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Two opposing effects of absorbing aerosols on global-mean precipitation

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[1] Absorbing aerosols affect global-mean precipitation primarily in two ways. They give rise to stronger shortwave atmospheric heating, which acts to suppress precipitation. Depending on the top-of-the-atmosphere radiative flux change, they can also warm up the surface with a tendency to increase precipitation. Here, we present a theoretical framework that takes into account both effects, and apply it to analyze the hydrological responses to increased black carbon burden simulated with a general circulation model. It is found that the damping effect of atmospheric heating can outweigh the enhancing effect of surface warming, resulting in a net decrease in precipitation. The implications for moist convection and general circulation are discussed.

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1. Introduction

[2] A robust characteristic of the simulated response to greenhouse gas warming is a modest increase in global-mean precipitation (2–3% K^{-1}) as more latent heating compensates for stronger radiative cooling, a necessary condition for establishing a new equilibrium climate state [e.g., *Allen and Ingram*, 2002; *Wetherald and Manabe*, 2002; *Held and Soden*, 2006]. A shift of the same balance (albeit to a colder climate) explains the reduction in precipitation caused by purely scattering aerosols such as sulfate [e.g., *Meehl et al.*, 1996; *Roeckner et al.*, 1999]. In comparison, the existing studies on the long-term hydrological impacts of absorbing aerosols such as black carbon (BC) are relatively few [e.g., *Menon et al.*, 2002; *Liepert et al.*, 2004; *Ramanathan et al.*, 2005; *Randles and Ramaswamy*, 2008], despite the fact that some of the future emission scenarios project more BC along with decreasing sulfur [*Levy et al.*, 2008].

[3] Although some particular aspects of the issue (e.g., reduced surface solar flux, atmospheric heating, stabilization of the troposphere and reduced precipitation) have been discussed, often in the context of the surface energy budget and on the regional scale, still missing is a theoretical framework in which one is able to quantify all the processes essential for determining the change in global-mean precipitation, and thus to devise an *a priori* measure of the ability of a particular climate perturbation to alter precipitation, analogous to what radiative forcing is for surface temperature. Such a measure would be highly desirable for purposes like model intercomparison and attribution of observed and model-simulated changes in precipitation.

[4] This study approaches the issue from the angle of energy balance constraint on the hydrological cycle. We argue that despite the large uncertainty in the current physical understanding and model representation of the radiative and/or microphysical effects of aerosols on individual precipitation events [*Khain*, 2009, and references therein], the global-mean precipitation has to vary under such a constraint. This would generate valuable insights into the robustness of model simulations. The same methodology has been utilized successfully to study decadal-scale hydrological response to greenhouse gases [*Allen and Ingram*, 2002; *Held and Soden*, 2006].

2. Design of Experiments

[5] We first use a modified version of the Geophysical Fluid Dynamics Laboratory (GFDL) AM2.1 atmosphere general circulation model (AGCM) [*The GFDL Global Atmospheric Model Development Team*, 2004] for evaluating the atmosphere-only perturbations, and then couple it to a mixed-layer ocean model for simulating the corresponding climate responses. This particular AGCM includes a prognostic treatment of the interactions between aerosols and liquid clouds as described by *Ming et al.* [2006, 2007]. More detailed description of the coupled model is given by *Ming and Ramaswamy* [2009]. We perturb the pre-industrial control case by adding 2.4×10^{-6} kg m^{-2} to the burden of BC within a σ-layer across the entire globe. The burden is chosen so that the corresponding radiative perturbations are comparable to that of the present-day anthropogenic BC (estimated at 0.53 W m^{-2} in AM2.1). This is done for a series of layers in the planetary boundary layer (PBL) and in the free troposphere (FT) (Table 1). We also examine a realistic distribution of present-day BC to illustrate whether the results vary with the spatial pattern of perturbation [*Ginoux et al.*, 2006].

3. Results

[6] This study employs the top-of-the-atmosphere (TOA) all-sky flux change (F, warming as positive), as opposed to the instantaneous forcing, as a measure of the radiative perturbation to the climate system [*Hansen et al.*, 2005]. The values of F, calculated with AGCM, and resulting changes in surface temperature (δT_s), simulated with the mixed-layer model, are listed in Table 1. For the same amount of increase in BC loading, F varies with altitude mainly through interacting with clouds, but is typically positive as expected. The only exception occurs for the layer with σ = 0.98. A close look suggests that the clear-sky component of

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Table 1. TOA Flux Change, Changes in Atmospheric Absorption, Surface Temperature and Surface Sensible Heat Flux, and Relative Changes in Precipitation and Convective Mass Flux[a]

σ (Altitude)	F	δAA	δT_s	δSH	$\delta P/P$	$\delta M_c/M_c$
0.99 (35)	0.92	4.8	1.5	−4.5	2.1	−1.5
0.98 (200)	−0.82	3.8	0.25	−3.8	−0.07	7.4
0.95 (460)	1.4	4.6	2.1	−3.4	2.1	−1.5
0.90 (850)	3.2	5.8	3.4	−2.3	3.0	−10.7
0.84 (1450)	1.4	5.7	1.6	−1.8	−1.6	−8.1
0.77 (2200)	1.2	6.2	1.3	−1.7	−2.7	−12.0
0.60 (4100)	1.6	7.9	1.8	−1.7	−3.5	−29.5
Realistic BC	0.53	1.4	0.40	−0.45	−0.4	−4.4

[a] δAA, change in atmospheric absorption (W m^{-2}); δT_s, change in surface temperature (K); δSH, change in surface sensible heat flux (W m^{-2}); $\delta P/P$, relative change in precipitation (%); $\delta M_c/M_c$, relative change in convective mass flux (%). The approximate altitude (m) for each σ-layer is given.

F, like those at the other layers, is positive, but is outweighed by the negative cloudy-sky component. Thus, the negative F is attributed to some peculiarities in the simulated changes in cloud fields. Also note that most of the idealized perturbations are 2–3 times of that of realistic BC.

[7] It is clear from Figure 1 that F is a reliable predictor of δT_s. This is true almost for the entire range of F (0.53–3.2 W m^{-2}). The only exception is the abnormal negative F at $\sigma = 0.98$. Note that the slope of the best linear fit with zero intercept (1.1 K m^2 W^{-1}) can be thought of as the model's equilibrium climate sensitivity (λ) for BC, which is reasonably close to that for greenhouse gases (1.3 K m^2 W^{-1}).

[8] Also in Table 1 are the percentage changes in simulated global-mean precipitation ($\delta P/P$). For warming caused by greenhouse gases including CO_2, $\delta P/P$ scales reasonably well with δT_s with a ratio of 2–3% K^{-1} [e.g., *Allen and Ingram*, 2002; *Held and Soden*, 2006]. This does not hold for absorbing aerosols. δT_s and δP even differ in sign for the perturbations imposed at three σ-layers (0.60, 0.77 and 0.84). In these cases, precipitation decreases despite considerable surface warming (1.3–1.8 K). Why does the hydrological cycle respond so differently to the two common climate perturbations?

[9] An analysis of the global-mean energy budget of the atmosphere provides a theoretical framework that is useful for answering this question. LW radiative cooling has to be balanced out by surface sensible and latent heating, and atmospheric absorption (both in SW and LW). When one considers how these factors would vary as absorbing aerosols force the shift in equilibrium climate state, the picture can be simplified into the following equation:

$$k\delta T_s = \delta AA + L\delta P + \delta SH. \qquad (1)$$

δAA represents the fraction of the overall variation in atmospheric absorption that is induced directly by absorbing aerosols through atmosphere-only processes, and thus is independent of subsequent δT_s. The combined change in atmospheric absorption and LW radiative cooling caused by δT_s through a variety of feedback mechanisms (including cloud feedback) is assumed to be proportional to δT_s, with k as proportionality constant [*Allen and Ingram*, 2002; *Andrews et al.*, 2009]. L is the specific latent heat of water, and SH is the sensible heat received by the atmosphere.

[10] The AGCM simulations suggest that irrespective of altitude, higher BC burden enhances atmospheric absorption considerably by 3.8–7.9 W m^{-2} (δAA in Table 1). This alone tends to suppress precipitation according to equation (1). Stronger absorption heats up the atmosphere first, and then the surface to a certain degree. This is why the sensible heat flux from the surface into the interior of the atmosphere decreases (δSH in Table 1). Unlike δAA, δSH varies strongly with the vertical location of perturbation. The three layers adjacent to the surface (with σ equal to or greater than 0.95) see the largest decreases in SH, which counteract most, if not all, of the enhancement in atmospheric absorption. This indicates that these near-surface layers and the surface are tightly "coupled". In comparison, the atmosphere retains most of the increase in absorption at the upper levels. These findings are consistent with *Chung and Zhang* [2004]. The same analysis as in Figure 1 of *Andrews et al.* [2009] (not shown) indicates that regardless of the altitude of BC, δSH does not correlate with δT_s for the first 30 years of the simulations, during which T_s gradually changes. This is consistent with the view that δSH is driven primarily by atmosphere-only processes, and thus can be classified as "fast response."

[11] By re-arranging equation (1) and taking into account the fact that the global-mean latent heat flux (LP) is 86.3 W m^{-2} in the pre-industrial control case, one can express the relative change in precipitation ($\delta P/P$) as

$$\delta P/P = 0.0116 \times (k\delta T_s - \delta AA - \delta SH). \qquad (2)$$

We estimate k at 1.8 W m^{-2} K^{-1}, based on the hydrological response to greenhouse gases-induced warming simulated with the same model. This translates into 2.0% K^{-1}, which is within the range of reported values for different models [see *Allen and Ingram*, 2002, Figure 2]. Note that the constant 0.0116 is the reciprocal of LP. Figure 2 shows that the values of $\delta P/P$ calculated with equation (2) are in good agreement with the simulations. This leads us to conclude that the above framework captures the factors key to determining the global-mean hydrological response to absorbing aerosols, and thus can be utilized to better understand its characteristics.

[12] Surface warming, as is the typical thermal response to absorbing aerosols, invariantly favors more precipitation. This effect is responsible for the positive scaling between

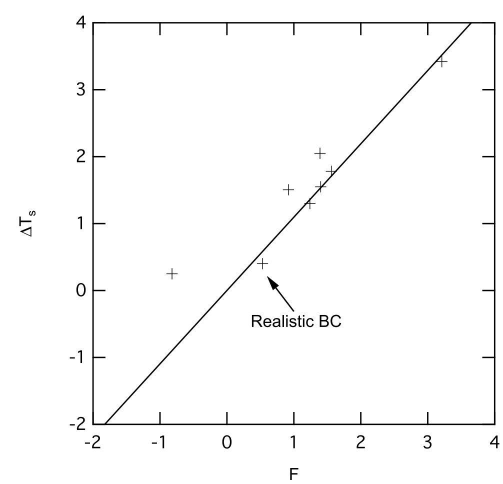

Figure 1. Scatter plot (crosses) of TOA flux change (F, W m^{-2}) and change in surface temperature (δT_s, K). The line represents the best linear fit with zero intercept ($\delta T_s = 1.1F$, $R^2 = 0.66$).

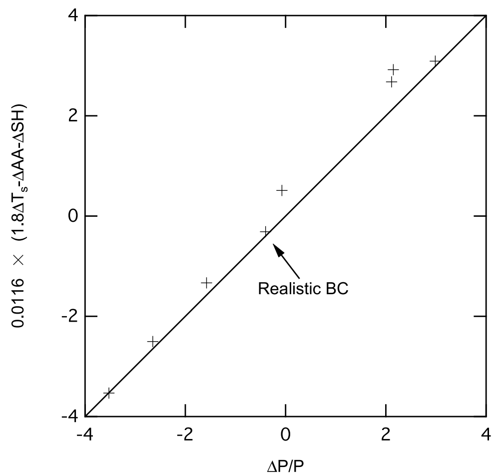

Figure 2. Scatter plot (crosses) of relative change in precipitation ($\delta P/P$, %) and a derived quantity ($0.0116 \times (1.8\delta T_s - \delta AA - \delta SH)$, %).

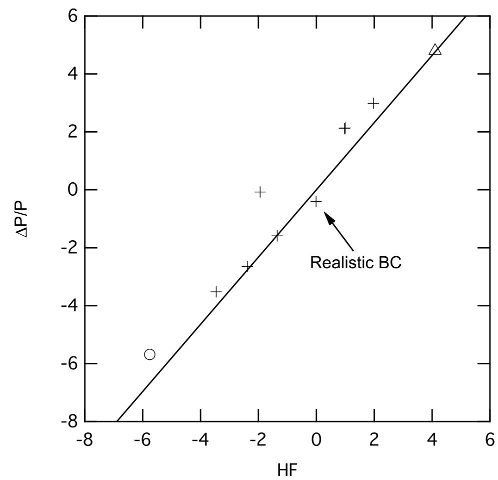

Figure 3. Scatter plot of hydrological forcing (HF, W m^{-2}) calculated as $2.0F - \delta AA - \delta SH$ and relative change in precipitation ($\delta P/P$, %). The crosses are for BC. The triangle and circle correspond to the perturbations caused by the pre-industrial-to-present-day increases in radiatively active gases and aerosols, respectively, as described in *Ming and Ramaswamy* [2009]. Note that the hydrological forcing of gases also includes the increase in the LW downward surface flux [*Allen and Ingram*, 2002]. The solid line denotes $\delta P/P = 0.0116\ HF$.

$\delta P/P$ and δT_s for greenhouse gas warming. On the other hand, stronger atmospheric absorption has a suppressing effect. The loss in sensible heating is usually small unless aerosols reside in the layers closest to the surface. The simulations show that the absorption effect can potentially dominate the warming effect when a large fraction of the enhanced absorption is kept within the atmosphere, thus giving rise to a net decrease in precipitation in spite of surface warming. This is accompanied by significant reduction in lapse rate [*Ramanathan et al.*, 2005; *Erlick and Ramaswamy*, 2003]. As aerosols are heavily concentrated in the lower troposphere in reality, an implication of the above finding is that their overall effect would be sensitive to the vertical profile. For the distribution of present-day BC in AM2.1, the two effects roughly cancel out, leaving little change in global-mean precipitation (Table 1). Despite these insights, more research is needed to shed light on the physical mechanisms, through which the hydrological responses discussed here are realized, especially on the regional scale.

4. Discussion

[13] We propose a concept of hydrological forcing (HF), which would provide a means to quantify the ability of a climate perturbation to modify global-mean *precipitation* without performing expensive coupled model simulations. It is analogous to radiative forcing (RF), except that RF measures the impact on *surface temperature*. Building upon the success of equation (2) in providing a mechanistic explanation of the simulated variations in precipitation, one may calculate the HF of absorbing aerosols as $k\lambda F - \delta AA - \delta SH$. As δAA can be written as the difference between TOA and surface flux changes (the latter denoted as F_s), an alternative expression for HF is $(k\lambda - 1)\,F + F_s - \delta SH$. Note that F, δAA (or F_s) and δSH can be evaluated readily with AGCM simulations. As discussed before, the climate sensitivity (λ) of the model is 1.1 K m^2 W^{-1}, and k is 1.8 W m^{-2} K^{-1}. Figure 3 compares HF computed as $2.0\ F - \delta AA - \delta SH$ with $\delta P/P$. $\delta P/P$ correlates well with HF for all BC cases. Furthermore, it is encouraging to see that the HF based on this formula represents reasonably well the hydrological perturbations posed by the total direct and indirect effects of aerosols and by radiatively active gases discussed by *Ming and Ramaswamy* [2009].

[14] Built upon the mass balance of water in FT, the thermodynamic argument, as laid out by *Held and Soden* [2006], dictates that the relative change in convective mass flux ($\delta M_c/M_c$) follows $\delta P/P - 0.07\delta T_s$. *Held and Soden* [2006] also showed that δM_c can be used as a proxy for the change in the tropical mean circulation. It appears from Figure 4 that the simulated $\delta M_c/M_c$ generally follows the thermodynamic argument. The pronounced decreases in M_c, especially for the perturbations at the upper layers, are presumably caused by reduced lapse rate (convective instability). For CO_2 and purely scattering aerosol effects, δP is always in

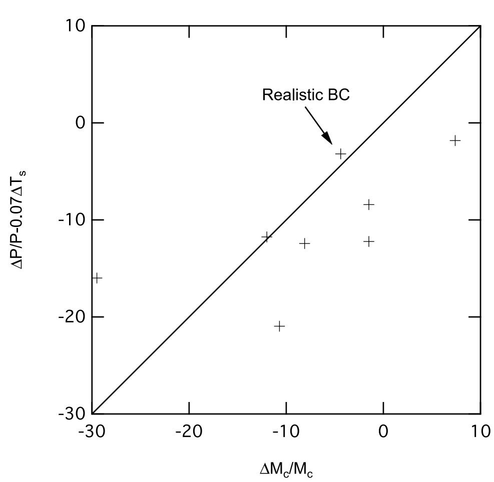

Figure 4. Scatter plot (crosses) of relative change in convective mass flux ($\delta M_c/M_c$, %) and a derived quantity ($\delta P/P - 0.07\delta T_s$, %).

the same sign as δT_s, thus mitigating δM_c implied by the Clausius-Clapeyron scaling. However, this is no longer true for absorbing aerosols since precipitation may be suppressed by absorption despite surface warming. As a result, they could be much more potent at altering circulation patterns than CO_2 and scattering aerosols (Table 1). Modest surface warming (a few tenths of a degree above 1 K) is often accompanied by substantial reductions in M_c (up to 29.5%).

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