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Transient and Equilibrium Climate Response to Anthropogenic Forcing

1. **Introduction**

Climate sensitivity refers to the change of globally averaged surface temperature in response to a specified external forcing, which is typically taken as the radiative flux change from doubling the preindustrial concentration of carbon dioxide in the atmosphere, at about 3.7 W/m2. Because of the increasing amount of atmospheric greenhouse gases as a result of the use of fossil fuels, understanding the magnitude of the response of surface temperature to anthropogenic activities, and thus the sensitivity of the climate system, has been an active area of research in the last thirty years (Randall et al. 2007).

Coupled General Circulation Models (CGCMs) are one of the few available tools to derive the climate sensitivity. Earlier studies used an atmospheric model with a mixed layer slab ocean to integrate a model to equilibrium state under constant forcing (e.g., Manabe et al., Hansen.). This sensitivity is referred to as the equilibrium sensitivity of a climate model. It has been known that atmospheric feedback processes, especially cloud- feedback, plays a major role in determining the sensitivity of the climate system or a climate model (Cess et al., Mitchell., Manabe Wetherald). The discrepancies in cloud feedback among the models imply several factors of difference in climate responses from these models to a doubling of CO2 forcing.

Results from the multi-model ensemble of IPCC AR4 (Inter-governmental Panel on Climate Change Fourth Assessment Report) have however shown that different climate models simulated similar magnitudes of global tempeature change for the 20th century (IPCC Meehl), even though their atmospheric feedback processes are known to be very different (Soden, Bony). As an example, Figure 1(a) shows the changes of temperature in the 140 years from 1860 to 2000 simulated by using two CGCMs, one from the National Center for Atmospheric Research (NCAR) version CCSM3, and the other from the Geophysical Fluid Dynamics Laboratory (GFDL) version CM2. Also plotted is the observed temperature variation during this period. A best estimate of the climate forcing, from greenhouse gases, from the combination of greenhouse gas and tropospheric aerosol, and the total (including greenhouse gases, tropospheric aerosol, solar activities, and volcanic aerosols) are shown in Figure 1. The industrialization has created a greenhouse forcing of about 2.5 W/m2; this is offset by about 1 W/m2 from increased aerosol loading in the atmosphere during the same period, leading to a total net forcing of about 1.5 W/m2. The two models in Figure 1(a) used a variant of the forcing in Figure 1(a).

The two models are known to have different cloud-climate feedbacks: the NCAR CCSM3 has a negative cloud feedback; while the GFDL AM2 has a positive cloud feedback. This is shown in Figure 2, in terms of the change of cloud radiative forcing at the top of the atmosphere (TOA) normalized to a unit change of surface temperature. Averaged from 60oS to 60oN, the changes of clouds in the CCSM have the effect of losing 1.2 W/m2 of energy for one degree change of surface temperature, while cloud changes in the model GFDL model would lead to a 0.2 W/m2 gain of radiative energy. Combining these values with the approximate 0.7oK climate change, one can infer the forcing from the sum of the externally imposed forcing and the cloud forcing in the CCSM to be about 0.5 W/m2, much less than the 1.5 W/m2 in Figure 1b, while that in the GFDL CM to larger than 1.5 W/m2. Given these differences, how can the two models both simulate similar climate for the 20th century?

This paper investigates the impact of atmospheric feedback (mainly cloud feedback) on the transient climate response to external forcing. The purpose is to answer the simple question: will models with different equilibrium climate sensitivities produce similar 20th century climate from the observed forcing? Answer to this question will provide guidance about what factors are the most important in determining climate change in the time horizon of several decades to a century. It will additionally address whether cloud feedback can be inferred from modern instrument records.

The paper starts in the follows with the description of a model in Section 2, and the terminology of forcing and sensitivity in Section 3. Results are presented in Section 4 for a special case and in Section 5 for a more general case. The last section contains a brief summary of the paper.

**2. The Model**

The atmosphere is described by a simple energy balance model averaged over the whole atmosphere written for a unit surface area as

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where is the moist static energy of the atmosphere:

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NS  and NT denote the net upward energy flux at the surface and TOA respectively with surface pressure *ps* and TOA pressure *pt*; *Ta, z, q* are air temperature, height, and water vapor mixing ration; *Cp, g*, and *L* are specific heat of air under constant pressure, gravitational constant, and latent heat of evaporation or condensation. Equation (1) is valid for a hydrostatic atmosphere. *pt* does not have to be the real top of the atmosphere. Instead, it is often selected at a level where external forcing is defined.

The Earth surface is assumed to be covered by water with a mixed layer; the depth of this layer is D. The horizontally averaged mixed-layer thermodynamic equation is

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where is the mixed-layer temperature, which we refer to as the sea surface temperature (SST), and are horizontal currents, and are the specific heat and density of sea water. *ND* is the upward heat flux at the bottom of the mixed layer. This heat flux can be expressed in terms of entrainment velocity at the bottom of the ocean mixed layer, just like the turbulent entrainment of heat and moisture at the top of atmospheric boundary layer (e.g., Lilly 1968). It is written as

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where is the deep water temperature immediately below the mixed layer.

Below the mixed layer for z< -D, the deep ocean temperature is controlled by

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where k is the vertical diffusivity; , v and are velocity components of the ocean currents. The upper boundary condition of the deep ocean

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The other boundary condition at the bottom of the deep ocean can assume two forms: one with no heat flux, and the other with fixed temperature, corresponding to a Dirichlet or Neumann boundary conditions:

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where H is the depth of the ocean. A more physically defensible boundary condition at the bottom of the ocean is to make the net heat flux to be equal to the thermal flux of the Earth crust, but this would introducing additional variables. For practical purposes of studying climate change on the time scale of less than a few centuries, (7) and (8) are equally valid since heat diffusion into the deep ocean is a very slow process. These assumptions are equivalent to setting the ocean to an infinite depth for the study of a forced solution from the atmosphere. Therefore, in the following study, we will use H=.

The governing equation of (1), (3) and (5) describe the three components – atmosphere, SST, and a deep ocean – of the climate system under external forcing through the atmosphere. These components are schematically shown in Figure 1. We will be only concerned with the tropical region, since it occupies the largest portion of the surface. For simplicity, horizontal transport of heat in this region is neglected, and the upwelling in that region is specified.

The coupling between the atmosphere and the mixed layer ocean is through the net surface heat flux *Ns*; the coupling between the mixed layer and the deep ocean is through the turbulent heat flux at the bottom of the mixed layer *ND*.

Combining (1) and (3) without the ocean currents yields the heat budget equation for the atmosphere and the ocean mixed-layer:

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**3. The formulation of the forcing and feedback**

The TOA net upward heat flux in (9) is the sum of shortwave and infrared longwave radiation and so it is only a function of the surface and atmospheric state. We can functionally write it as

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where and are the three-dimensional air temperature and water vapor respectively; represents greenhouse gas concentration; other variables are self explanatory.

Other independent variables can be added if appropriate.

To study climate change under an external forcing, all state variables are written as perturbation to a reference state such as the current climatological condition. The above equation can be therefore, after linearization, written as

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All partial derivatives are taken at the reference state. We have used CO2 as a surrogate for all greenhouse gases.

Climate forcing can be imposed by changes in the anthropogenic or natural event-driven independent variables. The anthropogenic forcing includes atmospheric greenhouse and aerosol forcing, while the natural forcing includes solar variability. These forcing terms are combined together as a total forcing written as a downward radiative flux:

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All other terms in equation (21) represent atmospheric feedback processes, often expressed using the surface temperature as the control variable. They are the negative Stefan-Boltzman radiative feedback, temperature lapse-rate feedback, water vapor feedback, snow and sea-ice feedback, and cloud feedback, which can combined as

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The symbol is added for clarity in this expression only. The coefficients of the partial derivatives to each independent variable are called as the radiative kernels in Soden (2009); these are found to be somewhat independent of which climate model is used. The coefficients of are called feedbacks, in units of (W/m2)/K.

To isolate the cloud feedback, which is the largest uncertainty and source of discrepancy in climate models, we will combine the Stefan-Boltzman term and those from temperature lapse rate, water vapor, snow/sea ice into a single parameter, λ0, expressed as W/m2/K, and write the cloud feedback as the change of cloud-radiative forcing (downward) as λc:

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where

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This cloud feedback represents the increase of net downward radiation at TOA due to change of clouds associated with a unit change of surface temperature.

A more widely used definition of cloud-radiative forcing, denoted by , is the difference of total-sky and clear-sky radiative flux (Ramanathan xx):

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This definition differs slightly from that in (26), but it can be observed from satellite measurements and can be more easily calculated in climate models. The two are related with each other

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We will not devolve further into the terminology (interested readers are referred to Zhang et al. (1994) and Soden and Held (2007) for additional discussions).

The net upward TOA radiative flux in (11) can thus be written as

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where the first term on the right hand side is the climate forcing; the second term is the atmospheric feedbacks. A positive represents positive cloud feedback, while represents a negative cloud feedback. It should be noted that in terms of difference among climate models, we can consider to be the ensemble mean of all terms in (13) and to represent all intermodal differences. But for ease of description, we refer to it as cloud feedback.

**4. Results**

*Terminology and specifications*

In the energy budget equation (9) for the atmosphere and the ocean mixed layer, if we assume an ocean mixed layer of 75 meters as observed or in climate models (e.g., Liu et al. 2010), the coefficient for the oceanic term on the left hand side of the equation is =3108 J/K, while the coefficient of the atmospheric term is on the order of *Cpps/g = 1107* J/K. Therefore, atmospheric term on the left hand side of (9) can be neglected. This is equivalent to assume a zero heat capacity atmosphere, but the atmosphere is needed to introduce the climate forcing. Together with equation (4), the controlling equation of SST can be written as

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The terms on the right hand of (20) describe the climate forcing, atmospheric feedbacks, and mixing of heat with the deep ocean.

We next introduce the terminology of climate sensitivity and response. The equilibrium sensitivity is in theory only defined when the forcing in (20) is steady, i.e., independent of time. It refers to the magnitude of SST change under a unit external forcing, but this is sometimes also referred to as the magnitude of SST change from forcing of a doubling of atmospheric CO2 concentration relative to preindustrial concentration.