Two opposing effects of absorbing aerosols on global-mean precipitation
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Received 11 February 2010; revised 19 May 2010; accepted 25 May 2010; published 2 July 2010.

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.


1. Introduction

A robust characteristic of the simulated response to greenhouse gas warming is a modest increase in global-mean precipitation (~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].

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 inter-comparison and attribution of observed and model-simulated changes in precipitation.

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

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

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 (Tσ), 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
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.

<table>
<thead>
<tr>
<th>$\sigma$ (Altitude)</th>
<th>$F$</th>
<th>$\delta AA$</th>
<th>$\delta T_s$</th>
<th>$\delta SH$</th>
<th>$\delta PP$</th>
<th>$\delta M/M_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.99 (35)</td>
<td>0.92</td>
<td>4.8</td>
<td>-4.5</td>
<td>2.1</td>
<td>-1.5</td>
<td></td>
</tr>
<tr>
<td>0.98 (200)</td>
<td>-0.82</td>
<td>3.8</td>
<td>0.25</td>
<td>-3.8</td>
<td>-0.07</td>
<td>7.4</td>
</tr>
<tr>
<td>0.95 (460)</td>
<td>1.4</td>
<td>4.6</td>
<td>2.1</td>
<td>-3.4</td>
<td>2.1</td>
<td>1.5</td>
</tr>
<tr>
<td>0.90 (850)</td>
<td>3.2</td>
<td>5.8</td>
<td>-2.3</td>
<td>3.0</td>
<td>-10.7</td>
<td></td>
</tr>
<tr>
<td>0.84 (1450)</td>
<td>1.4</td>
<td>5.7</td>
<td>-1.8</td>
<td>-1.6</td>
<td>-8.1</td>
<td></td>
</tr>
<tr>
<td>0.77 (2200)</td>
<td>1.2</td>
<td>6.3</td>
<td>-1.7</td>
<td>-2.7</td>
<td>-12.0</td>
<td></td>
</tr>
<tr>
<td>0.60 (4100)</td>
<td>1.6</td>
<td>7.9</td>
<td>1.8</td>
<td>-1.7</td>
<td>-3.5</td>
<td>-29.5</td>
</tr>
<tr>
<td>Realistic BC</td>
<td>0.53</td>
<td>1.4</td>
<td>0.4</td>
<td>-0.45</td>
<td>-0.4</td>
<td>-4.4</td>
</tr>
</tbody>
</table>

$\delta AA$, change in atmospheric absorption (W m$^{-2}$); $\delta T_s$, change in surface temperature (K); $\delta SH$, change in surface sensible heat flux (W m$^{-2}$); $\delta PP$, change in precipitation (%); $\delta M/M_c$, relative change in convective mass flux (%). The approximate altitude (m) for each $\sigma$-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 $\Delta T_s$. 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 ($\lambda$) 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 ($\delta PP$). For warming caused by greenhouse gases including CO$_2$, $\delta PP$ scales reasonably well with $\Delta T_s$ 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. $\Delta T_s$ and $\delta PP$ even differ in sign for the perturbations imposed at three $\sigma$-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:

$$k\delta T_s = \delta AA + L\delta P + \delta SH.$$  

(1)

$\delta AA$ 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 $\delta T_s$. The combined change in atmospheric absorption and LW radiative cooling caused by $\delta T_s$ through a variety of feedback mechanisms (including cloud feedback) is assumed to be proportional to $\delta T_s$, with k as proportionality constant [Allen and Ingram, 2002; Andrews et al., 2009]. $L$ is the specific latent heat of water, and $\delta SH$ 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$^{-2}$ ($\delta AA$ 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 ($\delta SH$ in Table 1). Unlike $\delta AA$, $\delta SH$ varies strongly with the vertical location of perturbation. The three layers adjacent to the surface (with $\sigma$ 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, $\delta SH$ does not correlate with $\delta T_s$ for the first 30 years of the simulations, during which $T_s$ gradually changes. This is consistent with the view that $\delta SH$ 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 ($\delta PP$) is 86.3 W m$^{-2}$ in the pre-industrial control case, one can express the relative change in precipitation ($\delta PP$) as

$$\delta PP = 0.0116 \times (k\delta T_s - \delta AA - \delta SH).$$

(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 $\delta PP$ 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 ($\delta T_s$, K). The line represents the best linear fit with zero intercept ($\delta T_s = 1.1F$, $R^2 = 0.66$).](image-url)
$\delta P/P$ and $\delta T_s$ 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 $k\Delta F - \delta AA - \delta SH$. As $\delta AA$ 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 $F$, $\delta AA$ (or $F_s$) and $\delta SH$ can be evaluated readily with AGCM simulations. As discussed before, the climate sensitivity ($\lambda$) 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 $\delta P/P - 0.078T_s$. Held and Soden [2006] also showed that $\delta M_c$ 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, $\delta P$ is always in positive range.
the same sign as $\delta T_s$, thus mitigating $\delta M_c$ 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%).

Acknowledgments. G.P. was supported by the Ernest F. Hollings Undergraduate Scholarship Program, administered by NOAA’s Oce of Education.

References


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