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 \(\sim 2.4\%\) 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 \(\sim 1.5\%\) 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, Lambert and Webb 2008). This response is somewhat smaller than some recent observations (\(\sim 7\%\ K^{-1}\)) but still consistent when interdecadal variability is considered (Liepert and Previdi 2009).

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 \(\sim 2–3\%\ K^{-1}\) (Held and Soden 2006, Lambert and Webb 2008). This response is somewhat smaller than some recent observations (\(\sim 7\%\ K^{-1}\)) 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 \(\sim 7\%\ K^{-1}\) (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 CO₂, 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 CO₂ 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 CO₂ is therefore a combination of the response that scales with global-mean surface-temperature change and the response to tropospheric temperature adjustment to the CO₂ 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 CO₂ 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’, \( \Delta P_{\text{fast}} \), and ‘slow’, \( \Delta P_{\text{slow}} \), responses respectively. During transient climate change experiments \( \Delta P_{\text{slow}} \) is proportional to global-mean surface-air-temperature change, \( \Delta T \). The constant of proportionality, \( \alpha \) (in units \( \% \text{ K}^{-1} \)), measures the percentage change in precipitation per unit of global-mean surface-air-temperature change. Thus a change in global-mean precipitation, \( \Delta P \), can be expressed as the sum of the fast and slow responses, \( \Delta P = \Delta P_{\text{fast}} + \Delta P_{\text{slow}} \), and so,

\[
\Delta P = \Delta P_{\text{fast}} + \alpha \Delta T.
\]

It is the purpose of this letter to evaluate the fast and slow precipitation responses to increasing, 1% yr⁻¹, CO₂ 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 CO₂ increases (in comparison to instantaneous CO₂ doubling experiments) where both changes in radiative forcing and \( \Delta T \) 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 \( \Delta P \) and \( \Delta T \) was different between transient experiments at the point of CO₂ doubling and those at equilibrium. Section 2 presents the model data, section 3 presents the results and section 4 discusses the conclusions.

### 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 CO₂ doubling scenario. Starting from a control run (usually, but not always, based on pre-industrial conditions) CO₂ was increased at a rate of 1% yr⁻¹ for 70 years, at which point CO₂ 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 CO₂ increase experiment. In the following analysis all results are based on annual and global averages. Each AOGCM contributed equally to the AOGCM-mean.

| Model         | \( \kappa \) | \( \alpha \) |
|---------------|--------------|--------------|
| 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  |
3. Results

Figure 1 shows the time series of the change in global-annual-mean precipitation rate (in %), the $\Delta P$ term in equation (1), and $\Delta T$ 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 $\Delta T$, and so precipitation, continue to respond to the forcing due to thermal lag, created by the large heat capacity of the oceans.

$\Delta P$ and $\Delta T$ 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 $\Delta P_{\text{int}}$ does not change (a reasonable assumption given the observed linearity and the short timescale of atmospheric adjustments to forcings), the gradient of $\Delta P$ as a function of $\Delta T$ represents the slow response of precipitation to $\Delta T$, the $\alpha$ 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 $\Delta T$; AOGCM-mean equals $2.40 \pm 0.04\% \, K^{-1}$. The individual model results are listed in table 1. There is good agreement across the models of a value of the order
imbalance and analogous to the proportionality between the global energy and Webb 2008, Andrews et al 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 $\Delta P$ and $\Delta T$, but on a different slope to $\alpha$ (figure 2 and table 1). We refer to this constant of proportionality as the ‘transient hydrological sensitivity’ (see below), termed $\kappa$, so that,

$$\Delta P = \kappa \Delta T. \quad (2)$$

During this period CO$_2$ levels are increasing and so $\kappa$ 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 $\alpha$ due to its response to $\Delta T$, but as CO$_2$ levels are increasing (and so inducing cumulative fast responses) it forces the response onto a different path ($\kappa$ diverges from $\alpha$).

The utility of $\kappa$ 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 $\Delta T$, 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 $\kappa$, as diagnosed from the models. In most cases $\kappa$ is significantly smaller than $\alpha$ (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 $\Delta T$. However, for the GFDL models, $\alpha$ and $\kappa$ are indistinguishable (table 1), and GISS-EH is particularly anomalous in that $\kappa$ is larger than $\alpha$. The reason for this different behaviour is unclear.

Given that $\Delta P$ is proportional to $\Delta T$ during the time period in which the fast response is changing this suggests that it is also proportional to $\Delta T$ (assuming $\alpha$ is constant). Substituting equation (2) into (1) gives,

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

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 $\Delta P_{\text{fast}}$ 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).

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 $\Delta T$, 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 $\Delta T$ at a rate of the order $\sim2.4\% \text{ K}^{-1}$, 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 $\sim1.5\% \text{ K}^{-1}$. 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 $\Delta T$ 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 $\Delta T$ 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 $\Delta T$, so that $F = \rho \Delta T$, where $\rho$ is the ‘climate resistance’ in units W m$^{-2}$ K$^{-1}$. Replacing $\Delta T$ in equation (2) we find,

$$\Delta P = \frac{k}{\rho} F.$$  

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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