

Fast and slow components of the extratropical atmospheric circulation response to CO2 forcing

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1	Fast and slow components of the extratropical atmospheric circulation
2	response to CO ₂ forcing
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ABSTRACT

Poleward shifts of the extratropical atmospheric circulation are a common 11 response to CO₂ forcing in global climate models (GCMs), but little is known 12 about the time dependence of this response. Here it is shown that in coupled 13 climate models, the long-term evolution of sea surface temperatures (SSTs) 14 induces two distinct time scales of circulation response to step-like CO2 forc-15 ing. In most Coupled Model Intercomparison Project phase 5 GCMs as well 16 as in the multi-model mean, all of the poleward shift of the midlatitude jets and 17 Hadley cell edge occurs in a fast response within 5 to 10 years of the forcing, 18 during which less than half of the expected equilibrium warming is realized. 19 Compared with this fast response, the slow response over subsequent decades 20 to centuries features stronger polar amplification (especially in the Antarctic), 21 enhanced warming in the Southern Ocean, an El Niño-like pattern of tropical 22 Pacific warming, and weaker land-sea contrast. Atmosphere-only GCM ex-23 periments demonstrate that the SST evolution drives the difference between 24 the fast and slow circulation responses, although the direct radiative effect of 25 CO_2 also contributes to the fast response. It is further shown that the fast and 26 slow responses determine the long-term evolution of the circulation response 27 to warming in the RCP4.5 scenario. The results imply that shifts in midlat-28 itude circulation generally scale with the radiative forcing, rather than with 29 global-mean temperature change. A corollary is that time slices taken from 30 a transient simulation at a given level of warming will considerably overesti-3 mate the extratropical circulation response in a stabilized climate. 32

2

1. Introduction

A well-known feature of the atmospheric circulation response to CO_2 forcing is the overall 34 poleward shift of extratropical circulation, including the jet streams (Kushner et al. 2001; Yin 35 2005; Barnes and Polvani 2013), the storm tracks (Chang et al. 2012; Harvey et al. 2014), and the 36 edge of the tropics (Lu et al. 2007; Kang and Polvani 2011; Ceppi et al. 2013). This poleward 37 shift is primarily mediated by sea surface temperature (SST) changes, as demonstrated by climate 38 model experiments forced only with a prescribed SST increase (Brayshaw et al. 2008; Staten et al. 39 2012; Grise and Polvani 2014), although the direct effect of CO₂ (in the absence of any SST 40 changes) also contributes to the poleward circulation shift (Deser and Phillips 2009; Staten et al. 41 2012; Grise and Polvani 2014). 42

In previous analyses of atmospheric circulation change under greenhouse gas forcing, the cir-43 culation response is typically defined as the difference in climatology between a control present-44 day (or pre-industrial) state, and a future warmer state. While convenient, such a definition con-45 ceals any possible time dependence of the forced circulation response. Since circulation shifts 46 are mainly driven by increasing SST, a simple, naïve assumption is that the circulation will shift 47 at the same rate as global-mean warming over the course of the transient response to greenhouse 48 gas forcing. A related assumption that spatial patterns of climate response scale with global-mean 49 temperature change, known as "pattern scaling," is commonly made for temperature and precipita-50 tion, for example when estimating regional climate responses under scenarios for which no global 51 climate model (GCM) simulations are available (e.g., Santer et al. 1990; Mitchell 2003; Tebaldi 52 and Arblaster 2014, and references therein). 53

It is known, however, that transient patterns of SST response evolve over time following CO_2 forcing – in violation of the pattern scaling assumption – primarily because the ocean system

includes processes characterized by multiple time scales. In particular, GCMs forced with an 56 abrupt CO_2 increase show that SST anomalies in regions such as the Southern Ocean, the North 57 Atlantic, and the tropical Pacific substantially deviate from linearity with respect to global-mean 58 warming over the course of the transient response (Manabe et al. 1990, 1991; Stouffer 2004; Held 59 et al. 2010; Armour et al. 2013; Geoffroy and Saint-Martin 2014; Long et al. 2014; Rugenstein 60 et al. 2016b). Since the extratropical circulation response depends sensitively on the spatial pattern 61 of warming (e.g., Butler et al. 2010; Chen et al. 2010; Harvey et al. 2014; Ceppi et al. 2014), this 62 suggests that midlatitude circulation changes may be characterized by multiple time scales, and 63 may not generally scale with global-mean temperature change. The impact of the evolution of 64 SSTs on the time scales of circulation change would be in addition to the previously identified 65 rapid dynamical adjustment to CO₂ forcing, which acts on a time scale of weeks to months (Deser 66 and Phillips 2009; Staten et al. 2012; Wu et al. 2013; Bony et al. 2013; Grise and Polvani 2014, 67 2017). 68

In this paper we demonstrate that the SST-mediated midlatitude circulation response to CO_2 69 forcing involves two distinct time scales, which can be explained by time-evolving patterns of 70 SST change. In the majority of CMIP5 GCMs and in the multi-model mean, all of the poleward 71 shift occurs in a fast response (including the direct CO₂ response) within 5 to 10 years of the 72 forcing. To demonstrate the existence of distinct time scales of atmospheric circulation change, 73 we analyze abrupt CO₂ forcing CMIP5 experiments (section 3), which provide the best possible 74 separation between the various time scales of climate response to radiative forcing. In section 75 4, we then show that the same time scales of response also operate in RCP4.5, a scenario with 76 gradually increasing forcing. Finally, we summarize and discuss our results in section 5. 77

78 2. Data and Methods

79 a. Climate model experiments

Most of the results presented in this paper are based on CMIP5 coupled atmosphere-ocean GCM 80 experiments (Taylor et al. 2012). The atmospheric circulation response to warming is assessed 81 in 28 140-year abrupt4×CO₂ simulations, in which atmospheric CO₂ concentration is instan-82 taneously quadrupled relative to pre-industrial values at the start of year 1, then held constant. 83 Climate anomalies are calculated by subtracting the parallel reference pre-industrial control inte-84 gration from the abrupt $4 \times CO_2$ simulation, to remove any model drift. Monthly-mean fields are 85 aggregated into annual-mean values prior to analysis. The models included in the analysis are 86 listed in Table 1. 87

By the end of the 140-year abrupt4 \times CO₂ experiments, climate has not yet reached a steady state 88 due to the long equilibration time scale of the ocean. To explore the relationship between circu-89 lation change and warming on time scales longer than 140 years, we use an ensemble of coupled 90 abrupt $4 \times CO_2$ integrations of the Community Earth System Model (CESM; Hurrell et al. 2013) 91 with the atmospheric component CAM4 (Neale et al. 2010) extending to 1000 years, described 92 in Rugenstein et al. (2016a). The ensemble includes 121 members during the first two years, 13 93 members between years 3 and 100, 6 members between years 101 and 250, and 1 member for the 94 remainder of the integration. The ensemble members are branched off in January of subsequent 95 years of the reference pre-industrial simulation. We use only the ensemble mean in our analysis. 96 In addition to these coupled simulations, we also perform atmosphere-only CAM4 experiments

⁹⁷ In addition to these coupled simulations, we also perform atmosphere-only CAM4 experiments ⁹⁸ with imposed patterns of SST change, designed to understand the role of time-varying patterns of ⁹⁹ surface warming for the circulation response. These experiments are run for 25 years after 1 year ¹⁰⁰ of spin-up. Both the coupled and the atmosphere-only integrations are performed at a resolution ¹⁰¹ of 1.9° latitude by 2.5° longitude with 26 vertical levels.

102 b. Atmospheric circulation metrics

¹⁰³ In this paper we focus on meridional shifts of the zonal-mean circulation, quantified by indices ¹⁰⁴ of jet latitude and poleward edge of the Hadley cells. The jet latitude is calculated separately ¹⁰⁵ for the Southern Hemisphere, the North Pacific basin (140° E to 120° W), and the North At-¹⁰⁶ lantic/European sector (60° W to 60° E). Jet latitude is defined as a centroid of the 850 hPa zonal ¹⁰⁷ wind distribution between 30° and 60°,

$$\phi_{\rm jet} = \int_{30^\circ}^{60^\circ} \phi \, \bar{u}^2 \, d\phi \Big/ \int_{30^\circ}^{60^\circ} \bar{u}^2 \, d\phi,$$

where ϕ is latitude and the overbar denotes a zonal average; latitudes with climatological easterlies 108 are excluded from the calculation. Using the square of the zonal wind ensures that more weight 109 is given to latitudes of strong westerly wind. Similar jet definitions have been used in previous 110 literature (Chen et al. 2008; Ceppi et al. 2014). For the Hadley cell edge, we use the latitude 111 where the meridional mass streamfunction crosses zero in the subtropics at 500 hPa, after cubically 112 interpolating the values onto a 0.1° latitude grid. Note that very similar results are obtained if the 113 latitude of zero surface zonal-mean zonal wind in the subtropics is used instead as a measure of the 114 Hadley cell edge, as in Vallis et al. (2015) (not shown). All shifts are defined as positive poleward. 115

3. Circulation response to abrupt CO₂ forcing

a. Two time scales of climate response

Plotting jet latitude against global-mean temperature anomaly reveals the existence of two distinct time scales of atmospheric circulation response to CO_2 forcing in abrupt4× CO_2 experiments

(Fig. 1). Following CO_2 quadrupling, the multi-model mean jets rapidly shift poleward with in-120 creasing temperature during the first few years of the integrations. However, the shifting tends 121 to cease after about 5 years, despite steadily increasing global-mean temperature; the mean trend 122 even reverses in the North Pacific basin, where the zonal-mean jet returns to its original latitude by 123 the end of the abrupt $4 \times CO_2$ simulations. Henceforth we define the "fast" and "slow" circulation 124 responses as the changes between the control climate and the mean of years 5-10, and between 125 years 5–10 and 121–140, respectively (black crosses in Fig. 1). During the fast response, the planet 126 warms by 3.0 K on average, less than half the expected equilibrium warming of 6.6 K based on 127 estimated forcing and feedback values in our set of GCMs (Caldwell et al. 2016). 128

Despite considerable inter-model spread in jet shift, as evidenced by the 75% intervals in Fig. 1, 129 the tendency for a weaker poleward shift in the slow response is robust across climate models 130 (Fig. 2). In the Southern Hemisphere (SH), this difference is present in all of the models; and 131 while the circulation systematically shifts poleward in the fast response, the shifts are as often 132 positive as negative in the slow response, with no shift in the multi-model mean. In the Northern 133 Hemisphere (NH), the spread is larger but only a few models show a more positive shift in the slow 134 response. The Hadley cell edge response is consistent with that of the midlatitude jets, suggesting 135 that coherent changes in large-scale circulation sensitivity to warming occur between the fast and 136 slow responses. 137

The direct response to CO₂ forcing, occurring on a time scale of weeks to months, is part of the fast response as defined here and may partly account for the nonlinear relationship between circulation shifts and global-mean temperature identified in Figs. 1 and 2 (Staten et al. 2012; Wu et al. 2013; Grise and Polvani 2014, 2017). However, this effect should be restricted to year 1, and therefore cannot account for the bulk of the circulation shift by years 5–10 (Fig. 1). To understand the time scales of atmospheric circulation shifts, we therefore turn to the evolution of patterns of SST change during the transient response to CO₂ forcing (e.g., Manabe et al. 1990;
Held et al. 2010; Long et al. 2014). The evolution of SST patterns could have implications for
changes in baroclinicity (i.e. meridional temperature gradients and vertical stability), important for
midlatitude circulation shifts. We investigate this possibility in the next subsection by considering
the joint evolution of the patterns of surface temperature and zonal wind response.

¹⁴⁹ b. Spatial patterns of temperature and zonal wind response

The multi-model mean fast and slow patterns of surface air temperature change, and the cor-150 responding 850 hPa zonal wind anomaly patterns, are shown in Fig. 3. Evident differences are 151 visible between the fast and slow warming patterns, which are robust across models (stippled re-152 gions in Fig. 3). Part of these differences are consistent with the rapid adjustment to CO_2 forcing 153 (taking place during the first few weeks to months following the CO_2 increase), associated with 154 enhanced warming over land relative to ocean areas in the fast response. Large differences in 155 warming pattern between fast and slow responses also occur over the ocean, however, reflect-156 ing differences in the pattern of SST change. The Southern Ocean particularly stands out due to 157 strongly suppressed warming in the fast response relative to the global mean, while in the slow re-158 sponse it warms on par with the global average. Instead of the interhemispheric gradient found in 159 the fast response, the slow response pattern is generally characterized by a more hemispherically 160 symmetric SST increase, with a tendency toward an El Niño-like pattern in the tropical Pacific 161 (Collins et al. 2005; Kohyama and Hartmann 2016), slightly suppressed subtropical warming rel-162 ative to the global mean, and suppressed warming in the North Atlantic, due to a weakening of the 163 meridional overturning circulation in that ocean basin (Drijfhout et al. 2012; Collins et al. 2013). 164 The slow response pattern also features a higher degree of polar amplification compared with the 165 fast response, particularly over the Antarctic cap. 166

The differences between fast and slow temperature and circulation responses are consistent with 167 the understanding that the ocean thermodynamic response to forcing is dominated by two time 168 scales: a fast time scale of a few years associated with the coupled atmosphere-mixed-layer ocean 169 system, and a much slower time scale (of the order of 100 years) determined by the large heat 170 capacity of the deep ocean (Dickinson 1981; Manabe et al. 1990; Gregory 2000; Held et al. 2010; 171 Olivié et al. 2012; Geoffroy et al. 2013). While the distinction between time scales of mixed-172 layer and deep ocean warming offers a plausible explanation for the time dependence of SST 173 warming patterns, various additional processes also contribute to local SST changes, including the 174 climatological ocean circulation (Armour et al. 2016), changes in ocean circulation (Drijfhout et al. 175 2012; Woollings et al. 2012), and coupled air-sea feedbacks (Bjerknes 1969; Xie and Philander 176 1994; Clement et al. 1996; Xie et al. 2010), to name a few. As an additional caveat, the time scales 177 of ocean heat uptake may well vary regionally, so that the evolution of SSTs cannot be entirely 178 captured by two time scales only. Understanding the evolution of transient SST anomaly patterns 179 is beyond the scope of this work, but we note that the fast and slow warming patterns in Fig. 3 are 180 highly consistent with those documented in previous work in different sets of GCMs (Held et al. 181 2010; Geoffroy and Saint-Martin 2014; Long et al. 2014), suggesting that the processes underlying 182 the time dependence of SST patterns are reasonably robust across GCMs. 183

The fast and slow zonal wind response patterns (right column of Fig. 3) reflect the evolution of jet latitude seen in Fig. 1: while the jets shift poleward in all regions in the fast response, a weak equatorward jet shift is visible in the North Pacific in the slow response, with little change in extratropical zonal wind elsewhere. To understand the relationship between circulation responses and warming patterns, it is helpful to consider the patterns in Fig. 3 along with the vertical structure of the changes in zonal-mean temperature and wind shown in Fig. 4. First focusing on the SH, we note that in the fast response, the delayed Southern Ocean warming causes an anomalously strong

meridional temperature gradient across the midlatitudes throughout the troposphere (Fig. 4a), fa-191 voring a strengthening and poleward shift of the eddy-driven jet (Butler et al. 2010; Chen et al. 192 2010; Harvey et al. 2014; Ceppi and Hartmann 2016). By contrast, the slow warming pattern 193 is associated with a clear weakening of the meridional temperature gradient at lower and middle 194 tropospheric levels, due to amplified Antarctic warming, which alone would favor an equatorward 195 jet shift (Butler et al. 2010). The lack of a clear SH zonal wind response to the slow warming 196 reflects cancelling effects of upper- and lower-level temperature gradient changes (Harvey et al. 197 2014; Ceppi and Hartmann 2016). 198

In the NH, the weaker fast jet response in the NH relative to the SH is consistent with the effect of 199 amplified Arctic warming on midlatitude baroclinicity (Fig. 4a,c). In the slow response, warming 200 becomes more muted in the subtropics to midlatitudes, so that the low-level temperature gradient 201 across the midlatitudes weakens further, which may contribute to the slight equatorward shift of the 202 zonal-mean circulation (Fig. 4b,d). However, zonal asymmetries in warming may also contribute 203 substantially to the NH jet and stationary wave response (Delcambre et al. 2013; Simpson et al. 204 2014). In particular, the slow warming pattern includes an El Niño-like component in the tropical 205 Pacific (Fig. 3b) which may contribute to the North Pacific jet response. In the next subsection, we 206 demonstrate that the SST anomaly patterns are primarily responsible for the differences between 207 fast and slow temperature and zonal wind responses. 208

²⁰⁹ c. Relative roles of direct and SST-mediated effects of CO₂

To confirm the key role of surface warming patterns for differences in circulation sensitivity to warming, and to disentangle the contributions of the direct component of CO₂ forcing and SST change to the atmospheric circulation response, we perform atmosphere-only GCM (AGCM) experiments in which we separately impose the multi-model mean fast SST change, the slow SST change, and the CO₂ increase while keeping SSTs unchanged. The perturbed SST experiments
also include the corresponding changes in sea ice cover. Climate responses are calculated relative
to an experiment with SSTs and sea ice taken from the pre-industrial control CMIP5 multi-model
mean.

We first consider the bottom two rows of Fig. 5, which can be directly compared with Fig. 4. 218 When forced with the multi-model mean SST and CO₂ changes¹, our AGCM produces temper-219 ature and zonal wind changes in close agreement with the CMIP5 model mean. In particular, it 220 recovers the large difference in jet sensitivity to global warming between the fast and slow re-221 sponse. The fast response can be further decomposed into contributions of direct radiative forcing 222 of CO₂ and SST changes (top two rows of Fig. 5). This reveals that SST changes account for most 223 of the tropospheric temperature changes and SH jet shift in the fast response; however, the direct 224 effect of CO₂ also causes a poleward jet shift in both hemispheres, associated with tropospheric 225 warming (particularly over NH landmasses) and strong stratospheric cooling. Note that the direct 226 effect of CO₂ on circulation seems to be larger in this AGCM compared with most CMIP5 models 227 (cf. the year 1 response in Fig. 1 and Grise and Polvani 2014). 228

229 d. Centennial changes in temperature and circulation

Because the ocean takes centuries to equilibrate with the imposed greenhouse gas forcing, the model climates have not reached equilibrium by the end of the CMIP5 abrupt4×CO₂ experiments. Consequently, the patterns of temperature and circulation response continue evolving after year 140 of the experiment. We investigate the centennial circulation response using a 1000-year abrupt4×CO₂ experiment with CESM (section 2a). As shown in Fig. 6, the relationship between

¹Note that the fast SST and CO₂ changes are imposed in separate experiments, and the responses are added to obtain the combined effect in Fig. 5c,g. Previous work suggests that these responses are approximately additive (Deser and Phillips 2009; Staten et al. 2012).

jet shift and global-mean temperature in CESM is in good qualitative agreement with the mean
CMIP5 model behavior: the jets shift poleward during the first few years of the integration, following which the jet latitude stabilizes – or decreases in the case of the North Pacific jet. The main
differences relative to the CMIP5 ensemble are (a) larger North Pacific jet fast and slow responses,
(b) a weaker SH jet shift, and (c) a shorter time scale for the fast response (the peak jet latitude
being reached by year 2 or 3).

Warming patterns being specific to each model, it is unsurprising that CESM's fast and slow 241 temperature and zonal wind patterns present differences relative to CMIP5 models (top two rows 242 of Fig. 7, vs. Fig. 3). In the fast (subdecadal) temperature response, Southern Ocean warming is 243 less suppressed compared with the CMIP5 ensemble, and larger zonal asymmetries are present 244 in the tropics. These features are consistent with a weak SH jet shift, and with a large tropical 245 zonal wind response that is absent from the CMIP5 multi-model mean (Fig. 7a,d). Nevertheless, 246 clear similarities are also visible in the temporal evolution of these patterns: as in CMIP5, the 247 slow response shows a transition to a more hemispherically symmetric temperature pattern, with 248 delayed Antarctic and Southern Ocean warming and an El Niño-like pattern of SST anomalies in 249 the tropical Pacific in the slow (decadal) response. 250

Beyond year 140 of the abrupt4 \times CO₂ experiment, the patterns of temperature and zonal wind 251 response continue evolving (the centennial response in Fig. 7c, f). The surface warming pattern be-252 comes increasingly hemispherically and zonally symmetric, being mainly characterized by polar 253 amplification. This favors a slight weakening of the midlatitude westerlies, particularly in the SH 254 and in the North Atlantic. The weak overall changes in extratropical winds once again suggest can-255 celing effects between polar-amplified warming at low levels, and tropically-amplified warming 256 aloft, causing meridional temperature gradient changes of opposite sign. Taken together, Figs. 6 257 and 7 suggest that the circulation response to CO_2 forcing is primarily determined by the changes 258

occurring during the first 140 years following the forcing; the very slow warming on time scales
of centuries to millennia does not strongly change the nature of the dynamical response, particularly in the extratropics, and does not cause further poleward circulation shifts. However, since the
ocean processes controlling long-term warming patterns remain poorly understood and are likely
to vary across models, this result will need to be further tested with other coupled GCMs.

4. Fast and slow circulation responses in RCP4.5

²⁶⁵ a. Relationship between step and gradual forcing experiments

The abrupt4×CO₂ experiments considered so far are helpful in understanding the relationship 266 between atmospheric circulation and global-mean temperature anomaly because they provide an 267 optimal time scale separation and a good signal-to-noise ratio thanks to the large forcing. However, 268 this understanding is interesting mainly to the extent that it can be applied to more realistic gradual 269 forcing scenarios. If the climate responses are linear in forcing magnitude, then any greenhouse 270 gas forcing experiment can be understood as consisting of a sum of responses to small abrupt 271 CO₂ forcings at various time scales (Good et al. 2011, 2013). Linearity in forcing magnitude has 272 been shown to hold to a good approximation for the temperature response (Good et al. 2013), 273 meaning that the gradual forcing responses can be traced back to abrupt experiments. In this 274 section, we demonstrate that the two time scales of circulation response identified in $abrupt4 \times CO_2$ 275 integrations are also expressed in gradual forcing experiments, causing a decrease in the tendency 276 for the circulation to shift poleward with warming as greenhouse gas concentrations stabilise and 277 climate approaches equilibrium. 278

To test the applicability of our findings to realistic future scenarios, we consider the RCP4.5 experiment in CMIP5, for which 12 GCMs have provided long integrations reaching year 2299 (Ta²⁸¹ ble 1). We select this experiment because the anthropogenic forcing agent concentrations are sta-²⁸² bilized relatively early in the experiment (around year 2080, compared with year 2250 in RCP8.5), ²⁸³ offering a chance to detect the various time scales of temperature and circulation response in the ²⁸⁴ experiment. Although the anthropogenic forcing peaks even earlier in RCP2.6 (around 2050), ²⁸⁵ the small magnitude of the forcing compared with RCP4.5 makes it more difficult to separate the ²⁸⁶ signal from the noise in the dynamical response.

The time series of the sum of anthropogenic forcing agents (expressed as CO_2 -equivalent concentrations in ppm; Meinshausen et al. 2011) and global-mean surface air temperature anomaly relative to 1900–1949 are shown in Fig. 8 (black curves). The total concentration of anthropogenic forcing agents (dominated by CO_2) quickly rises between the late twentieth century and about 2080, after which it remains approximately stable. Consistent with this, global-mean temperature rises rapidly until the late twenty-first century, but continues increasing more slowly for the following two centuries as the deep ocean slowly adjusts to the forcing.

To relate the RCP4.5 responses to the abrupt $4 \times CO_2$ experiments, a few assumptions are nec-294 essary. In addition to assuming that the response is linear in forcing magnitude, we make the 295 simplification that the response to abrupt CO₂ forcing can be fully characterized by a combination 296 of the two patterns identified in section 3a. We also make the further assumption that all anthro-297 pogenic forcing agents produce the same patterns of response as CO₂. This assumption is likely 298 to be inaccurate in the case of aerosol forcing, whose warming patterns are distinct from those 299 induced by CO_2 (Wang et al. 2016) – even though the patterns also include common features due 300 to similar ocean-atmosphere feedbacks (Xie et al. 2013). To the extent that the above assumptions 301 are true, the climate responses in RCP4.5 can be entirely characterized as linear combinations of 302 the fast and slow responses identified in $abrupt4 \times CO_2$. 303

We test these assumptions by regressing the annual-mean, multi-model mean surface air tem-304 perature anomaly in RCP4.5 (relative to the 1900–1949 historical climate, in K), separately for 305 each year, onto the fast and slow warming patterns (in K K^{-1} ; Fig. 3a,b). This yields two re-306 gression coefficients that quantify the relative contributions of the fast and slow patterns to the 307 RCP4.5 global-mean temperature anomalies in any given year, plus an intercept which we de-308 scribe as a residual (Fig. 8b, colored curves). By construction, the regression coefficients and 309 the intercept all have units of K, making their physical interpretation straightforward. Since the 310 fast response occurs within 10 years of the forcing, we expect the fast contribution to warming to 311 closely track the evolution of radiative forcing, while the slow contribution should increase more 312 gradually and continue growing well after the forcing agents stabilize. The regression coefficients 313 are in excellent agreement with our expectation, and the sum of the fast and slow contributions 314 (the "reconstructed" global-mean warming, red curve) closely follows the actual values (Fig. 8b). 315 The coefficient of determination of the regression (R^2) – a measure of the fraction of the spatial 316 variance in the warming pattern that can be explained by our regression model – increases from 317 about 80% in year 2000 to over 95% in year 2050 and beyond. The lower values during the twen-318 tieth century could reflect the effects of aerosol forcing on temperature anomaly patterns (next 319 paragraph), but more likely result from the low signal-to-noise ratio during this period when the 320 forcing is still relatively small. From the above results we conclude that to a good approximation, 321 the responses to gradually increasing forcing at any point in time can be understood as a linear 322 combination of fast and slow responses to abrupt CO₂ forcing. 323

As an aside, we note that during the late twentieth century, the slow contribution grows more rapidly than the fast contribution; this may reflect the mid-century dip in radiative forcing associated with aerosols, to which the fast component responds while the slow component is more sensitive to the cumulative forcing. The partitioning between fast and slow contributions is likely to be less accurate in the mid-twentieth century than in subsequent periods, because the temperature fingerprint of aerosol forcing may not be entirely captured by the fast and slow warming patterns of CO₂. This seems consistent with the regression residual developing during the late twentieth century, and remaining nearly constant thereafter, once the warming becomes dominated by greenhouse gases (purple curve in Fig. 8b). It is also consistent with the low value of R^2 prior to about the year 2000.

b. Contributions of fast and slow responses to RCP4.5 jet shifts

The varying relative importance of the fast and slow patterns of response suggests that the cir-335 culation shifts per unit warming should also vary with time in RCP4.5. Since the SH and North 336 Atlantic jets shift only in the fast response, we expect the shifts of these jets to scale with the 337 fast contribution to warming in RCP4.5, and therefore approximately with the radiative forcing, 338 rather than with warming. The North Pacific jet response should depend on both the fast and slow 339 contributions, but should exhibit a more marked equatorward shifting tendency as climate nears 340 equilibrium, when the slow warming pattern becomes more dominant. These predictions can be 341 made quantitative by reconstructing the zonal wind response as a linear combination of the fast 342 and slow patterns (Fig. 3c,d) multiplied by the respective regression coefficients (Fig. 8, middle). 343 It should be borne in mind that this zonal wind reconstruction is entirely based on the patterns of 344 SST change, and therefore it cannot include the effects of stratospheric ozone depletion on the SH 345 jet, as discussed below. 346

Figure 9 shows the jet latitude as a function of global-mean warming for the actual (black curves) and the reconstructed (red) zonal wind fields. Overall, the jet responses tend to scale more linearly with warming than in abrupt $4 \times CO_2$, as expected if the fast and slow time scales of response overlap because of the gradually increasing forcing. However, the SH and North Atlantic jets still

show separate time scales of response (black curves in Fig. 9), with an initial poleward shift with 351 warming followed by a stabilization once the forcing has reached its peak (grey vertical bars at 352 year 2080). The zonal wind reconstruction captures these different time scales well (red curves). 353 In the SH, until about 2050 the jet shifts further poleward than would be anticipated based on 354 SST anomaly patterns alone, but this is perfectly consistent with the effect of ozone depletion and 355 recovery (Arblaster and Meehl 2006; Son et al. 2010; McLandress et al. 2011; Barnes et al. 2014). 356 The North Atlantic poleward jet shift is also somewhat overpredicted, but the temporal evolution is 357 well captured by the zonal wind reconstruction. The reconstructed North Pacific jet shift shows no 358 clear response until 2080, followed by a very weak equatorward shift, in agreement with the actual 359 jet behavior. To gain additional insight into the circulation response, we calculate separate jet shift 360 indices for the fast and slow contributions, by using only either the fast or the slow component of 361 the zonal wind change. This confirms that the SH and North Atlantic jet responses are entirely 362 due to the contribution of the fast response to CO_2 forcing – and therefore occur only as long as 363 the radiative forcing keeps increasing – whereas the North Pacific jet remains at a nearly constant 364 latitude owing to competing effects of the fast and slow zonal wind changes. 365

To fully appreciate the significance of the results in Fig. 9, it is worth keeping in mind that, similar to the abrupt4×CO₂ integrations, the RCP4.5 runs have not reached equilibrium by the end of the simulations. Hence substantial further warming could occur beyond year 2300 with no accompanying circulation shift. To highlight this, we approximate the equilibrium warming following the method of Gregory et al. (2004), as described in the Appendix, and calculate the equilibrated jet response under the assumption that all of the long-term warming is associated with the slow pattern.² This calculation suggests that the planet would warm by a further 0.75 K beyond

 $^{^{2}}$ As a caveat, Fig. 7 suggests that at least in CESM, the latter assumption would not be entirely accurate and would lead to an equatorward bias of the North Pacific jet response, for example.

³⁷³ year 2300, with the North Pacific jet shifting slightly equatorward while the SH and North Atlantic ³⁷⁴ jets would remain at near-constant latitude (red dots in Fig. 9). Note that our simple calculation ³⁷⁵ of equilibrium warming likely underestimates the true value (see Appendix). Overall, the clear ³⁷⁶ deviation from linearity in warming indicates that pattern scaling would be a poor assumption to ³⁷⁷ estimate equilibrium circulation responses to greenhouse gas forcing from the transient responses, ³⁷⁸ as discussed in the next section.

5. Discussion and Conclusions

The purpose of this paper is to show that owing to the evolution of spatial patterns of SST in-380 crease, the extratropical atmospheric circulation response to greenhouse gas forcing involves two 381 distinct time scales with different characteristics, and consequently midlatitude circulation shifts 382 do not generally scale with global-mean temperature change. Following abrupt CO_2 forcing, pole-383 ward circulation shifts occur mainly during the first 5 to 10 years. In subsequent decades, the 384 multi-model mean SH and North Atlantic jets remain at a nearly constant latitude despite sub-385 stantial global warming, while the North Pacific jet shifts back equatorward. AGCM experiments 386 demonstrate that the two time scales of circulation response are primarily determined by distinct 387 patterns of SST change. "Slow" warming on time scales longer than 10 years is associated with 388 a pattern that has a relatively high degree of low-level polar amplification and is therefore less 389 effective at causing poleward circulation shifts compared with the "fast" warming in the initial 5 390 to 10 years. In addition to the effect of SSTs, the direct radiative effect of CO_2 also contributes 391 to the fast poleward circulation shift, in line with previous results (Staten et al. 2012; Grise and 392 Polvani 2014). However, the direct response should be restricted to year 1, and therefore cannot 393 account for the bulk of the circulation shift by years 5-10. 394

Our results imply that poleward circulation shifts generally scale with the cumulative amplitude 395 of the radiative forcing, rather than with the global-mean warming. This is shown to be true in 396 the RCP4.5 experiment, whose response is determined by the same fast and slow patterns as in 397 abrupt $4 \times CO_2$. Under a scenario in which forcing agents peak and stabilize, we can therefore 398 expect the extratropical circulation to rapidly reach a near-equilibrium, in considerably less time 399 than it takes the climate system to equilibrate. As a corollary, if radiative forcing were to decrease 400 in the future, for example by means of carbon dioxide removal, atmospheric circulation would be 401 expected to respond within a few years. Thus, our results imply that climate change mitigation 402 actions would have a more rapid impact on extratropical atmospheric circulation than on other 403 aspects of climate change related to global-mean temperature. 404

We have not discussed the seasonality of the time scales of circulation change. In their analysis 405 of the evolution of SH circulation response to CO_2 forcing, Grise and Polvani (2017) found that 406 the jet shift was faster during austral winter than during summer, and the evolution of jet latitude 407 in summer was more similar to that of global-mean temperature. We have analyzed the evolution 408 of SSTs and circulation separately for half-year seasons (November–April and May–October), 409 and found a qualitatively similar evolution in both seasons: the overall features of the fast and 410 slow patterns of SST change show little seasonality, and the majority of the poleward shift occurs 411 within the fast response in each extended season (not shown). In agreement with Grise and Polvani 412 (2017), a weak poleward shift persists in the slow response during austral summer, which these 413 authors ascribe to the evolution of polar lower stratospheric temperature. Hence, the specific 414 character of the slow response may vary seasonally, but the annual-mean perspective is sufficient 415 to demonstrate how the fast and slow time scales in the SST response trigger very different global 416 circulation changes. 417

Our results suggest that care is warranted when using pattern scaling approaches to estimate at-418 mospheric circulation responses at different levels of equilibration from transient simulations. As 419 an example, the impacts of 2 K global-mean warming – a common policy target (Randalls 2010) – 420 are sometimes assessed by taking a time slice around the time of 2 K warming in transient simula-421 tions that are far from reaching steady state (e.g., Schleussner et al. 2016). Applying this method 422 yields an estimated SH jet shift of 1.0° , about two-thirds larger than the estimated *equilibrium* shift 423 of 0.6° for a 2 K warming scenario (calculated by rescaling the equilibrium jet shift in Fig. 9 for a 424 warming of 2 K). Similar errors could occur when using a pattern scaling approach to reconstruct 425 circulation changes under different scenarios with different forcing histories and levels of equili-426 bration. This does not invalidate pattern scaling in general, however; there is no indication based 427 on our results that pattern scaling would not yield accurate results when reconstructing scenarios 428 at similar levels of equilibration. 429

To conclude, we note that future SST anomaly patterns will have important implications not 430 only for changes in atmospheric circulation and rainfall (Xie et al. 2010; Chadwick et al. 2014), 431 but also for the magnitude of climate feedbacks and therefore climate sensitivity, arguably the 432 most fundamental metric of global climate change (Andrews et al. 2015; Gregory and Andrews 433 2016; Zhou et al. 2016). Current GCMs predict a wide range of patterns of SST response to 434 greenhouse gas forcing, and our understanding of the responsible processes remains too limited to 435 determine which of these various possible responses are more realistic (Vecchi et al. 2008; Collins 436 et al. 2010; Kohyama and Hartmann 2017). Further work is also needed to test the linearity of the 437 patterns of SST change and their associated time scales, for example by comparing the responses 438 to positive and negative radiative forcing (Held et al. 2010; Good et al. 2016). We hope that our 439 results will motivate further theoretical and observational work to better understand the patterns 440 and time scales of SST change in GCMs. 441

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APPENDIX

Estimation of the equilibrium global-mean warming in RCP4.5

Here we describe our approach to estimate the equilibrium global-mean warming values shown in Fig. 9. An external forcing F causes a top-of-atmosphere radiative flux imbalance N according to

$$N = F + \lambda \Delta T, \tag{A1}$$

where λ is the feedback parameter (in W m⁻² K⁻¹), ΔT is the global-mean surface temperature 453 anomaly, and radiative fluxes are positive downward. The feedback parameter λ , which must be 454 negative for a stable system, determines how efficiently the system can restore radiative balance 455 with warming and is treated as a property of the climate model for a given forcing. Once the 456 system has reached equilibrium, N = 0 on average, so we may rewrite Equation A1 as $\Delta T_{eq} =$ 457 $-F_{eq}/\lambda$, where the subscript "eq" denotes equilibrium values. If the forcing is held constant at its 458 equilibrium value, the values of F_{eq} and λ can be calculated for each model as the intercept and 459 slope of a least-squares fit of annually-averaged values of N versus ΔT (Gregory et al. 2004). We 460 use the N and ΔT time series during 2100–2299, when the forcing agents are held constant and 461 the pattern of SST increase is dominated by the slow response. This yields a multi-model mean 462 equilibrium warming value $\Delta T_{eq} = 3.86$ K (Fig. 9). 463

Although we assume the feedback parameter to be a fixed value in our calculation, analyses of coupled atmosphere-ocean CMIP5 GCMs suggest that λ tends to increase (i.e., becomes less negative) over time in abrupt4×CO₂ simulations in most models (Andrews et al. 2012, 2015). As a result, the values of λ calculated by the method of Gregory et al. (2004) may underestimate the effective feedback values, which would result in underestimated equilibrium warming values in Fig. 9. These values should therefore be taken as a likely lower bound for the equilibrium warming in RCP4.5.

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742	Table 1.	List of CMIP5 models used	in t	he a	naly	sis.	Cro	sses	inc	licat	e av	vaila	able	e da	ta fo	or		
743		the respective experiments.		•		•	•	•	•	•								35

TABLE 1. List of CMIP5 models used in the analysis. Crosses indicate available data for the respective
 experiments.

Model name	piControl &	historical &
	$abrupt4 \times CO_2$	RCP4.5
ACCESS1.0	×	
ACCESS1.3	×	
BCC-CSM1.1	×	×
BCC-CSM1.1(m)	×	
BNU-ESM	×	
CanESM2	×	×
CCSM4	×	×
CNRM-CM5	×	×
CSIRO-Mk3.6.0	×	×
FGOALS-g2	×	
FGOALS-s2	×	
GFDL-CM3	×	
GFDL-ESM2G	×	
GFDL-ESM2M	×	
GISS-E2-H	×	×
GISS-E2-R	×	×
HadGEM2-ES	×	
INM-CM4	×	
IPSL-CM5A-LR	×	×
IPSL-CM5A-MR	×	×
IPSL-CM5B-LR	×	
MIROC5	×	
MIROC-ESM	×	×
MPI-ESM-LR	×	×
MPI-ESM-MR	×	
MPI-ESM-P	×	
MRI-CGCM3	×	
NorESM1-M	×	×

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747 748 749 750 751 752 753 754	Fig. 1.	Jet shifts in abrupt4×CO ₂ integrations as a function of global-mean surface air temper- ature anomaly. The curves denote multi-model means, while shading indicates the 75% range (12.5 to 87.5 percentiles of the distribution) of model values. Annual-mean values are shown for years 1–10 (circles) and decadal-mean values for years 11–140 (diamonds). Black crosses indicate the means for years 5–10 and 121–140, and dashed lines represent linearly interpolated values between these points. Zonal wind values are ensemble-averaged year by year prior to calculating jet indices, and are plotted against the multi-model mean temperature.	37
755 756 757 758	Fig. 2.	Fast and slow atmospheric circulation responses to warming in individual models (open circles) and in the multi-model mean (thick crosses). The fast response is defined as the difference in climate between the pre-industrial control and years 5–10, while the slow response is the change between years 5–10 and 121–140.	38
759 760 761 762 763 764	Fig. 3.	Multi-model mean patterns of change in surface air temperature (left) and 850 hPa zonal wind (right) in abrupt4×CO ₂ , all normalized by global-mean warming during the respective periods. To highlight the spatial patterns, we subtract 1 from the temperature patterns to yield a global-mean of 0. Thick grey contours denote the control zonal wind climatology (contours at 5 and 10 m s ⁻¹). Areas where 90% of the models agree on the sign of the response are stippled.	39
765 766 767	Fig. 4.	As in Fig. 3, but for zonal-mean temperature and zonal wind. The global-mean temperature response has been subtracted at each level. Thick grey contours denote the control zonal wind climatology (contours at 10, 20, and 30 m s ^{-1}).	40
768 769 770	Fig. 5.	As in Fig. 4, but for CAM4 AGCM experiments. Panels a, b, e, and f are all normalized by the combined global-mean surface warming due to CO_2 forcing and fast SST pattern, so that the sum of the first two rows equals the third row.	41
771 772 773 774	Fig. 6.	Jet shifts as a function of global-mean surface warming in a 1000-year abrupt $4 \times CO_2$ experiment with CESM. Decadal-mean values are shown for years 11-150 (diamonds) and 50-year means for years 151-1000 (squares). The values are ensemble averages up to year 250 (see text).	42
775 776 777 778	Fig. 7.	As in Fig. 3, but for the 1000-year CESM abrupt $4 \times CO_2$ experiment. The subdecadal and decadal responses correspond to the fast and slow responses in Fig. 3. The centennial response (panels c, f) is defined as the normalized difference between years 121-140 and 951-1000.	43
779 780 781 782 783	Fig. 8.	(a) Time series of CO_2 -equivalent concentration of anthropogenic forcing agents, (b) global- mean, multi-model mean surface temperature anomaly in RCP4.5, relative to the 1900-1949 climatology, and (c) coefficient of determination (R^2) from the regression model. The verti- cal grey bar in panels a–b indicates year 2080, at which point atmospheric CO_2 concentration approximately stabilizes.	44
784 785 786 787 788 789	Fig. 9.	Jet shifts in RCP4.5 as a function of global-mean warming. Open circles denote individual years, while the black curves show 20-year running averages. The red curve is the reconstructed jet latitude evolution, and the open red circle indicates the estimated equilibrium global warming and jet response (see text). Blue and green curves represent the fast and slow warming contributions to jet shifts (see text). The vertical grey bar indicates year 2080, when CO_2 concentration approximately stabilizes.	45



Jet shifts in abrupt4xCO₂ simulations (degrees poleward)

FIG. 1. Jet shifts in abrupt4×CO₂ integrations as a function of global-mean surface air temperature anomaly. The curves denote multi-model means, while shading indicates the 75% range (12.5 to 87.5 percentiles of the distribution) of model values. Annual-mean values are shown for years 1–10 (circles) and decadal-mean values for years 11–140 (diamonds). Black crosses indicate the means for years 5–10 and 121–140, and dashed lines represent linearly interpolated values between these points. Zonal wind values are ensemble-averaged year by year prior to calculating jet indices, and are plotted against the multi-model mean temperature.



FIG. 2. Fast and slow atmospheric circulation responses to warming in individual models (open circles) and in the multi-model mean (thick crosses). The fast response is defined as the difference in climate between the pre-industrial control and years 5–10, while the slow response is the change between years 5–10 and 121–140.



⁷⁹⁹ FIG. 3. Multi-model mean patterns of change in surface air temperature (left) and 850 hPa zonal wind (right) ⁸⁰⁰ in abrupt4×CO₂, all normalized by global-mean warming during the respective periods. To highlight the spatial ⁸⁰¹ patterns, we subtract 1 from the temperature patterns to yield a global-mean of 0. Thick grey contours denote ⁸⁰² the control zonal wind climatology (contours at 5 and 10 m s⁻¹). Areas where 90% of the models agree on the ⁸⁰³ sign of the response are stippled.



FIG. 4. As in Fig. 3, but for zonal-mean temperature and zonal wind. The global-mean temperature response has been subtracted at each level. Thick grey contours denote the control zonal wind climatology (contours at $10, 20, \text{ and } 30 \text{ m s}^{-1}$).



FIG. 5. As in Fig. 4, but for CAM4 AGCM experiments. Panels a, b, e, and f are all normalized by the combined global-mean surface warming due to CO_2 forcing and fast SST pattern, so that the sum of the first two rows equals the third row.



FIG. 6. Jet shifts as a function of global-mean surface warming in a 1000-year abrupt $4 \times CO_2$ experiment with CESM. Decadal-mean values are shown for years 11-150 (diamonds) and 50-year means for years 151-1000 (squares). The values are ensemble averages up to year 250 (see text).



FIG. 7. As in Fig. 3, but for the 1000-year CESM abrupt $4 \times CO_2$ experiment. The subdecadal and decadal responses correspond to the fast and slow responses in Fig. 3. The centennial response (panels c, f) is defined as the normalized difference between years 121-140 and 951-1000.



FIG. 8. (a) Time series of CO₂-equivalent concentration of anthropogenic forcing agents, (b) global-mean, multi-model mean surface temperature anomaly in RCP4.5, relative to the 1900-1949 climatology, and (c) coefficient of determination (R^2) from the regression model. The vertical grey bar in panels a–b indicates year 2080, at which point atmospheric CO₂ concentration approximately stabilizes.



FIG. 9. Jet shifts in RCP4.5 as a function of global-mean warming. Open circles denote individual years, while the black curves show 20-year running averages. The red curve is the reconstructed jet latitude evolution, and the open red circle indicates the estimated equilibrium global warming and jet response (see text). Blue and green curves represent the fast and slow warming contributions to jet shifts (see text). The vertical grey bar indicates year 2080, when CO₂ concentration approximately stabilizes.