Equilibrium response to carbon dioxide and aerosol forcing changes in a 1D air–sea interactive model

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Abstract

A 1D air–sea interactive model that couples an atmospheric column model with a slab ocean model was introduced. The model simulated an exact balance between radiative cooling and convective heating in the free troposphere. Within the planet boundary layer, the turbulent mixing produces cooling in the upper part and warming in the lower part, which is compensated by radiative and convective heating together. The model was then used to explore the equilibrium response of sea surface temperature (SST) and precipitation to carbon dioxide (CO2) and aerosol forcing changes. Results show a warming or cooling signal owning to increased CO2 or aerosols can be well captured by the ocean, leading to an increasing or decreasing of the SST, and hence the precipitation. Cutting off the interactions between atmosphere and ocean however renders different results. By applying the relaxed weak temperature gradient (WTG) approximation, the local response to aerosol forcing perturbations was investigated, which was shown to be largely different from the global response in spite of the same forcing. To understand the nonlinear interactions among different forcings, the equilibrium responses to multiple combinations of CO2 and aerosol forcings are analyzed.

Keywords: 1D model; carbon dioxide; aerosol; radiative-convective equilibrium; weak temperature gradient

1. Introduction

It is well established that external forcings, such as greenhouse gases, ozone, aerosols, etc., play an important role in modulating regional and global climate (Knutson and Manabe, 1995; Meehl et al., 2003). Knowing how the climate system responds to these forcings is important to understand current climate and produce believable predictions of the future. Typically, people resort to coupled ocean–atmosphere general circulation models (CGCMs) to numerically simulate the forcing effect (Murphy, 1995; Stott et al., 2000; Soden and Held, 2006; Andrews and Allen, 2008). Although CGCMs are state-of-the-art models that contain multiple components, the computational cost on running such models is dauntingly huge. Moreover, it is sometimes difficult to get conclusive results among these models, as GCMs usually disagree on the effects of some climate feedbacks. Alternatively, people choose to use atmosphere general circulation models (AGCMs) to simulate climate evolution in response to external forcings, with the sea surface temperature (SST) specified as a prerequisite (Li et al., 2007a, 2007b; He et al., 2012; Dong et al., 2015). Actually, SST being prescribed instead of projected implies the SST per se is an external forcing. The assumption behind is the time scale over which the atmosphere adjusts to the underlying surface is fast compared to the time scale that characterizes the adjustment of SST to changing conditions (Emanuel and Sobel, 2013). However, the long lasting forcings in the atmosphere may exert nonnegligible impacts on the ocean before their demise. In this sense, these simulations are far from a full climate response, but a fast atmospheric response to imposed forcings including SST.

If the global response to external forcings is primarily concerned, the 1D model suffices as well, i.e. the radiative-convective model (RCM), which dates back as early as to Manabe and Moller (1961). Although RCM represents the entire atmosphere by a single vertical column, it simulates in detail the transfer of energy through the depth of the atmosphere. Owning to its high computational efficiency, RCM has been widely used in scientific issues such as radiative-convective equilibrium (RCE), the role of convection, and tropical thermodynamics (Sobel and Bretherton, 2000; Sobel et al., 2007; Emanuel and Sobel, 2013).

Observations indicate a global warming trend during the last five decades, which is linked to increased greenhouse gases. However, a cooling trend is observed against this background of global warming in some regions, which is believed to be partly caused by the local increasing of aerosols (Xu et al., 2006). It is thus necessary to investigate the relative contribution of multiple external forcings to the climate system. Instead of...
using traditional AGCMs as in many previous studies (Li et al., 2007a; He et al., 2012), this study uses a 1D air–sea interactive model to explore such an issue, with the following questions primarily concerned: how SST and precipitation respond to different CO2 and aerosol forcing scenarios? Can the total response be approximated by linearly adding these individual responses? And what are the results if for a local response?

2. Model introduction

2.1. Atmospheric model

The atmosphere is represented by a single-column model that was originally described in Wang (2015). The governing equations for temperature and water vapor are expressed as:

\[
\frac{dT}{dt} = -\frac{\partial T}{\partial p} + \frac{R_d T}{C_p} + \frac{S_c + S_R + S_{\text{diff}}}{\rho} \tag{1}
\]

\[
\frac{dq}{dt} = -\frac{\partial q}{\partial p} + Q_c + Q_{\text{diff}} \tag{2}
\]

where \(\omega\) denotes surface velocity, \(S_c, S_R,\) and \(S_{\text{diff}},\) respectively, stands for convective heating, radiative cooling, and the convergence of turbulent heat flux. \(Q_c\) is convective moisture sink and \(Q_{\text{diff}}\) is the convergence of turbulent moisture flux. The model was employed with a relaxed convection scheme proposed by Emanuel (1991), the closure of which prompts convection quickly toward quasi-equilibrium with the large-scale forcing. The Planet Boundary Layer (PBL) turbulence is parameterized by a ‘non-local’ scheme, which calculates the eddy-diffusivity profile based on boundary-layer height and turbulent velocity scale (Holtslag and Boville, 1993). Radiative transfer process was calculated every 3 h using the Rapid Radiative Transfer Method for GCMs (RRTMG) package (Clough et al., 2005). Other physical parameterizations such as stratiform cloud condensation processes are switched off to exclude influences caused by clouds, which is of great uncertainty in GCMs (Zhang et al., 2005; Wang et al., 2012; Wang et al., 2014).

2.2. Ocean model

The sea is simplified as a slab ocean, with the mixed layer depth fixed at 40 m. Thus, SST can be readily computed according to the ocean heat budget equation:

\[
C_m \frac{dT_m}{dt} = F_{\text{rad}} - SH - LH \tag{3}
\]

where \(T_m\) stands for SST, \(F_{\text{rad}}\) is the net downward radiative flux and \(C_m = \rho_{\text{sea}} c_{\text{sea}}\) is the heat capacity of the slab ocean. The seawater density \(\rho_{\text{sea}}\) and heat capacity \(c_{\text{sea}}\) are set as 1025 kg m\(^{-3}\) and 3994.7 J kg\(^{-1}\) K\(^{-1}\), respectively.

Surface sensible and latent heat fluxes are calculated using the aerodynamic flux formulate, which are written as:

\[
SH = \rho_A V^* C_p C_H \Delta \theta \tag{4}
\]

\[
LH = \rho_A V^* L C_H C_m \Delta q \tag{5}
\]

where \(\rho_A\) is atmospheric surface density, \(V^*\) is the surface wind speed fixed at 5 m s\(^{-1}\), \(C_H\) is the drag coefficient specified as \(1.2 \times 10^{-3}\), \(\Delta \theta\) and \(\Delta q\) denotes the difference between surface layer and the lowest model level for potential temperature and water vapor, respectively.

Prognostic Equations (1)–(3), along with the two diagnostic Equations (4) and (5) constitute the skeleton of the 1D air–sea interactive model. The timestep of the model is set as 20 min. The model was integrated using a leapfrog time-differencing scheme combined with an Asselin time filter. The initial data comes from a typical sounding profile of tropics that is spaced every 25 hPa from 1000 hPa to a top of 5 hPa, which is also used as the model vertical coordinate.

Other configurations of the model are set as follows: sea surface albedo is 0.3, and the solar constant is 1360 W m\(^{-2}\) with a zenith angle of 71.8\(^\circ\) thus there is no diurnal and seasonal cycle in the simulations. The CO2 concentration is set as 360 ppm and varied under different warming scenarios. Other greenhouse gases such as CH4, N2O, CFC11, CFC12 are fixed at 1.72 ppm, 310 ppb, 280 ppt, and 484 ppt, respectively. Aerosols are switched off unless otherwise stated.

3. Model results

3.1. General performance

Prior to exploring SST and precipitation responses to external forcings, this section first exhibits the basic performance of the model.

3.1.1. RCE simulation

In this case, the vertical velocity was set to 0 and the model ran into radiative-convective equilibrium at Day 80. The final SST equilibrates at 30.04 °C and the precipitation rate is about 4.07 mm day\(^{-1}\). The sources and sinks of heat (solid lines) and moisture (dashed lines) due to individual physical process at the equilibrium state are shown in Figure 1(a). As expected, the radiative cooling is well balanced by the convective heating in the free troposphere, with a typical cooling/heating rate about 1–2 K day\(^{-1}\). Within PBL, the turbulent mixing yields a cooling tendency in the upper part and a warming tendency in the lower part, which is then largely offset by convective cooling. This convective cooling is caused by unsaturated cumulus downdrafts, which import low enthalpy of the middle troposphere into the subcloud layer, thus cooling and drying the air near the surface. The convective drying is then balanced by the turbulent moistening. Since turbulence is very weak in the free troposphere, the moisture sink due to convective condensation can only be compensated by
the moisture source due to convective-scale transport, resulting in null moisture sink and source above PBL.

The equilibrium thermal profiles are shown in Figure 1(b). In the lower troposphere, the moist static energy $h$ is much larger than the dry static energy $s$, owing to the presence of water vapor. The curve $h$ approaches saturated moist static energy $h^*$ at 950 hPa, which is close to the cumulus base.

### 3.1.2. WTG Simulation

In this case, the vertical velocity $\omega$ was not set as 0 but determined according to the relaxed weak temperature gradient (WTG) approximation, which was first advanced by Sobel and Bretherton (2000) and has been extensively studied since then (e.g. Wang et al., 2013; Romps, 2012; Raymond et al., 2015). The WTG approximation holds well in the tropics, as the Rossby deformation radius is so large that fast gravity waves can keep the temperature of the column the same as that of the surrounding environment (Emanuel and Sobel, 2013). Following Raymond and Zeng (2005), $\omega$ was calculated at each level above 850 hPa by assuming that the potential temperature relaxes to a reference profile in a time scale $T_{\text{relax}}$. In this study, the initial sounding is taken as the reference profile and $T_{\text{relax}}$ is set to 3 h as in Wang and Sobel (2011). Below 850 hPa, $\omega$ was determined by interpolating its value between 850 hPa and the surface, where $\omega$ was set to 0. The derived $\omega$ was then used to advect moisture. When a local forcing is imposed, the local response will lead to an atmospheric circulation that in turn modifies the response. For instance, the WTG approximation in the context of a single-column model can render a tropical response to abrupt aerosol increasing due to volcanic eruptions.

Figure 2 shows the time evolution of simulated SST, outgoing longwave radiation (OLR), precipitation, evaporation, and pressure velocity. The model tends to besteadyat Day 200. The precipitation increases in the first 10 days and then decreases abruptly, to a value of about 2.5 mm day$^{-1}$ at the equilibrium state. This is in accordance with the evolution of SST, which drops from the initial 30 °C to the final 25.6 °C. The evaporation declines at a relatively slow rate, with the final value of about 3.55 mm day$^{-1}$, in contrast to 4.07 mm day$^{-1}$ in RCE run. The imbalance between precipitation and evaporation under the WTG approximation is caused by the presence of large-scale subsidence (Figure 2(e)), which continuously dries out the column and acts as a moisture sink. Consequently, the water vapor was markedly decreased, leading to a significant increasing of OLR. Although deep convection is shown to be suppressed in this experiment, it does not mean the WTG approximation necessarily lead to weakened convections. In fact, multiple equilibria can be obtained depending on the humidity of initial atmosphere (Sobel et al., 2007; Sessions et al., 2010).

To explore the model sensitivity to relaxation time scale $T_{\text{relax}}$, we follow Wang and Sobel (2011) by varying $T_{\text{relax}}$ from 20 min to 2 days: 1/3, 1, 3, 6, 9, 12, 24, 36, and 48 h, which are shown in Figure 3(a). The result under strict WTG (Sobel and Bretherton, 2000) is also overlaid, corresponding to $T_{\text{relax}}=0$. There is a local minimum in the precipitation at $T_{\text{relax}}=1$ h. The precipitation then increases with $T_{\text{relax}}$ and approaches the value in RCE at $T_{\text{relax}}=48$ h, because the system must return to RCE as $T_{\text{relax}}$ goes to infinity. The precipitation at the smallest $T_{\text{relax}}$ is close to that under strict WTG. In fact, $T_{\text{relax}}$ modulates the interactive intensity between convection and WTG vertical velocity. As shown in Figure 3(b), the WTG vertical velocity is strongest in the zero limit of $T_{\text{relax}}$ and tends to be smaller when $T_{\text{relax}}$ becomes larger.

### 3.2. Response to external forcing changes

#### 3.2.1. CO$_2$ forcing

To mimic different warming scenarios, CO$_2$ raises from its control value of 360–2160 ppm by a successive

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Figure 1. (a) Heat (solid) and moisture (dashed) tendencies due to radiation (blue), convection (red) and turbulence (green) at the equilibrium state (unit: K day$^{-1}$); (b) dry (solid), moist (point) and saturated moist (dashed) static energy at the equilibrium state (unit: K).
Equilibrium response to CO$_2$ and aerosol forcings in a 1D model

Figure 2. Time evolutions of (a) SST (unit: °C), (b) OLR (unit: W m$^{-2}$), (c) precipitation (unit: mm day$^{-1}$), (d) evaporation (unit: mm day$^{-1}$) and (e) WTG vertical velocity (unit: mb h$^{-1}$) under the relaxed WTG approximation.

increasing of 360 ppm. Four experiments were carried out, which distinguished each other in the surface boundary conditions. The experiment ‘SST prognostic’ uses prognostic SST from the slab ocean model. ‘SST fixed’ keeps SST at its initial value. ‘SST&CO$_2$’ uses prescribed SST from the ‘SST prognostic’ run. ‘SST specified’ is similar to ‘SST&CO$_2$’ but fixes CO$_2$ at its control value. Each experiment reached steady before Day 200.

The SST and precipitation responses are shown in Figure 4. In the ‘SST prognostic’ experiment, the increased CO$_2$ leads to increased SST, and hence the increased precipitation because of the enhanced latent heat flux. In the ‘SST fixed’ experiment the precipitation decreases in spite of the warming atmosphere. This is because the SST keeps unchanged, thus the atmospheric instability decreases resulting in weakened convections. In the ‘SST specified’ experiment, the precipitation increases even faster than in ‘SST prognostic’. This is because the longwave cooling reinforces after CO$_2$ falling down to its control value, thus requiring more latent heating to balance the radiative

Figure 3. (a) Precipitation (unit: mm day$^{-1}$) versus the relaxation time scale and (b) WTG vertical velocity (unit: mb h$^{-1}$) under different relaxation time scale.
cooling. Varying CO₂ and SST simultaneously as in ‘SST&CO₂’ produces a precipitation response much close to that in ‘SST prognostic’ (The small differences are likely numeric artifacts.). This by itself is trivial, as the external forcings at the equilibrium state are the same in these two experiments. In fact, ‘SST&CO₂’ is a prototype of many experiments that use AGCMs to study climate response to external forcings. However, their results should be viewed cautiously if the SST specified in the AGCM deviates from the one projected in the CGCM under the same forcing.

3.2.2. Aerosol forcing

To reveal the impact of aerosol forcings on climate, the aerosol optical depth $\tau$ successively increases from 0 to 6 times of its averaged value in the tropics, which is based on aerosol mass dataset simulations from NCAR CAM-Chem (Lamarque et al., 2012). The global and local responses of precipitation and SST to the aerosol forcing changes are shown in Figure 5 under the RCE and WTG assumption respectively (Using the tropical aerosol forcing in RCE run is unrealistic, but the results of doing so are instructive.). In RCE run, the SST decreases with the increasing of aerosols, and hence the decrease of precipitation. This is because increased aerosols lead to decreased net downward radiative flux at the surface. As a result, SST decreases and hence the latent heat flux. In WTG run, in addition to the similar declining characteristics as in RCE run, both SST and precipitation at equilibrium are getting smaller. The reason is that the induced subsidence in WTG run significantly dries out the atmospheric column and leads to increased OLR, thus further cooling the air–sea system. This demonstrates that a regional response may differ from a global response even if under the same external forcing.

3.2.3. Linear versus nonlinear response

The above two sections estimated the response of SST and precipitation to individual external forcings, but how will climate respond to multiple forcings? Can the total response be simply approximated by linearly adding these individual responses together? To answer this question, a total of 30 experiments were carried out, each with a different combination of CO₂ and aerosol forcings. Figure 6 displays the scatter plot of the linear SST response versus the total SST response under various forcing combinations. The SST in the control experiment ($1 \times CO₂, \tau = 0$) is subtracted from each experiment and only their differences are shown. As noticed, the points do not strictly collapse on the diagonal line, indicating the linear approximation is not of great accuracy. This also demonstrates that the nonlinear interactions among these forcings are important, as also found in He et al. (2012). The bias between the linear and total response can be as large as 2.8 °C, i.e. for the experiment ($2 \times CO₂, \tau$), where the total response shows a value of 0 while the linear response exhibits a value of −2.8 °C. Most points lie above the diagonal line, implying that the nonlinear effects act to offset the cooling response caused by aerosols. The linear and total responses are getting agreeable under high CO₂ and low aerosol scenarios, presumably owning to the dominant role of water vapor feedback in the warming atmosphere.

4. Concluding remarks

The 1D air–sea interactive model that couples an atmospheric column model with a slab ocean model was introduced. The model performances under both RCE and WTG assumptions are analyzed. The model was then used to study the response of SST and precipitation to CO₂ and aerosol forcing changes. Results show a warming or cooling signal owing to increased CO₂ or aerosols can be well captured by the ocean, which is manifested in increasing or decreasing of the SST. However, cutting off the interactions between atmosphere and ocean renders different responses. By properly specifying the SST, the atmospheric model alone can produce similar precipitation responses as in the coupled simulation. This means AGCMs can be used to study climate response to external forcings, however, their results should be viewed cautiously if the specified SST in the AGCM deviates significantly from the projected SST in the CGCM.

Under the relaxed WTG approximation, the decreasing of SST as a response to increasing aerosols is even more pronounced in comparison with that in RCE run. This is because the imposed WTG assumption induces large-scale subsidence, which continuously dries out the column and leads to increased OLR, thus further cooling the air–sea system. This implies different regions may have different responses even if under the same external forcing. The relaxation time scale $T_{relax}$ plays an important role in modulating the interactions between convection and WTG vertical velocity. An infinitely small/large $T_{relax}$ leads to results that are close to those in strict WTG/RCE, respectively.

When multiple forcings exist in the climate system, the method that approximates the total response by linearly adding individual responses is found to be inaccurate. The nonlinear effects are found to be significantly
Equilibrium response to CO$_2$ and aerosol forcings in a 1D model

Figure 5. Precipitation (unit: mm day$^{-1}$) and SST (unit: °C) as a response to aerosol forcing changes under the (a) RCE and (b) WTG assumption.

Figure 6. Scatter plot of the linear SST response (Abscissa) versus the total SST response (Ordinate).

important, which act to offset the cooling response caused by aerosols. However, under high warming scenarios, the linear and total responses are fairly agreeable, presumably owning to the dominant role of water vapor feedback in the warming atmosphere.

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