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The transient response of global-mean precipitation to increasing carbon dioxide levels

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Abstract
The transient response of global-mean precipitation to an increase in atmospheric carbon dioxide levels of 1% yr^{-1} is investigated in 13 fully coupled atmosphere–ocean general circulation models (AOGCMs) and compared to a period of stabilization. During the period of stabilization, when carbon dioxide levels are held constant at twice their unperturbed level and the climate left to warm, precipitation increases at a rate of ∼2.4% per unit of global-mean surface-air-temperature change in the AOGCMs. However, when carbon dioxide levels are increasing, precipitation increases at a smaller rate of ∼1.5% per unit of global-mean surface-air-temperature change. This difference can be understood by decomposing the precipitation response into an increase from the response to the global surface-temperature increase (and the climate feedbacks it induces), and a fast atmospheric response to the carbon dioxide radiative forcing that acts to decrease precipitation. According to the multi-model mean, stabilizing atmospheric levels of carbon dioxide would lead to a greater rate of precipitation change per unit of global surface-temperature change.

Keywords: precipitation, carbon dioxide, surface-temperature change, climate models

1. Introduction

Global-mean precipitation is an important part of the Earth's climate system; it links the global water and energy cycles through condensational heating of the atmosphere, providing a link between the hydrological cycle and radiative processes such as cloud feedback (Stephens 2005). It is useful to compare changes in global-mean precipitation against the expectations of the Clausius–Clapeyron relation (Held and Soden 2006) and recent global observations (e.g. Wentz *et al* 2007, Adler *et al* 2008), but it may not be so relevant to understanding climate impacts, because regional changes can be significantly larger and of opposite sign to the global-mean change (e.g. Meehl *et al* 2007).

Changes in the Earth's global-mean surface temperature induce various climate feedbacks, such as changes in water vapour, clouds, atmospheric stability and lapse rates, that can influence precipitation processes and lead to changes in precipitation (e.g. Trenberth *et al* 2003). Climate models simulate a change in global precipitation with global surface-temperature change of the order ∼2–3% K^{-1} (Held and Soden 2006, Lambert and Webb 2008). This response is somewhat smaller than some recent observations (∼7% K^{-1}) but still consistent when interdecadal variability is considered (Liepert and Previdi 2009).

As atmospheric moisture storage is small compared to fluxes, global precipitation can be approximated by surface evaporation (Wild and Liepert 2010). The precipitation response can therefore be understood from a surface perspective, where small changes in the atmospheric boundary layer play an important role (e.g. Richter and Xie 2008, Lu and Cai 2009). For example, in response to surface-temperature change alone, we might expect global precipitation to increase at a rate of ∼7% K^{-1} (Richter and Xie 2008). The smaller responses simulated by climate models are achieved by an increase in relative humidity, a decrease in wind speed and an increase in stability near the surface with global-mean surface-temperature change, all of which acts to dampen evaporation and hence precipitation (Richter and Xie 2008, Lu and Cai 2009).

In addition to changing with global-mean surface-temperature change, precipitation is also affected by the change in atmospheric radiative heating caused by the presence of the forcing agent (e.g. Allen and Ingram 2002, Lambert and Webb 2008, Andrews *et al* 2009). In the case of CO_2, whose radiative forcing is mostly felt in the troposphere, this leads to a tropospheric temperature adjustment that occurs before the oceans have time to warm (e.g. Gregory and Webb 2008). This tropospheric temperature adjustment can increase atmospheric stability and reduce convection, leading to a reduction in convective precipitation (Dong *et al* 2009). The easiest way of demonstrating this effect is in climate model experiments whereby the CO_2 level is instantaneously changed but sea-surface-temperatures are held fixed. In such experiments the evaporation and precipitation rate are observed to go down (e.g. Mitchell 1983, Yang *et al* 2003, Dong *et al* 2009, Bala *et al* 2009).

The overall response of precipitation to a change in CO_2 is therefore a combination of the response that scales with global-mean surface-temperature change and the response to tropospheric temperature adjustment to the CO_2 radiative forcing. These two responses emerge on different timescales due to the differing heat capacities of the atmosphere and ocean: the atmospheric response comes about quickly, within a few weeks of the CO_2 perturbation (Dong *et al* 2009), while the response to global-mean surface-temperature change (and the various climate feedbacks that it induces) acts on a multi-annual timescale due to the time it takes for the oceans to warm. In the long term, the response to global-mean surface-temperature change dominates, but in the short term the tropospheric temperature adjustment to radiative forcing is important. We refer to these precipitation, P, responses as the 'fast', $\Delta P_{\rm fast}$, and 'slow', $\Delta P_{\rm slow}$, responses respectively. During transient climate change experiments $\Delta P_{\rm slow}$ is proportional to global-mean surface-air-temperature change, ΔT. The constant of proportionality, α (in units % K^{-1}), measures the percentage change in precipitation per unit of global-mean surface-air-temperature change. Thus a change in global-mean precipitation, ΔP, can be expressed as the sum of the fast and slow responses, $\Delta P = \Delta P_{\rm fast} + \Delta P_{\rm slow}$, and so,

$$\Delta P = \Delta P_{\rm fast} + \alpha \Delta T. \qquad (1)$$

It is the purpose of this letter to evaluate the fast and slow precipitation responses to increasing, 1% yr^{-1}, CO_2 levels in fully coupled atmosphere–ocean general circulation models (AOGCMs) and compare this to a period of stabilization. This scenario is more representative of real world CO_2 increases (in comparison to instantaneous CO_2 doubling experiments) where both changes in radiative forcing and ΔT will occur at the same time, and so separating the fast and slow responses will be difficult as they will both evolve together. In addition, we anticipate that accounting for the fast response may shed light on why Allen and Ingram (2002) noticed that the relationship between ΔP and ΔT was different between transient experiments at the point of CO_2 doubling and those at equilibrium. Section 2 presents the model data, section 3 presents the results and section 4 discusses the conclusions.

Table 1. The transient hydrological sensitivity, κ, and the differential hydrological sensitivity, α, (in % K^{-1}) for various AOGCMs. α represents the 'slow' precipitation response to global-mean surface-air-temperature change while κ represents the combined 'slow' and 'fast' precipitation response (see text) to increasing CO_2 levels as they evolve together. κ and α are diagnosed from the gradient of the change in global-mean precipitation rate against global-mean surface-air-temperature change during the 70 years in which CO_2 levels are increased by 1% yr^{-1} and during the stabilization period where CO_2 levels are held at twice their unperturbed value respectively (see figure 2). Uncertainties represent the 1-σ uncertainty from the regression.

Model	κ	α
CCSM3	1.77 ± 0.07	2.60 ± 0.14
CGCM3.1(T47)	1.60 ± 0.06	2.91 ± 0.25
CNRM-CM3	1.34 ± 0.08	2.60 ± 0.09
GFDL-CM2.0	1.38 ± 0.10	1.51 ± 0.13
GFDL-CM2.1	1.08 ± 0.09	1.01 ± 0.17
GISS-EH	2.30 ± 0.06	1.96 ± 0.07
INM-CM3	1.62 ± 0.09	2.15 ± 0.11
IPSL-CM4	1.66 ± 0.06	3.29 ± 0.09
MIROC3.2(medres)	1.69 ± 0.05	2.40 ± 0.10
ECHO-G	0.76 ± 0.06	1.70 ± 0.18
ECHAM5-MPI/OM	1.79 ± 0.07	2.61 ± 0.11
MRI-CGCM2.3.2a	2.18 ± 0.05	3.57 ± 0.15
UKMO-HadGEM1	1.14 ± 0.05	2.07 ± 0.06
AOGCM-mean	1.53 ± 0.02	2.40 ± 0.04

2. Climate model data

Climate model data was taken from the World Climate Research Programme's (WCRP) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset. This large database archives numerous AOGCM simulations: here we make use of the CO_2 doubling scenario. Starting from a control run (usually, but not always, based on pre-industrial conditions) CO_2 was increased at a rate of 1% yr^{-1} for 70 years, at which point CO_2 levels are then held constant at twice their unperturbed levels for a further 150 years. We examined all of the models that contributed to the CMIP3 database; only 13 had the sufficient 220 years of relevant data and corresponding control runs. The 13 models are listed in section 3 (see table 1) and are referred to by their official CMIP3 name. For details of individual models see the online model documentation (www-pcmdi.llnl.gov/ipcc/model_documentation/ipcc_model_documentation.php). Note that Sun *et al* (2007) provide a detailed analysis of the CMIP3 model simulated changes in precipitation, evaporation and water vapour under a range of different emission scenarios for the 21st century. For each model, surface-air-temperature and the precipitation rate were extracted. Changes in these terms were calculated by subtracting corresponding linear fits of the control integration from the 1% yr^{-1} CO_2 increase experiment. In the following analysis all results are based on annual and global averages. Each AOGCM contributed equally to the AOGCM-mean.

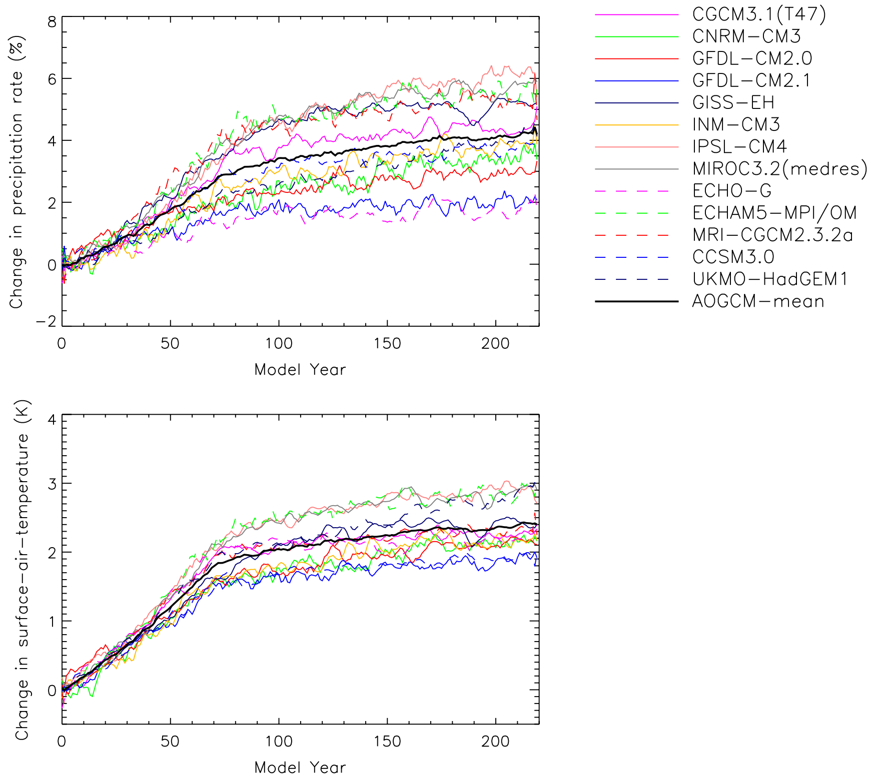

Figure 1. Time series of the change in global-mean precipitation rate (in %) and global-mean surface-air-temperature change (in K) for the 1% yr^{-1} CO_2 increase experiment compared to the control state for various fully coupled CMIP3 climate models. Note that after the 70th year CO_2 levels are double those in the control simulation, and subsequently held constant for the rest of the run. A five-year boxcar smoothing has been applied.

3. Results

Figure 1 shows the time series of the change in global-annual-mean precipitation rate (in %), the ΔP term in equation (1), and ΔT for all of the models and the AOGCM-mean. Throughout the 220 years of integration both global-mean surface-air-temperature and precipitation increases. The rate of these changes is greatest during the first 70 years, during which CO_2 levels are increasing. After the 70th year CO_2 levels are held constant but ΔT, and so precipitation, continue to respond to the forcing due to thermal lag, created by the large heat capacity of the oceans.

ΔP and ΔT appear to follow the same overall trend. Figure 2 shows a strong correlation between the two, but the relationship between them changes during the experiment. After the 70th year the points lie on a straight line with gradient $\sim$2.4% K^{-1}, but during the first 70 years the points lie on a straight line with gradient $\sim$1.5% K^{-1}. A similar difference was also noticed by Allen and Ingram (2002) in an older set of models (CMIP2). We now investigate the reason for this change in behaviour.

After the 70th year the forcing is constant. Therefore, assuming that ΔP_{fast} does not change (a reasonable assumption given the observed linearity and the short timescale of atmospheric adjustments to forcings), the gradient of ΔP as a function of ΔT represents the slow response of precipitation to ΔT, the α term in equation (1). This term, which we refer to as the 'differential hydrological sensitivity' (Andrews *et al* 2009), represents an increase in precipitation with positive ΔT; AOGCM-mean equals 2.40 ± 0.04% K^{-1}. The individual model results are listed in table 1. There is good agreement across the models of a value of the order

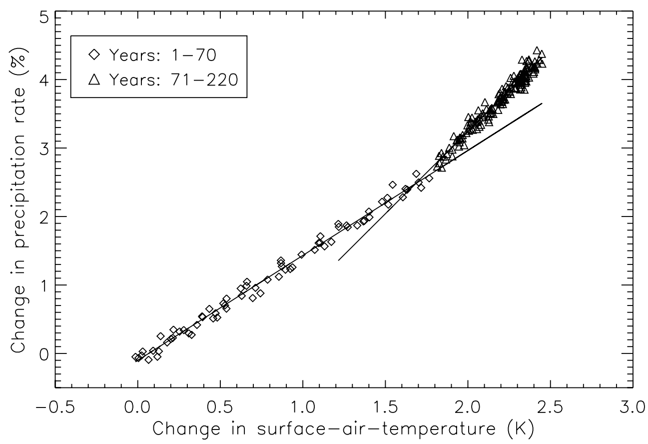

Figure 2. Relationship between the change in global-mean precipitation rate (in %) and global-mean surface-air-temperature change (in K) for the AOGCM-mean. Points are annual-global-means. Diamonds correspond to the first 70 years of integration, during which CO_2 is ramped up at 1% yr^{-1} and triangles correspond to the remaining 150 years during which CO_2 is held constant at twice its unperturbed level. Solid lines correspond to linear fits through the first 70th years and remaining years.

$\sim$2.4% K^{-1}, although GFDL-CM2.1 and GFDL-CM2.0 have particularly weak responses, $\sim$1.0% K^{-1} and $\sim$1.5% K^{-1} respectively, while MRI-CGCM2.3.2 has a particularly strong response, $\sim$3.6% K^{-1}. Previous studies (e.g. Lambert and Webb 2008, Andrews *et al* 2009) have diagnosed the response of precipitation to ΔT using a similar regression technique in models with a thermodynamic mixed-layer ocean component, as opposed to a fully dynamic ocean used here, whose responses may not necessarily be the same (Boer and Yu 2003). For the relevant models we find no systematic difference in the precipitation response to ΔT between the mixed-layer and fully dynamic ocean models (although the sample of relevant models is small). In fact our model ensemble-mean lies in the middle of the 1.4–3.4% K^{-1} range determined by Lambert and Webb (2008) and is in agreement with the model ensemble-mean determined by Andrews *et al* (2009).

During the first 70 years we also observe a linear relationship between ΔP and ΔT, but on a different slope to α (figure 2 and table 1). We refer to this constant of proportionality as the 'transient hydrological sensitivity' (see below), termed κ, so that,

$$\Delta P = \kappa \Delta T. \tag{2}$$

During this period CO_2 levels are increasing and so κ represents both the fast and slow response of precipitation as they evolve together. In other words, in the absence of any fast response to CO_2 it would take a value α due to its response to ΔT, but as CO_2 levels are increasing (and so inducing cumulative fast responses) it forces the response onto a different path (κ diverges from α).

The utility of κ is limited; it can only apply during the time period in which CO_2 levels are increasing. It is analogous to the proportionality between the global energy imbalance and ΔT, the 'ocean heat uptake efficiency' (see Gregory and Mitchell 1997, Raper *et al* 2002, Gregory and Forster 2008). Yet it is useful for predicting the precipitation response during increasing CO_2 radiative forcing, a scenario relevant for real world prediction. Table 1 lists the individual results for κ, as diagnosed from the models. In most cases κ is significantly smaller than α (table 1, AOGCM-mean $\sim$1.5% K^{-1} compared to $\sim$2.4% K^{-1}, respectively) and so the fast response to CO_2 is to suppress the precipitation response to ΔT. However, for the GFDL models, α and κ are indistinguishable (table 1), and GISS-EH is particularly anomalous in that κ is larger than α. The reason for this different behaviour is unclear.

Given that ΔP is proportional to ΔT during the time period in which the fast response is changing this suggests that it is also proportional to ΔT (assuming α is constant). Substituting equation (2) into (1) gives,

$$\Delta P_{\text{fast}} = (\kappa - \alpha)\Delta T. \tag{3}$$

The fast precipitation response can therefore be calculated during the first 70 years of the model experiments according to this equation, see figure 3. At the point of CO_2 doubling, year 70, precipitation is suppressed by $\sim$1.5% according to the multi-model mean due to the fast response (figure 3). This result can be compared to those of Andrews *et al* (2009) who diagnosed ΔP_{fast} due to an instantaneous doubling of CO_2 in models with a thermodynamic mixed-layer ocean component. A multi-model mean comparison suggests that the fast precipitation response to CO_2 forcing may be slightly smaller in the fully coupled AOGCMs rather than their thermodynamic mixed-layer counterparts, model ensemble-means of $\sim$−1.5% and $\sim$−2.5% respectively, but the qualitative responses are similar. Alternatively it could suggest the fast response is not fully realized at the point of CO_2 doubling in the transient experiments because the timescale of the fast response is longer in the fully coupled AOGCMs (see below).

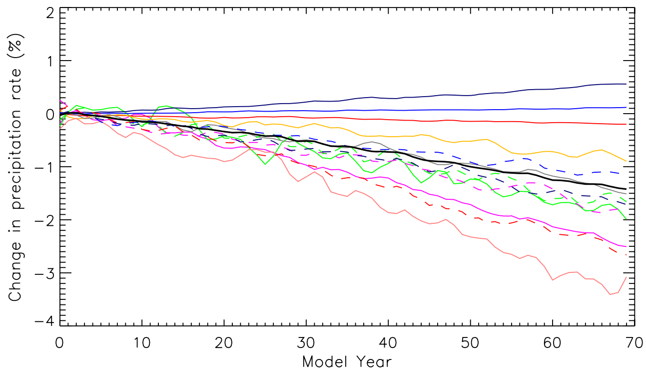

Figure 3. Diagnosed time series of the fast precipitation response (in %) during the years in which CO_2 levels are increased by 1% yr^{-1} for various fully coupled CMIP3 climate models. Lines are the same as in figure 1. A three-year boxcar smoothing has been applied.

The timescale in which the response of precipitation turns from the transient to the differential hydrological sensitivity, the kink in figure 2, depends on the timescale of the fast response. If the fast precipitation response occurs almost simultaneously with the change in CO_2, i.e. days to weeks, as suggested by Dong *et al* (2009), then on the multi-annual timescale considered here the change in response will be immediate after the 70th year, when the CO_2 forcing is stabilized and the kink in figure 2 is more pronounced. If, however, the fast precipitation response to the increasing CO_2 levels is only realized after a few decades, perhaps due to a forcing dependent response in the ocean (Williams *et al* 2008), then the kink will be smoothed out over a longer time period. Inspection of figure 2 suggests that the transition is sharp, but as CO_2 is only increasing by 1% yr^{-1} the forcing is probably not large enough to make a conclusion. A full analysis would require a large step change in forcing, such as an instantaneous quadrupling of CO_2, this would also allow a detailed analysis of the individual model results as the signal-to-noise ratio would be much larger.

4. Discussion and conclusions

We have examined the transient change of global-mean precipitation in response to a steadily increasing forcing scenario, 1% yr^{-1} increase in atmospheric CO_2 levels, in fully coupled AOGCMs. Results show that the change in global-mean precipitation rate is proportional to ΔT, but the relationship is different between results when the forcing is increasing or held constant. When the forcing is held constant the models suggest that the precipitation rate intensifies with ΔT at a rate of the order $\sim$2.4% K^{-1}, in line with previous estimates. During the time period of increasing forcing this response is suppressed by a fast atmospheric response to the increasing CO_2 radiative forcing, to $\sim$1.5% K^{-1}. We refer to the two proportionality factors as the 'differential' and 'transient' hydrological sensitivities respectively.

The differential hydrological sensitivity applies at all times; it represents the precipitation response to ΔT and should be independent of the forcing scenario (Andrews *et al* 2009, Bala *et al* 2009). In contrast, the transient hydrological sensitivity applies only to a scenario of increasing CO_2 radiative forcing, it represents the sum of the fast atmospheric response to CO_2 and its indirect effect through ΔT on precipitation as they both evolve together. It could be useful for predicting global-mean precipitation changes over a timescale of decades, when CO_2 radiative forcing is increasing. For example, Gregory and Forster (2008) observed a linear relationship between steadily increasing top-of-atmosphere/tropopause CO_2 radiative forcing, F, and ΔT, so that $F = \rho\Delta T$, where ρ is the 'climate resistance' in units W m^{-2} K^{-1}. Replacing ΔT in equation (2) we find,

$$\Delta P = \frac{\kappa}{\rho}F. \tag{4}$$

Hence, for increasing CO_2 levels, given the transient hydrological sensitivity and the climate resistance, the response of global-mean precipitation can be predicted from knowledge of the CO_2 radiative forcing alone.

Separating the fast and slow responses has applications to predicting time-dependent climate change (Gregory and Webb 2008, Williams *et al* 2008, Andrews 2009). However, in coupled transient climate change simulations, where both the radiative forcing and global surface-temperature change at the same time, the fast and slow responses will evolve together and are not easy to separate. According to the multi-model mean, stabilizing CO_2 radiative forcing would lead to a greater rate of precipitation change per unit surface warming for years to come. However, some models, namely the GFDL and GISS models, show little change in the relationship between precipitation changes and global surface-temperature change. In future research it would be interesting to investigate why the precipitation responses in the GFDL and GISS models are different.

Finally, this study has only evaluated the precipitation response to CO_2. Other forcing agents, such as other greenhouse gases and different species of aerosols, are also expected to influence precipitation. In particular, aerosols have a strong influence on the amount of solar radiation absorbed by the Earth's surface (e.g. Ramanathan *et al* 2001, Wild 2009), which is a driver of evaporation. Therefore, it would be useful if future research focused on evaluating the response of precipitation to many different forcing agents, as this study has done for CO_2.

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