Fast-Forward to Perturbed Equilibrium Climate

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Fast forward to perturbed equilibrium climate

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Key Points:

- A simple method for estimating the equilibrium climate sensitivity is proposed.
- The method allows to simulate the stationary climate corresponding to any given radiative perturbation with a limited computational cost.
- The method can be applied to any atmosphere-ocean coupled climate model.

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Abstract

The equilibrium climate sensitivity, i.e. the global-mean surface temperature change in response to a doubling of the carbon dioxide concentration is a widely used metric in climate change studies. Its exact value is rarely known because its estimation requires a long integration time of several thousand years. We propose a method to estimate an accurate value of the equilibrium response from fully coupled climate models at a reasonable computational cost. Using this method, our state-of-the-art climate model CNRM-CM6-1 reaches a stationary state after only few hundred of years of integration. This 'Fast-Forward' method consists of an optimal two-step forcing pathway designed using the framework of a two-layer energy-balance model. It can be applied easily to any coupled climate model.

1 Introduction

The equilibrium climate sensitivity (ECS) is commonly defined as the stationary-state global mean surface-air temperature change in response to a doubling of the atmospheric carbon dioxide (CO_2) concentration relative to the pre-industrial era. The ECS is a widely used metric to characterize the magnitude of global warming projected by climate models. Its contribution to the uncertainty of transient warming is preponderant and significantly larger than the contribution of ocean thermal inertia (e.g. Dufresne & Bony, 2008; Geoffroy et al., 2012). Moreover, many climate variable responses scale, at least partly, with the magnitude of global warming across multiple scenarios or across time in a given scenario (e.g. Geoffroy & Saint-Martin, 2014; Pfahl et al., 2017; Ceppi et al., 2018).

Originally, the ECS was computed using atmospheric models coupled with a simple thermodynamic mixed-layer ocean (Stouffer & Manabe, 1999). Such estimates may differ from those obtained from complete climate models that take oceanic transport into account. The use of a fully-dynamic ocean requires more than 3000 years of simulation to reach a stationary state (Stouffer, 2004), which explains the scarcity of studies documenting such millennial-length experiments (e.g. Danabasoglu & Gent, 2009; Jonko et al., 2012; Li et al., 2013; Paynter et al., 2018). Hence, despite the fact that ECS is a common metric of climate change studies, the true ECS of climate models is rarely known.

Gregory et al. (2004) proposed an alternative method to estimate the ECS of a model from centennial simulations. This method relies on the assumption that, in response to an externally imposed radiative perturbation $\mathcal{F}$, the top-of-the-atmosphere radiative net flux change ΔN evolves linearly with the global-mean surface-air temperature change ΔT , $\Delta N = \mathcal{F} - \lambda \Delta T$, where λ is the radiative feedback parameter. Using this method, the forcing and the feedback parameter can be estimated as the coefficients of the linear regression between ΔN and ΔT , and ECS can be obtained by extrapolating the response to $\Delta N = 0$. This linear regression is often applied to the **abrupt-4 $\times$ CO₂** experiment (Andrews et al., 2012), i.e. a 150-year simulation where the atmospheric CO₂ concentration is quadrupled instantaneously, while the initial condition is taken from a pre-industrial simulation. Under the same assumptions, the full transient evolution of energy imbalance ΔN and temperature change ΔT can be computed using a simple two-layer energy balance model (EBM) calibrated with the climate-model **abrupt-4 $\times$ CO₂** simulation (Held et al., 2010; Geoffroy, Saint-Martin, Olivié, et al., 2013, hereafter H10 and G13, respectively). Such estimates of ECS rely on the assumption that the feedback parameter λ is constant. However, λ may vary in magnitude as time goes by, due to changing sea-surface temperature patterns along the course of a transient warming (Winton et al., 2010; Geoffroy, Saint-Martin, Bellon, et al., 2013). Despite this limitation, the linear regression method has until now provided the most commonly used estimates of ECS for a large number of climate models.

In this paper, we present a simple and elegant method (‘Fast-Forward’ method) to estimate the ECS from fully coupled atmosphere-ocean general circulation models (AO-GCMs) from short simulations of a few hundred years only. More generally, this method allows to simulate the quasi-stationary climate corresponding to any given radiative perturbation. We design an optimal forcing scenario to minimize the simulation time needed to reach stationary state. Recently, Sanderson et al. (2017) designed a set of scenarios in order to achieve long-term +1.5 K and +2 K global-mean temperatures in a stable climate. However, they did not test the stationarity at multi-century timescales. Moreover, their protocol is not easily portable to other AO-GCMs, in particular because of the complexity of the forcing pathways. Our method overcomes these limitations by providing, for each warming target, simple forcing pathways, which can be easily computed by any climate modelling center participating in the Coupled Model Intercomparison Project - phase 6 (CMIP6; Eyring et al., 2016). The only coordinated experiment needed to im-

plement our Fast-Forward method is a 150-year long **abrupt-4×CO₂** experiment, which is required within CMIP6. Using the EBM framework calibrated on this experiment, it is possible to derive an optimized forcing pathway to reach long-term equilibrium and accurately and efficiently estimate the ECS of the corresponding AO-GCM.

2 Methods and experimental setup

Two-layer EBMs can be used to emulate the global-mean surface-air temperature response of a given fully coupled AO-GCM to an externally imposed radiative perturbation. In this framework, the climate system can be simply described by 5 parameters: 2 radiative parameters - the forcing reference (such as $\mathcal{F}_{4\times}$, the net radiative forcing associated with a quadrupling of the atmospheric CO₂ concentration) and the feedback parameter λ - and 3 thermal-inertia parameters: the first-layer specific surface heat capacity C , the second-layer (deep-ocean) specific surface heat capacity C_d and the heat exchange coefficient between the two layers γ (Gregory, 2000, H10, G13).

In response to any radiative perturbation $\mathcal{F}(t)$, the temperature responses of the two layers are the sum of the balanced temperature $\Delta T_{eq} = \mathcal{F}(t)/\lambda$ and two modes characterized by distinct timescales, τ_f (fast) and τ_s (slow). The relative contributions of the two modes are quantified by parameters (a_f , a_s , ϕ_f and ϕ_s) depending on C , C_0 , γ and λ , for which expressions are given in Table 1 of G13. Assuming that $C \ll C_0$, the values of the timescales are approximately equal to $\tau_f = C/(\lambda+\gamma)$ and $\tau_s = (\lambda+\gamma)C_0/(\lambda\gamma)$ (H10). G13 proposed a calibration method to derive the five EBM parameters from an AO-GCM step-forcing experiment, typically the experiment **abrupt-4×CO₂** carried out by all climate modelling centers participating in the CMIP6 intercomparison project.

In a step-forcing experiment corresponding to a CO₂ concentration increase of n_∞ times the pre-industrial carbon dioxide concentration, $n_\infty = [\text{CO}_2]_\infty/[\text{CO}_2]_{pi}$, the radiative perturbation is well approximated by $\mathcal{F}(t < 0) = 0$ and $\mathcal{F}(t \geq 0) = \mathcal{F}_\infty = \mathcal{F}_{4\times} \log_4(n_\infty)$. Stationary state ΔT_{eq} is obtained when the net radiative budget at the top of the atmosphere (TOA) is reduced to zero: $\Delta N = \mathcal{F}_\infty - \lambda \Delta T_{eq} = 0$. Within the EBM framework, it is possible to compute the time necessary for the deep-ocean temperature response to reach α percent of its equilibrium change. This stabilization time (t_α) is independent of the magnitude of the forcing and is approximately equal to: $t_\alpha \approx \tau_s \ln[(\phi_s a_s)/(1-\alpha)]$ (see detailed derivations in the Supplementary Information). In the

EBM framework, estimated values for stabilization time are of the magnitude of thousands of years.

This stabilization time can be significantly reduced through the use of a two-step forcing pathway. The idea is simply to initially impose a radiative forcing that is stronger than the target forcing in order to warm the deep ocean faster. The strong forcing is maintained until the target equilibrium of the deep ocean is reached (see Fig. S1). Once this equilibrium is reached, the forcing is revised downwards to the target forcing. By successively imposing an initial forcing n_0 higher than the target forcing n_∞ ($n_0 > n_\infty$) followed by the target forcing n_∞ for the remainder of the simulation, the global-mean surface temperature change will tend to the equilibrium temperature response ΔT_{eq} faster than if the target forcing was applied all along. The optimal duration t_0 of the initial forcing is linked to the amplitude of the initial forcing through the equation (see Supplementary Information for details of derivation):

$$t_0 = -\tau_s \ln \left[1 - \frac{\ln(n_\infty)}{\ln(n_0)} \right]. \quad (1)$$

By imposing a 'Fast-Forward', two-step forcing scenario, $[\text{CO}_2](t < t_0) = n_0[\text{CO}_2]_{pi}$ and $[\text{CO}_2](t \geq t_0) = n_\infty[\text{CO}_2]_{pi}$, the time to reach the n_∞ stationary state is approximately equal to $t_0 + \tau_f$. This is much shorter than the stabilization time $t_{0.99}$ necessary in a step-forcing experiment. For example, in the case of the CMIP5 multi-model mean, with a Fast-Forward experiment, the time to reach a $n_\infty = 2$ stationary state is approximately equal to 180 years. This is an order of magnitude smaller than the stabilization time for an **abrupt-2×CO2** experiment. In Eq. (1), it is possible to tune either the duration, t_0 , or the amplitude, n_0 , of the initial forcing. The time t_0 can be chosen to be as small as possible, but then the initial forcing and the temperature change at t_0 can be very large. To avoid spurious threshold effects, a compromise to impose a reasonable initial forcing is desirable. Here we propose to use the pre-existing 150-year long **abrupt-4×CO2** simulation as the initial part of the two-step forcing pathway and to set $n_0 = 4$. If the optimal duration time is smaller than the duration of the pre-existing simulation ($t_0 < 150$ years), the stationary state can be obtained after only a few years of simulation. Note that, for a target forcing of $n_\infty > 4$, performing an additional simulation with $n_0 > 4$ will be necessary, since no experiment with such a large forcing is available in the CMIP6 dataset.

If t_0 exceeds 150 years, the existing **abrupt-4×CO2** needs to be extended to t_0 in the two-step forcing pathway. An alternative method is to apply an optimal two-step forcing pathway from the end of the pre-existing **abrupt-4×CO2** simulation. In this case, after imposing an initial forcing $n_0 = 4$ during a duration of $t_0=150$ years, we successively impose an intermediate forcing n_m until t_m , followed by the stationary-state forcing $\mathcal{F}_\infty$ for the remainder of the simulation. Identically to the two-step forcing pathway, an optimal t_m can be determined to minimize the stabilization time in this three-step forcing pathway (see Supplementary Information for details of the derivation).

Finally, it is also possible to achieve a Fast-Forward stabilization of the global mean surface-air temperature by imposing an exponentially decreasing forcing (see Supplementary Information for details of derivation). Within this exponential pathway, by optimally choosing the decay time of the exponential forcing $\tau_e = C_0/\gamma$ and the amplitude of the initial forcing $n_e = e^{\tau_s/\tau_e} n_\infty$, the temperature adjustment of the first layer is very fast, with timescale τ_f , but the TOA radiative imbalance decays more slowly, with timescale τ_e , so even though the temperature is adjusted after a short time, the slowly adjusting components of the climate system are not in equilibrium until much later.

We test the different pathways of our method described above with the Centre National de Recherches Météorologiques’ AO-GCM, CNRM-CM6-1 (<http://www.umr-cnrm.fr/cmip6/references>). The parameters of the surrogate two-layer EBM were estimated from the 150-year CMIP6 **abrupt-4×CO2** experiment, following the method described in G13. In particular, the value of τ_s is computed by linear regression of $\ln(\Delta T_{eq} - \Delta T)$ against t over the 100-year period spanning from year 51 to year 150. Results are summarized in Table S1. The estimated value of the slow timescale is $\tau_s = 415$ years. In CNRM-CM6-1, the relative contribution of the slow mode is quantified by $a_s = 0.43$ and $\phi_s = 2.35$. As a result, the 99% stabilization time of a step-forcing experiment is $t_{0.99} = 1925$ years for this model.

As a reference, the CMIP6 **abrupt-2×CO2** experiment is extended from 150 to 750 years. Note that this experiment contributes to the Cloud Feedback Model Intercomparison Project (CFMIP; Webb et al., 2017). A Fast-Forward experiment with the two-step pathway (FF-2×CO2) was performed for $n_\infty = 2$. The value of the initial forcing was set to $n_0 = 4$, in order to use the existing **abrupt-4×CO2** simulation. In this case, the value of t_0 is equal to 287 years. For the target CO₂-doubling concentration ($n_\infty =$

2), we performed two additional Fast-Forward experiments: **expo-2×CO2**, with the exponential pathway with $n_e = 3.34$ and $\tau_e = 238.2$, and **FF-2×CO2-3step**, with the three-step pathway using 150 years of the **abrupt-4×CO2** experiment ($n_0 = 4$; $t_0 = 150$ years), an intermediate forcing with $n_m = 8$ and $t_m = 224$ years, and $n_\infty = 2$. Each of these Fast-Forward experiments was carried out for at least 400 years. The simulated climate states are analyzed as deviations from the model’s unperturbed climate state, as simulated by the first 500 years of the CMIP6 piControl experiment. The complete set of experiments is summarized in Table S2. Over all, four types of experiments are available to estimate the ECS (i.e., 2×CO₂ equilibrium): the abrupt forcing **abrupt-2×CO2**, the two-step forcing **FF-2×CO2**, the three-step forcing **FF-2×CO2-3step** and the exponential forcing **expo-2×CO2**. The forcing pathways of these four experiments are plotted in Fig. 1a.

3 Results

Figure 1b shows the temporal evolution of the annual-mean global-mean surface-air temperature response ΔT in all the step-forcing and Fast-Forward experiments carried out with CNRM-CM6-1. As predicted by the EBM framework, in the Fast-Forward experiment **FF-2×CO2**, the surface-air temperature response reaches equilibrium after a few dozen years following the end of the initial forcing ($t_0 = 287$ years). In the case of CNRM-CM6-1, for a target forcing of $n_\infty = 2$, the stabilization is effective after about 400 years. In the 3-step forcing experiment (**FF-2×CO2-3step**), the surface temperature response reaches quasi-stationary state after an even smaller duration, of about 350 years. Because of the large peak warming above the target equilibrium (of roughly 12 K), the three-step pathway might fail in some models due to hysteresis effects. In the absence of dynamic ice sheets or dynamic vegetation, CNRM-CM6-1 does not exhibit such effects. Similar results were obtained in the ‘recovery’ experiments of H10. In the Fast-Forward exponential-forcing simulation (**expo-2×CO2**), the surface temperature response is close to its long-term mean temperature response after only three decades. The ECS values predicted by these Fast-Forward experiments lie in the range of 4.2 K to 4.4 K, very close to the equilibrium temperature response estimated from the 150-year linear regression, $\Delta T_{eq}^{2\times} = 4.3$ K.

The joint evolution of ΔT and ΔN is plotted in Fig. 2 for all experiments. It approximately follows the EBM prediction $\Delta N(t) = \mathcal{F}(t) - \lambda \Delta T(t)$. In all step-forcing

pathways and the abrupt experiment, ΔN decreases linearly with ΔT , with an intercept at $\Delta T = 0$ that depends on the imposed CO_2 concentration, and with a slope close to $-\lambda$ estimated by linear regression using the **abrupt-2 $\times$ CO2** experiment. In experiments **FF-2 $\times$ CO2** and **FF-2 $\times$ CO2-3step**, ΔN becomes negative when the CO_2 concentration is reduced to the target value 2 $\times$ CO₂ (at t_0 or t_m); ΔN and ΔT subsequently relax to their equilibrium values following the same linear relationship. In the Fast-Forward exponential-forcing simulation (**expo-2 $\times$ CO2**), after reaching the surface stationary state (ΔT quasi-constant), the radiative response ΔN remains positive and exponentially decreasing, as predicted by the EBM: in that phase of the simulation, by design, the TOA radiative imbalance is entirely transferred to the deep ocean. After 750 years of integration, the surface-air temperature response in the **abrupt-2 $\times$ CO2** experiment is close to the extrapolated equilibrium value, $\Delta T_{eq}^{2\times}$. However, this experiment is not yet at equilibrium. It still has a positive net radiative TOA budget ΔN at the end of the simulation. Moreover, after 400 years, corresponding to the time needed for the Fast-Forward experiments to reach stationary state, the mean tendency of the 2000-m ocean heat content is about three times larger in the **abrupt-2 $\times$ CO2** experiment than in the Fast-Forward experiments (not shown).

In summary, the Fast-Forward experiments reach equilibrium after only a few hundred of years. If we assume that the 150-year **abrupt-4 $\times$ CO2** experiment already exists, the ECS of the fully coupled AO-GCM can be estimated from an additional 250-year simulation, at least four times less than the thousand(s) years needed to reach equilibrium by simply extending the step-forcing **abrupt-2 $\times$ CO2** experiment. The time needed to reach equilibrium at smaller values of CO_2 concentration is almost negligible, approximately an additional decade of integration (not shown).

However, CNRM-CM6-1 does not behave exactly as the EBM predicts. Indeed, experiment **FF-2 $\times$ CO2** still has a positive net radiative TOA budget ΔN at the end of the simulations. Likewise, the temperature response in experiment **expo-2 $\times$ CO2** presents an overshoot before reaching its stationary state. These two features suggest that the slow timescale τ_s is underestimated by the EBM calibration method. This value is also smaller than the typical timescales necessary to stabilize a fully coupled AO-GCM after an abrupt CO_2 doubling or quadrupling (e.g. Danabasoglu & Gent, 2009; Paynter et al., 2018). This would mean that the duration $t_0 = 150$ years of the initial **abrupt-4 $\times$ CO2** is indeed too short to properly estimate τ_s . If we use the first 287 years of the **abrupt-4 $\times$ CO2** exper-

iment to calibrate the EBM parameters, we find an estimate of $\tau_s = 530$ years for the slow timescale, significantly longer than the 415 years estimated from the first 150 years of the same simulation. This results in an estimate of $t_{0.99} = 2440$ years for the stabilization time at 99%. This estimate is closer to the empirical value derived from millennial-length experiments (e.g. Danabasoglu & Gent, 2009; Paynter et al., 2018).

With a better estimate of τ_s , it is likely that the Fast-Forward method would be more accurate. The advantages of extending experiment **abrupt-4 $\times$ CO2** longer than 150 years to better estimate τ_s have to be balanced with the Fast-Forward method's objective to limit computation time. Some sensitivity tests show however that the error in τ_s induces only a small error in the corresponding equilibrium response (see Fig. S2). More fundamental limitations might also contribute to the inaccuracies of the Fast-Forward method. In particular, the use of only two timescales to describe the ocean heat uptake might be questionable. The use of an improved version of the two-layer EBM with an efficacy for deep-ocean heat uptake (Geoffroy, Saint-Martin, Bellon, et al., 2013) could also yield better results.

Beyond the estimate of the global-mean temperature response, we can estimate the geographical distribution of the climate perturbation with the Fast-Forward method. Here, we define the equilibrium pattern as the zonally averaged and time-mean responses normalized by the global-mean equilibrium response for the same period of time. Figure 3 shows the surface-air temperature equilibrium patterns for the different Fast-Forward experiments. To highlight the interest of the Fast-Forward method, we compare the mean equilibrium pattern obtained after 350 years of integration in the three Fast-Forward experiments and in the **abrupt-2 $\times$ CO2** experiment (dotted lines). The long-term mean equilibrium pattern is estimated as the average over the last 40 years of the **FF-2 $\times$ CO2** experiment (solid red line).

Our results confirm the results of polar amplification of the equilibrium warming, in both the Arctic and the Antarctic. All the equilibrium patterns of the Fast-Forward experiments lie within the range predicted by the final period. The structure of the warming predicted is also very similar in the three Fast-Forward pathways, confirming the uniqueness of the equilibrium warming pattern and the absence of climate hysteresis in this AO-GCM. This also confirms that all Fast-Forward experiments converge towards the long-term equilibrium response in less than 400 years. On the other hand, the **abrupt-2 $\times$ CO2**

experiment is still far from equilibrium at year 350. Even at the end of the **abrupt-2×CO₂** experiment (after 750 years), the equilibrium pattern is still far from the equilibrium pattern (not shown). The 350-year **abrupt-2×CO₂** pattern (black dotted line) differs from equilibrium patterns mainly in the Southern ocean. These results are consistent with previous studies (e.g. Manabe et al., 1991; Geoffroy & Saint-Martin, 2014, H10)

4 Conclusion

By using the two-layer EBM framework, it is possible to design optimal forcing pathways to obtain a quasi-stationary state in an AO-GCM while minimizing the required computing resources. One optimal pathway is simply a two-step forcing scenario in which, before setting the target CO₂ concentration, a higher CO₂ concentration is imposed during a well-chosen period. The optimal duration of this period depends on the thermal inertia characteristics of the AO-GCM considered, which can be derived by calibrating the surrogate EBM parameters on an existing idealized experiment. Hence, this method can be easily applied to any AO-GCM.

Tests of this method using the state-of-the-art AO-GCM CNRM-CM6-1 are conclusive. Results from experiments at doubling of the CO₂ concentration show that the method performs well and that the model reaches its new equilibrium after about 350 years. Even with reasonable errors in the calibration of the EBM parameters, the model tends rapidly towards a quasi-stationary perturbed climate. However, a test with a single AOGCM is not sufficient to demonstrate that the method is generalizable and it would be interesting to test the method in an inter-comparison project. The lack of hysteresis effect in the current generation of climate models should guarantee the validity of the method for other AO-GCMs.

The main weakness of the method resides in an accurate estimation of the slow timescale, which is crucial to optimize the forcing pathway. A more complex adaptive method could be considered in the future. The forcing pathway could be changed interactively depending on the results of the first years of the Fast-Forward experiment. Another refinement of the method would be to take into account the effects of the surface warming pattern in the estimation of the slow timescale. But even with a poorly estimated slow timescale, the Fast-Forward method is a significant improvement over abrupt experiments. In the **abrupt-2×CO₂** experiment, the surface-air temperature reaches a value close to the ECS

in about twice the time required in the Fast-Forward experiments, but even at that time the stationary state is not reached, as the TOA energy budget and the temperature latitudinal pattern need more time to reach their equilibrium.

The Fast-Forward method provides an easily implemented and efficient framework to produce perturbed stationary climates at any level of carbon dioxide and at any temperature target (e.g. 1.5 K, 2 K). Such stationary-state simulations would be useful to quantify the state-dependency of climate sensitivity and to investigate the underlying mechanisms. They could also be helpful to understand and quantify regional impacts. Finally, they could be used to study the frequency of extreme climate events and the related societal impacts. The set of experiments provided by the Fast-Forward method can benefit other initiatives such as the international modelling efforts, HappiMip ('Half a degree Additional warming, Prognosis and Projected Impacts'; Mitchell et al., 2016) or nonlinMIP (Good et al., 2016).

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Figure 1. Temporal evolution of (a) CO_2 concentration in the step-forcing and Fast-Forward experiments and (b) corresponding global mean surface-air temperature responses (deviation from the temporal mean of the piControl experiment). The black circle denotes year 150 of the abrupt- $4\times\text{CO}_2$ experiment.

Figure 2. Scatterplot of the global mean surface-air temperature response (ΔT , K) and net TOA radiative imbalance (ΔN , W m⁻²) in the step-forcing and the Fast-Forward experiments (anomaly from the temporal mean of the piControl experiment). The black circle denotes year 150 of the abrupt-4xCO₂ experiment.

Figure 3. Pattern response of the zonal mean surface-air temperature in the step-forcing and the Fast-Forward experiments. For the FF-2×CO₂ (solid line), the equilibrium pattern response is calculated as the average over the last 50 years of the experiment. Plus/minus one interannual standard deviation is plotted as grey shading. The dotted lined corresponds to the pattern response evaluated as the 40-year mean centered over year 350.