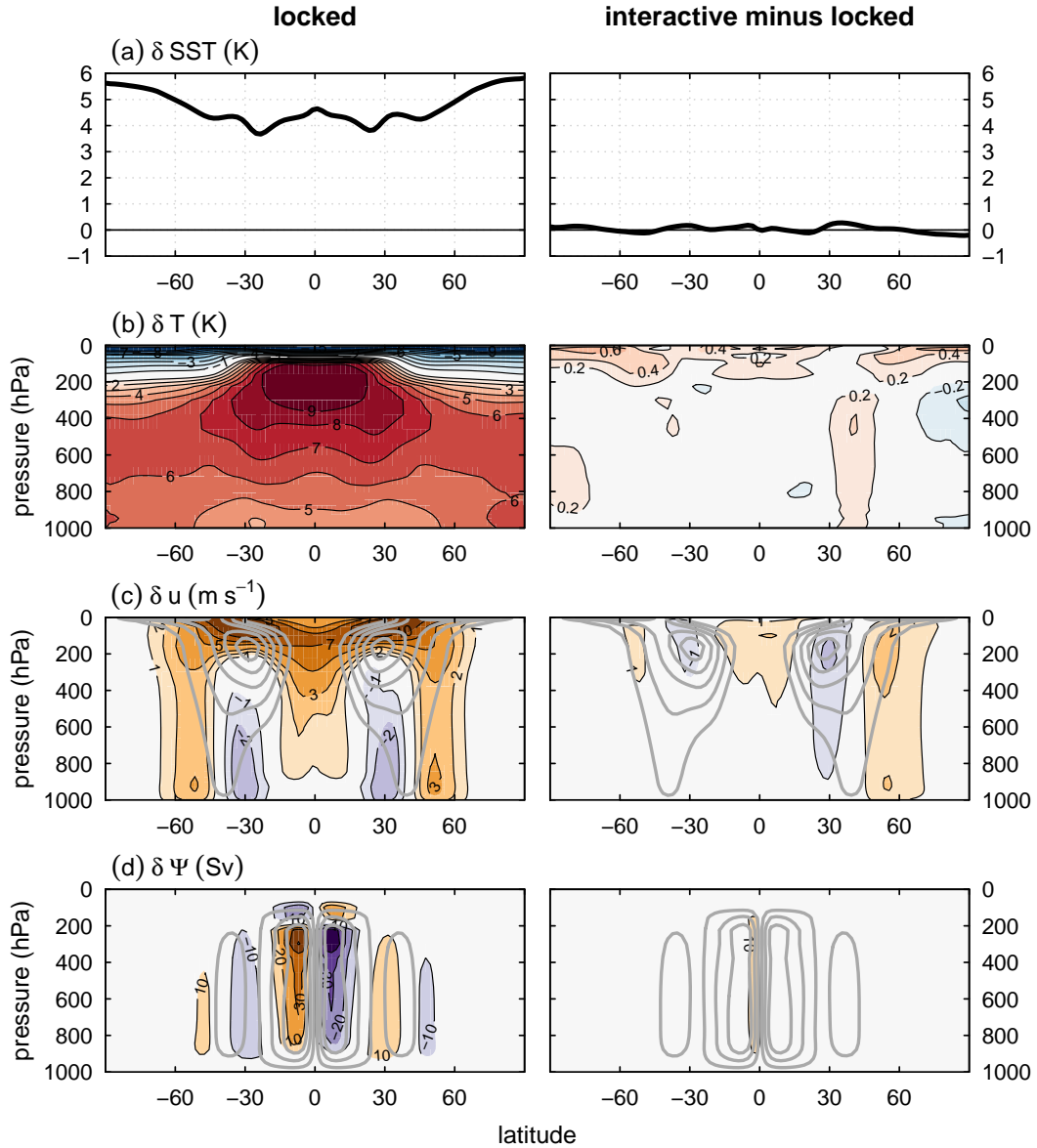
801 **Fig. 3.** Temperature (left column) and zonal wind (right column) responses to CO₂ quadrupling,
802 broken down into contributions from CO₂ forcing and clouds. Shading denotes the response.
803 In the right column, thick grey contours represent the control climatology (contour interval
804 10 m s⁻¹, only positive values shown). In panel (a), the CO₂, SW cloud, and LW cloud
805 effects are calculated with Eqs. 2, 3, and 4, respectively. 42

806 **Fig. 4.** (a,c,e) As in Fig. 2a, but showing the response of the vertically-averaged tropospheric tem-
807 perature. $\langle T \rangle_{\text{upper}}$, $\langle T \rangle_{\text{lower}}$, and $\langle T \rangle$ denote upper-tropospheric (100 to 500 hPa), lower-
808 tropospheric (500 to 1000 hPa), and tropospheric (100 to 1000 hPa) vertical-mean tempera-
809 ture, respectively. (b,d,f) Changes in the meridional gradient of $\langle T \rangle$ at various tropospheric
810 levels, calculated as the change in tropical-mean $\langle T \rangle$ (30° S to 30° N) minus the change in
811 extratropical-mean $\langle T \rangle$ 43

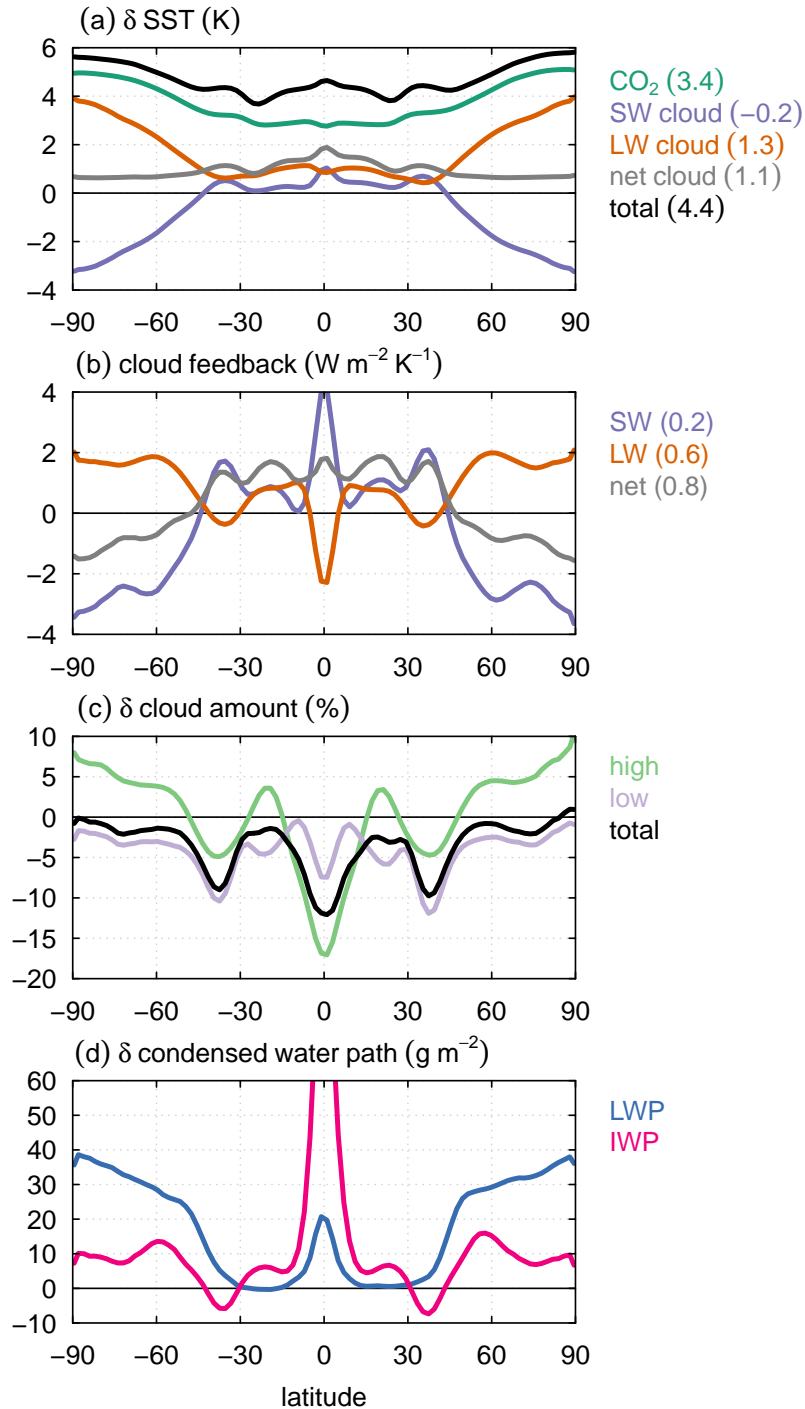
812 **Fig. 5.** As in Fig. 3, but for eddy kinetic energy (EKE, left) and meridional mass streamfunction
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818 mean zonal wind (ϕ_{jet}); storm-track latitude defined as the peak in sea-level pressure variance
819 ($\phi_{\sigma^2(\text{SLP})}$). All results are averaged over both hemispheres. 45

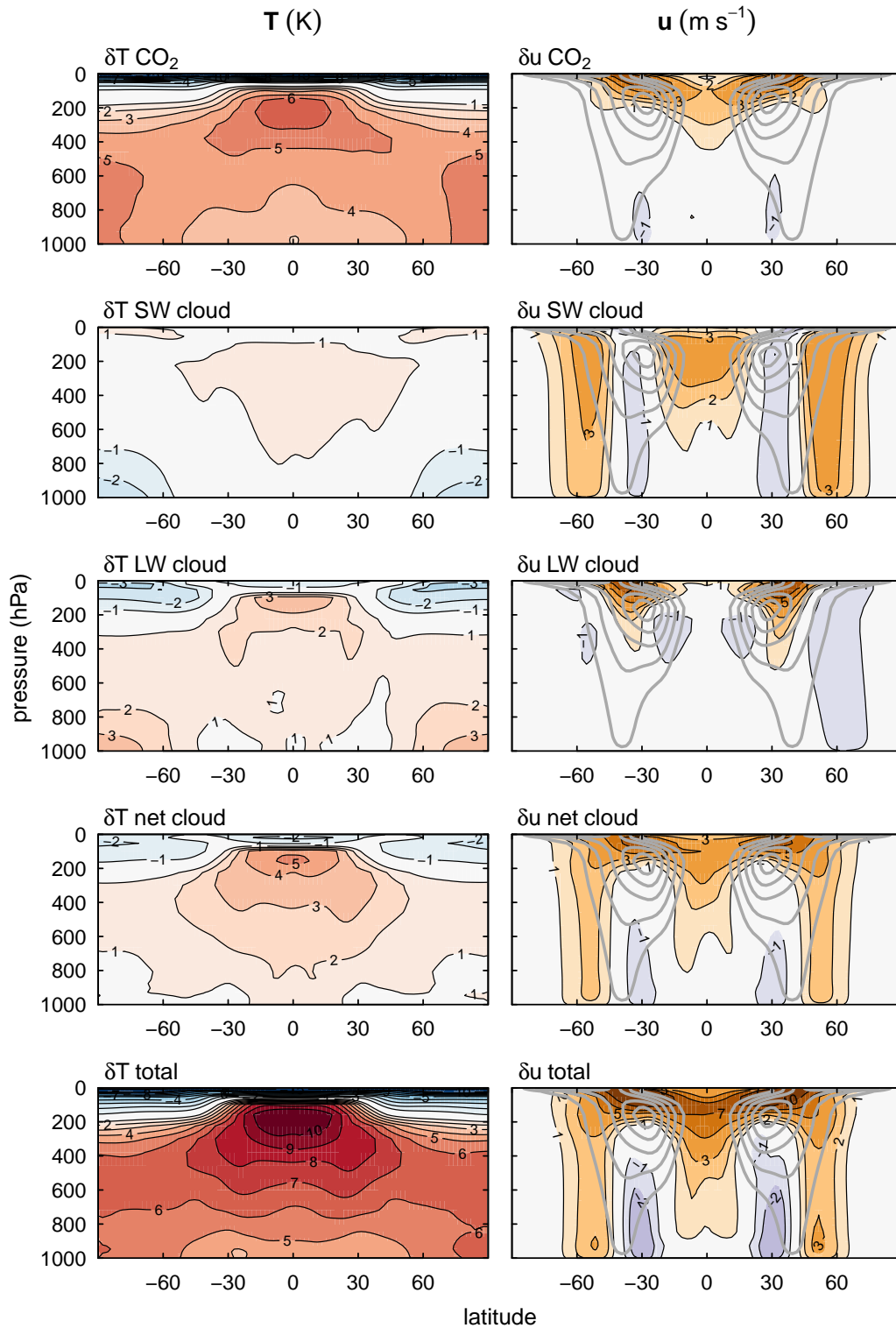
820 **Fig. 7.** Cloud feedback components in the abrupt4xCO₂ experiment of 28 CMIP5 models, all calcu-
821 lated as years 121–140 minus the pre-industrial control climatology. Grey curves represent
822 individual models, with the multi-model mean in thick black. The cloud feedback is cal-
823 culated using radiative kernels, following the method of Soden et al. (2008), and includes
824 rapid adjustments to CO₂ forcing (Sherwood et al. 2015). 46



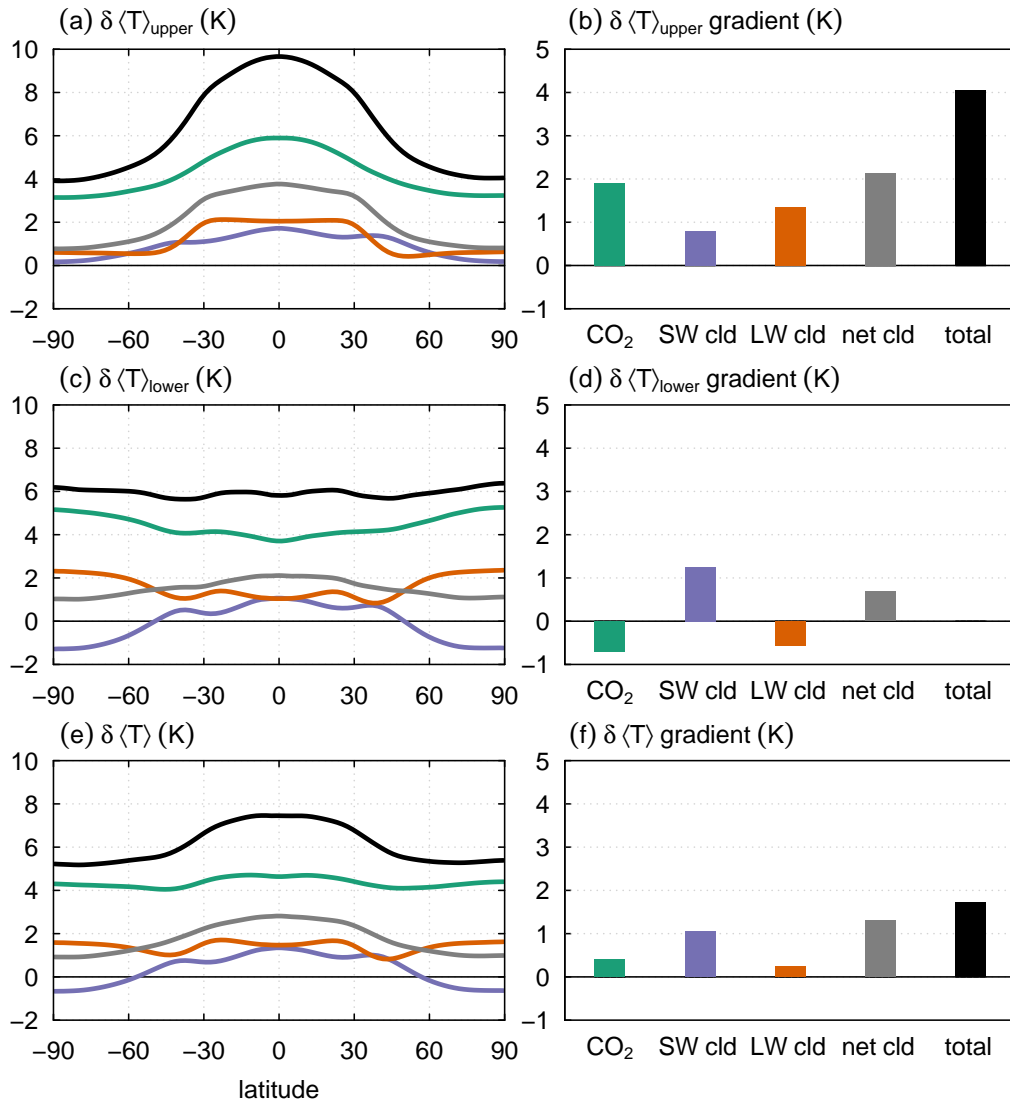
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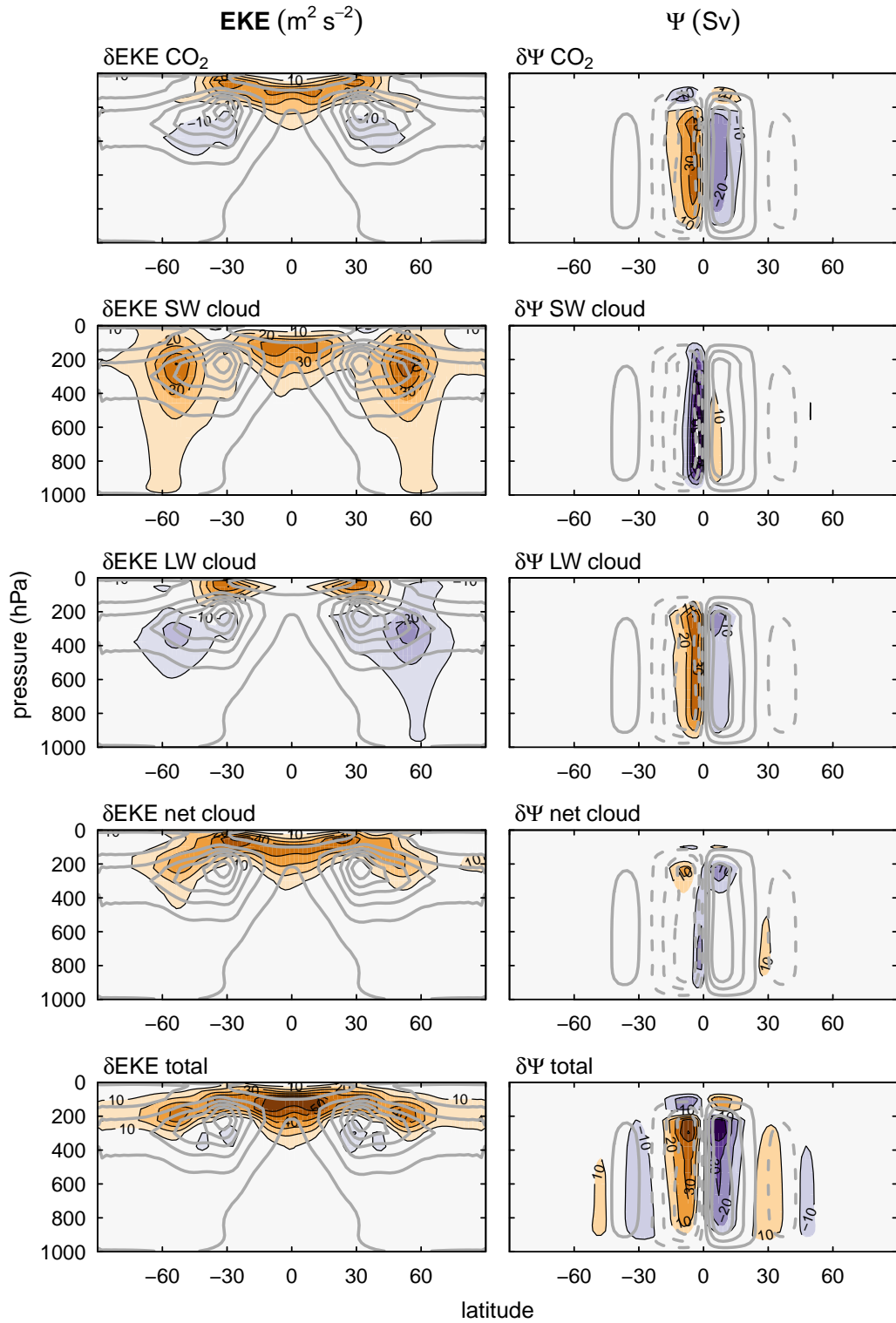
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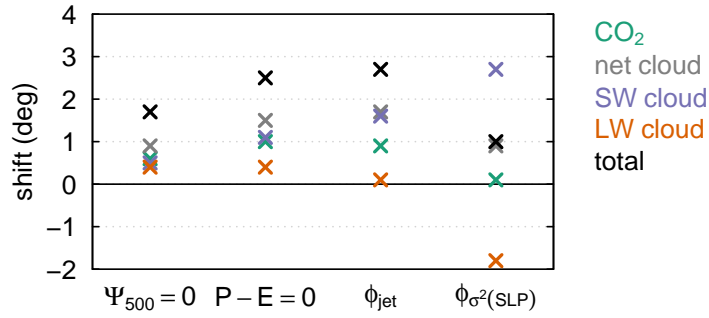
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 835 contours represent the control climatology (contour interval 10 m s^{-1} , only positive values shown). In panel (a),
 836 the CO_2 , SW cloud, and LW cloud effects are calculated with Eqs. 2, 3, and 4, respectively.



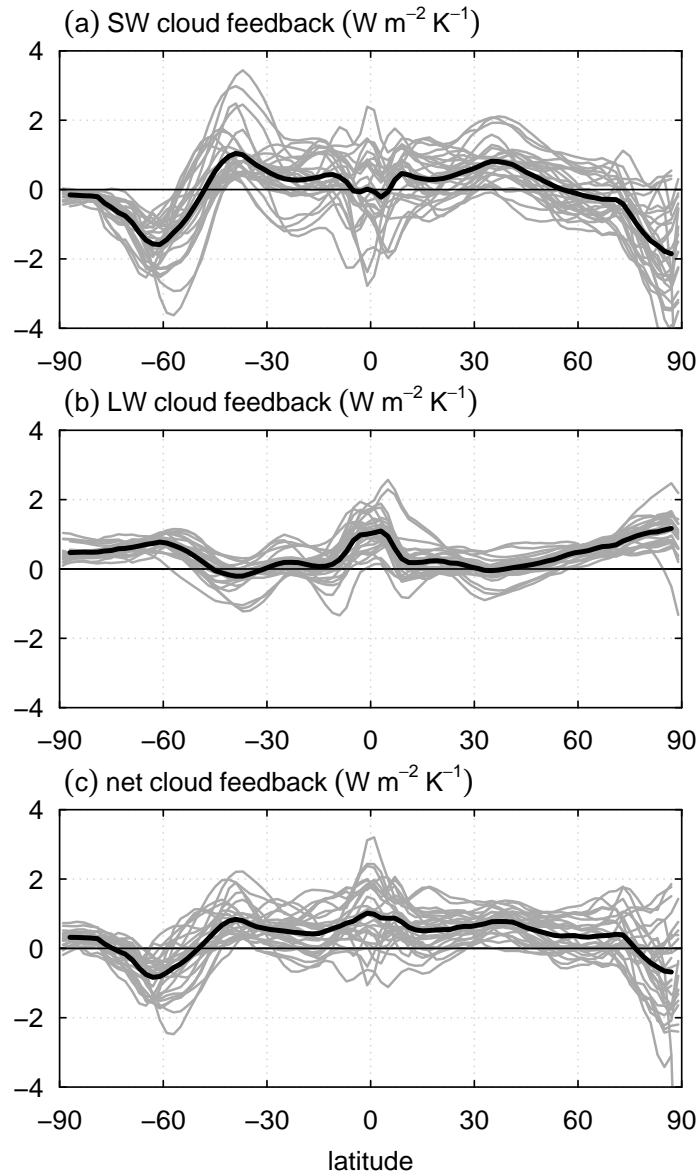
837 FIG. 4. (a,c,e) As in Fig. 2a, but showing the response of the vertically-averaged tropospheric temperature.
 838 $\langle T \rangle_{\text{upper}}$, $\langle T \rangle_{\text{lower}}$, and $\langle T \rangle$ denote upper-tropospheric (100 to 500 hPa), lower-tropospheric (500 to 1000 hPa),
 839 and tropospheric (100 to 1000 hPa) vertical-mean temperature, respectively. (b,d,f) Changes in the meridional
 840 gradient of $\langle T \rangle$ at various tropospheric levels, calculated as the change in tropical-mean $\langle T \rangle$ (30° S to 30° N)
 841 minus the change in extratropical-mean $\langle T \rangle$.



842 FIG. 5. As in Fig. 3, but for eddy kinetic energy (EKE, left) and meridional mass streamfunction (Ψ , right).
 843 Grey contours show the control climatology in intervals of $40 m^2 s^{-2}$ (EKE, left) and $60 Sv$ (Ψ , right), with
 844 negative values dashed and the zero contour omitted.



845 FIG. 6. 4xCO₂ response of various circulation metrics: Hadley cell edge defined as the first zero-crossing of
 846 the mass streamfunction at 500 hPa ($\Psi_{500} = 0$); latitude where precipitation equals evaporation in the subtropics
 847 ($P - E = 0$); jet latitude defined as the peak in zonal-mean zonal wind (ϕ_{jet}); storm-track latitude defined as the
 848 peak in sea-level pressure variance ($\phi_{\sigma^2(SLP)}$). All results are averaged over both hemispheres.



849 FIG. 7. Cloud feedback components in the abrupt4xCO₂ experiment of 28 CMIP5 models, all calculated as
 850 years 121–140 minus the pre-industrial control climatology. Grey curves represent individual models, with the
 851 multi-model mean in thick black. The cloud feedback is calculated using radiative kernels, following the method
 852 of Soden et al. (2008), and includes rapid adjustments to CO₂ forcing (Sherwood et al. 2015).