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Understanding hydrological sensitivities induced by various forcing agents with a climate model

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Abstract

The apparent hydrological sensitivity, defined as the global-mean precipitation change per increase of the global-mean temperature, is investigated for scenarios induced by different forcing agents. Simulations with a climate model driven individually by four different climate forcings, i.e. sulfate, black carbon, solar insolation and carbon dioxide (CO₂), are analyzed in the context of energy balance controls on global precipitation to explore how different forcing agents perturb different energy components grouped into fast and slow responses. Similarities and differences among the forcing agents are found in ingredients of the tendency contributing to the hydrological sensitivity from various energy budget components. Specifically, the sulfate and solar forcings induce the hydrological sensitivity of $\sim 3.0\%K^{-1}$ due to the slow response of radiative cooling whereas the black carbon induces a significantly negative hydrological sensitivity ($\sim$

6.0%K⁻¹) due to the strong atmospheric heating. The CO₂-induced hydrological sensitivity is found in between (~1.2%K⁻¹) as a result from the slow response of radiative cooling and its partial compensation by the atmospheric heating. The findings provide a quantitative basis for interpreting climatic changes of global precipitation driven by a mixture of various natural and anthropogenic forcings.

1. Introduction

Prediction of the global precipitation change is still subject to large uncertainty particularly when the climate change is driven by various forcing agents including different types of aerosols as well as long-lived greenhouse gases such as carbon dioxide (CO₂). The climatic change of global precipitation is often quantified by the hydrological sensitivity (HS), defined as the global-mean precipitation change per increase of the global-mean surface air temperature. The HS has earlier been studied for CO₂-induced global warming scenario based on the fact that the global-mean precipitation change is largely constrained by energy balance controls [e.g. *Allen and Ingram*, 2002; *Held and Soden*, 2006; *Stephens and Ellis*, 2008; *Pendergrass and Hartmann*, 2014; *DeAngelis et al.* 2015; *Watanabe et al.* 2018]. The energy balance argument has more recently been extended to analysis of HS manifested in more realistic historical and future scenario simulations driven by a mixture of various forcing agents [*Shiogama et al.* 2010; *Pendergrass and Hartmann*, 2012; *Wu et al.* 2013; *Suzuki et al.* 2017]. These studies have pointed out that the global precipitation change contains a component that does not simply scale with the global-mean surface air temperature. In this context, the global-mean precipitation change containing the “temperature-independent” component simply

divided by the global-mean surface air temperature change is referred to as the *apparent* hydrological sensitivity (hereafter AHS) [Fläschner *et al.* 2016], which is convenient to estimate in historical and future scenario simulations. Factors influencing the atmospheric energy budget can contribute to AHS as argued by Stephens and Hu [2010], who highlight aerosol and cloud feedback as most uncertain factors. This study is intended to quantitatively understand AHS induced by various climate forcers.

A major factor making AHS different from the pure HS is the rapid adjustment (or the fast response) of precipitation to climate forcing. The rapid adjustment is particularly pronounced by atmospheric radiative heating of light-absorbing aerosols, such as black carbon (BC), which substantially deviates the historical precipitation change from the temperature-dependent trend [Suzuki *et al.* 2017] and is a major cause for the inter-model spread of AHS in future climate change scenarios [Pendergrass and Hartmann, 2012]. These studies underscore the importance of separating the fast and slow response components for better quantification of AHS.

These research needs motivate a recent community effort of the model inter-comparison called Precipitation Driver and Response Model Intercomparison Project (PDRMIP) [Samset *et al.* 2016; Myhre *et al.* 2017]. This is coordinated to perform simulations with multiple global climate models individually driven by various forcing agents including sulfate (SF) and BC aerosols, solar insolation (Sol) and CO₂. The results from PDRMIP highlight how AHS differs among forcing agents [Samset *et al.* 2016] and how this difference is attributed to differing contributions from fast responses of energy budget components [Samset *et al.* 2018; Richardson *et al.* 2018]. The relative contributions from fast and slow responses are better understood in a whole picture of global energy budget [Suzuki and Takemura, 2019; hereafter ST19], which also explains

1 stark differences in global-mean responses of temperature [Takemura and Suzuki, 2019]
2 and precipitation to BC and SF.

3 This study builds upon these previous studies to better understand AHS induced by
4 various forcing agents through quantifying relative contributions of different factors to
5 AHS. The factors influencing the atmospheric energy budget are identified in the manner
6 tied to different physical processes. For this purpose, we employ climate model
7 simulations individually driven by four major forcing agents (SF, BC, Sol and CO₂) in
8 the way that extends PDRMIP. The results are analyzed in the context of atmospheric
9 energy budget that is broken down into component fluxes corresponding to physical
10 processes to quantify their corresponding contributions to AHS. The ingredients of AHS
11 tendencies due to different factors are then compared among the forcing agents to
12 characterize how different forcing agents induce various AHS values as a result from
13 different compositions of energy balance component perturbed. As opposed to the multi-
14 model analysis of PDRMIP, the single-model analysis of this study compares
15 precipitation responses to different forcing agents under a common set up of model
16 physics. This has an advantage of avoiding contaminations of the model-to-model
17 discrepancy in representation of physical processes.

18 **2. Model experiments**

19 We perform and analyze the model experiments using MIROC5 [Watanabe *et al.* 2010]
20 coupled with the SPRINTARS aerosol model [Takemura *et al.* 2005; 2009]. The model
21 experiments are those extended from the PDRMIP protocol, which increased BC and
22 sulfur dioxide (SO₂) emissions by ten times (x10) and five times (x5), respectively,
23 relative to the present emission conditions (control experiment). In this study, the BC and

SO₂ emissions are perturbed for their fuel sources by multiplying various globally uniform factors of 10, 5, 2, 1.5, 0.8, 0.5, 0.3, 0.1 and 0.0 to obtain more systematic and robust climate responses than PDRMIP (see Methods in *Takemura and Suzuki, 2019*) for these short-lived climate forcings with inhomogeneous spatial distributions. Simulations with increased solar insolation by +2% (denoted by Sol) and with CO₂ concentration doubled (denoted by CO₂) relative to the level in 2000 are also conducted but with no varying magnitudes for their relatively homogeneous nature. The change to quantity of interest caused by each of these forcings relative to the control experiment (denoted as Δ) is analyzed throughout this paper.

For each of the forcing, sea surface temperature (SST)-prescribed and ocean-coupled simulations are conducted to decompose the total climate response into its fast and slow response components. Following PDRMIP, the fast response is estimated from the SST-prescribed run and is subtracted from the total response obtained from the ocean-coupled run to infer the slow response as residuals. The SST-prescribed and ocean-coupled runs are performed for 15 and 100 years, respectively, and the last 10 and 50 years are used for analysis, respectively, also following the PDRMIP analysis. For all the simulations of this study, horizontal and vertical resolutions of the model are T85 and 40 layers, respectively.

3. Analysis

3.1 Hydrological sensitivity for different forcing agents

Figure 1 shows how the global-mean precipitation change (ΔP) is related to the global-mean surface air temperature change (ΔT_s) in response to the four forcings. The statistics for the total response (Fig. 1a) show how diverse the AHS is among different forcing

agents. The linear fittings to responses to varying SF and BC perturbations quantify their respective AHS to be $\sim +3.0\%K^{-1}$ for SF and $\sim -6.0\%K^{-1}$ for BC (as represented by blue and red lines in Fig. 1a). The substantial negative magnitude of the BC-induced AHS is consistent with PDRMIP [Samset *et al.* 2016]. The response to increased solar insolation is located close to the blue line, suggesting the Sol-induced AHS is more or less similar to the SF-induced AHS. This is reasonable given that the two forcings have common characteristics of modulating the solar energy input through either changing the insolation or the planetary albedo. The CO₂-induced AHS is somewhat smaller than the SF and Sol cases although still positive.

To exclude the fast response contributions, the statistics for the slow-response component alone is shown in Fig. 1b. The comparison of Figs. 1a and 1b highlights that a vast diversity of HS for the total response to different forcing agents (Fig. 1a) is remarkably narrowed down to a common value across various forcings when the slow response component is isolated (Fig. 1b). This is also consistent with findings of Richardson *et al.* [2018] for PDRMIP multi-model ensemble. In the following analysis, the precipitation responses are investigated in the context of global energy budget depicted in Fig. 10 of ST19.

3.2. Atmospheric energy balance

The atmospheric energy budget is given as the balance between the atmospheric radiative cooling change ($-\Delta R^{ATM}$) and the latent ($L\Delta P$; $L=2.5\times 10^6$ Jkg⁻¹) plus sensible (ΔS) heating change:

$$L\Delta P + \Delta S = -\Delta R^{ATM}. \quad (1)$$

Due to a relatively small heat capacity of atmosphere, the balance (1) also applies to

each of the fast and slow response components as (see Figure 10 of ST19)

$$L\Delta P_{fast} + \Delta S_{fast} = -\Delta R_{eff}^{ATM}, (2)$$

$$L\Delta P_{slow} + \Delta S_{slow} = -\Delta R_{slow}^{ATM}, (3)$$

where the subscript “fast” and “slow” denotes the fast and slow response components, respectively, of each quantity. R_{eff}^{ATM} represents the effective radiative forcing (ERF) to atmosphere, which is obtained as the net radiative imbalance of atmosphere in the SST-prescribed simulations. When the sensible heat (SH) contribution is neglected, the energy balances (2) and (3) reduce to the radiative-convective equilibrium (RCE) given as

$$L\Delta P_{fast} \sim -\Delta R_{eff}^{ATM}, (2)'$$

$$L\Delta P_{slow} \sim -\Delta R_{slow}^{ATM}. (3)'$$

Figure 2 shows how these relationships apply to the model output for the fast and slow response components to the four different forcings. Figure 2a illustrates that the latent plus sensible heat changes balance the atmospheric radiative cooling change for the fast response component, following (2). The latent heat (LH) change has a major contribution to this energy balance as illustrated in Fig. 2c, which depicts how RCE given by (2)' applies to most of the forcing agents except BC. In the BC-forced scenarios, SH significantly decreases in response to the BC-induced stabilized atmosphere to substantially deviate the balance relationship from the RCE.

Figures 2b and 2d show the corresponding balance relationships (3) and (3)', respectively, for the slow-response component. Similar to the fast-response case, the latent plus sensible heat change balances the atmospheric radiative cooling change (Fig. 2b). The RCE relationship is found to better apply to the slow-response component (Fig. 2d) for all the forcing agents including BC.

3.3. Scaling with global-mean temperature

a. Slow response

The slow response of atmospheric radiative cooling can further be decomposed into the clear-sky (denoted by the subscript “clr”) and cloud radiative effect (CRE; denoted by C) components as

$$\Delta R_{slow}^{ATM} = \Delta R_{slow,clr}^{ATM} + \Delta C_{slow}^{ATM}.$$

The clear-sky radiative cooling primarily comes from the column water vapor change [Stephens and Ellis, 2008] as

$$-\Delta R_{slow,clr}^{ATM} \sim \alpha \Delta W_{slow}/W,$$

where the coefficient α represents the radiative effect of water vapor as determined from the gas absorption parameterization of radiation physics [Stephens and Ellis, 2008; Stephens and Hu, 2010; Suzuki et al. 2017]. This relationship is illustrated in Fig. 3a that provides an approximate estimate of $\alpha \sim 0.34 \text{ W m}^{-2} \%^{-1}$. The water vapor change then occurs with ΔT_s through the Clausius-Clapeyron (C-C) relationship as

$$\Delta W_{slow}/W \sim \kappa \Delta T_s,$$

where $\kappa \sim 7.6\% \text{ K}^{-1}$ according to Fig. 3b. The clear-sky radiative cooling change therefore scales with the temperature response as

$$-\Delta R_{slow,clr}^{ATM} \sim \alpha \Delta W_{slow}/W \sim \alpha \kappa \Delta T_s.$$

The CRE component is also related to the global-mean temperature change as

$$\Delta C_{slow}^{ATM} \sim \mu \Delta T_s,$$

where μ represents the cloud radiative feedback onto atmosphere.

These arguments propose

$$-\Delta R_{slow}^{ATM} \sim (\alpha \kappa - \mu) \Delta T_s.$$

This relationship is illustrated in Fig. 4b to be independent of forcing agents, implying the universal nature of the cloud feedback and the water vapor-mediated radiative response for a given model with common physics. Similarly, the slow response of SH is also found to scale with the temperature change in Fig. 4d at a rate much smaller than for the radiative response (Fig. 4b). Denoting this relationship as $-\Delta S_{\text{slow}} \sim \nu \Delta T_s$, the slow response of the global-mean precipitation constrained by (3) scales with the temperature change as

$$L\Delta P_{\text{slow}} \sim (\alpha\kappa - \mu + \nu)\Delta T_s.$$

This provides a breakdown of the temperature-mediated slow response into components relevant to different physical processes.

b. Fast response

In contrast to the slow-response component, the fast-response component, by its definition, has no *direct* dependency on the global-mean temperature. However, the fast-response component relates to the global-mean temperature *indirectly* through linkages of the fast response of energy budget to the top-of-atmosphere radiative forcing that somehow induces the global-mean temperature change via the climate response as typically represented by the so-called feedback parameter. For the purpose of interpreting AHS containing both fast and slow responses, it is useful to investigate how the fast-response components, represented by each term of (2), relate to the global-mean temperature change.

Figure 4a shows the correlation between the atmospheric ERF and ΔT_s . The SF- and BC-induced ERFs are found to have starkly different correlations with ΔT_s . The SF forcing exerts a small magnitude of the atmospheric ERF with a wide range of the temperature response whereas the BC-induced atmospheric heating exerts significant

ERF on atmosphere for a small range of the temperature response. This difference can be primarily explained by different vertical structures of the instantaneous radiative forcing between the two aerosols although the $\text{ERF}-\Delta T_s$ correlations can also be influenced by aerosol impacts on clouds: BC heats atmosphere and cools surface whereas SF directly cools surface with little impact on atmosphere [ST19]. The atmospheric heating of BC triggers rapid adjustment of atmosphere that weakens the global-mean temperature response [Stjern *et al.* 2017] whereas the SF forcing produces tiny magnitudes of atmospheric ERF accompanied by pronounced temperature responses. Such characteristics are also hinted in the CO_2 - and Sol-induced responses in Fig. 4a where the Sol response is located closer to the SF relationship (blue line) than is the CO_2 case, reflecting their different energy redistributions into atmosphere and surface.

The corresponding relationships for the SH fast response also have starkly different relationships with ΔT_s between SF and BC (Fig. 4c), reflecting a significant decrease of SH in response to BC that induces stabilized atmosphere and a small change of SH due to the tiny impact of SF on atmosphere. The SH fast responses to CO_2 and Sol are also found to be similar to those to SF.

3.4. Contributions from various factors to AHS

Combining the fast and slow responses described above, the total response of the global-mean precipitation is expressed as

$$L\Delta P = L\Delta P_{fast} + L\Delta P_{slow} \sim -\Delta S_{fast} - \Delta R_{eff}^{ATM} + (\alpha\kappa - \mu + \nu)\Delta T_s,$$

or dividing by ΔT_s and decomposing ΔR_{eff}^{ATM} into the clear-sky and CRE components as

$$L\Delta P/\Delta T_s \sim -\Delta S_{fast}/\Delta T_s - \Delta R_{eff,clr}^{ATM}/\Delta T_s - \Delta C_{eff}^{ATM}/\Delta T_s + \alpha\kappa - \mu + \nu. \quad (4)$$

(4) provides a breakdown of AHS into contributions from different components of atmospheric energy budget relevant to different physical processes.

Figure 5 summarizes the AHS tendencies of different terms in (4) for the four forcing agents. It is found that SF and Sol have similar ingredients of contributions, which are dominated by the slow-response component of the clear-sky radiative cooling. This is mostly the only factor that determines AHS ($\sim 3\% \text{K}^{-1}$) for these two forcing agents and comes from the column water vapor change through the C-C relationship (Fig. 3). The values of $\alpha \sim 0.34 \text{Wm}^{-2}\%^{-1}$ and $\kappa \sim 7.6\% \text{K}^{-1}$ result the AHS tendency due to this component of $\sim 2.6 \text{Wm}^{-2}\text{K}^{-1}$ (or $\sim 2.8\% \text{K}^{-1}$). This sets a component common among the forcing agents analyzed and dominating the slow response (Fig. 5), explaining a universal nature of the pure HS in Fig. 1b. This component is also found to exert a large contribution to the CO_2 -induced AHS, which is partially offset by the negative tendency due to the clear-sky ERF induced by CO_2 atmospheric heating to result a net positive AHS of $\sim 1.2\% \text{K}^{-1}$ that is somewhat smaller than the SF- and Sol-induced AHS ($\sim 3\% \text{K}^{-1}$).

The BC-induced AHS is dominated by a large negative tendency due to the clear-sky ERF component associated with the atmospheric heating. This negative tendency is partially offset by fast responses of CRE and SH as well as by the slow response of the clear-sky radiative cooling to result a pronounced net negative AHS of about $-6\% \text{K}^{-1}$. The positive offset largely comes from the fast SH response, suggesting a significant role of SH in determining the precipitation response, as also pointed out by *Myhre et al.* [2018]. The slow response of the clear-sky radiative cooling, common among the forcing agents, exerts the second largest compensation for the clear-sky ERF. The positive tendency due to the fast response of CRE is also not negligible, suggesting a significance of cloud responses to BC. The positive and negative tendencies represent two opposing pathways

of precipitation responses to BC [Ming *et al.* 2010].

In MIROC, the fast response of CRE is negligible for forcing agents except BC but is subject to uncertainties arising from aerosol impacts on clouds for SF and BC cases. The slow-response component of CRE, representing the cloud feedback to the global precipitation, also has only a small negative contribution to AHS for all the forcing agents analyzed (Fig. 5). It should be noted, however, that there is no a priori reason that determines whether the cloud feedback on precipitation is positive or negative [Stephens and Hu, 2010; Pendergrass and Hartmann, 2014; Mauritsen and Stevens, 2015]. Also given the link of HS to the equilibrium climate sensitivity via the low-cloud feedback [Watanabe *et al.* 2018], the cloud feedback uncertainty and its influence on AHS warrant further investigations in future studies.

4. Conclusion

This study investigates how different forcing agents induce different AHS through different ingredients of contributions from various components of atmospheric energy budget. Relevance of energy budget components to global-mean temperature change is analyzed to quantify contributions of different components to AHS. Different forcing agents are found to have varying ingredients of the AHS tendencies due to different factors. In particular, SF and Sol primarily induce the slow response of radiative cooling that determines AHS of $\sim 3.0\%K^{-1}$ whereas BC induces a strong atmospheric heating, which is partially compensated for by the fast response of SH and CRE and the slow response of radiative cooling to result a significantly negative AHS ($\sim -6.0\%K^{-1}$). The CO₂-induced AHS is found in between ($\sim 1.2\%K^{-1}$) as determined by the slow response

of radiative cooling and its partial compensation by the atmospheric heating. This study offers a theoretical basis for interpreting the historical and future climatic changes of global precipitation driven by a mixture of various forcings.

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References

- Allen, M. R., and W. J. Ingram, 2002: Constraints on future changes in climate and the hydrologic cycle. *Nature*, **419**: 224-232.
- DeAngelis, A. M., X. Qu, M. D. Zelinka, and A. Hall, 2015: An observational radiative constraint on hydrologic cycle intensification. *Nature*, **528**: 249-253.
- Fläschner, D., T. Mauritsen, and B. Stevens, 2016: Understanding the intermodal spread in global-mean hydrological sensitivity. *J. Climate*, **29**, 801-817, doi:10.1175/JCLI-D-15-0351.1.
- Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686-5699.
- Mauritsen, T., and B. Stevens, 2015: Missing IRIS effect as a possible cause of muted hydrological change and high climate sensitivity in model. *Nature Geoscience*, **8**, 346-351.
- Ming, Y., V. Ramaswamy, and G. Persad, 2010: Two opposing effects of absorbing

- 1 aerosols on global-mean precipitation. *Geophys. Res. Lett.*, **37**, L13701,
2 doi:10.1029/2010GL042895.
- 3 Myhre, G., and Coauthors, 2017: PDRMIP: a precipitation driver and response model
4 intercomparison project – protocol and preliminary results. *Bull. Am. Meteorol.*
5 *Soc.*, **98**, 1185-1198.
- 6 Myhre, G., and Coauthors, 2018: Sensible heat has significantly affected the global
7 hydrological cycle over the historical period. *Nature Communications*, **9**:1922,
8 doi:10.1038/s41467-018-04307-4.
- 9 Pendergrass, A., and D. L. Hartmann, 2012: Global-mean precipitation and black
10 carbon in AR4 simulations. *Geophys. Res. Lett.*, **39**, L01703,
11 doi:10.1029/2011GL050067.
- 12 Pendergrass, A., and D. L. Hartmann, 2014: The atmospheric energy constraint on
13 global-mean precipitation change. *J. Climate*, **27**, 757-768, doi:10.1175/JCLI-D-
14 13-00163.1.
- 15 Richardson, T. B., and Coauthors, 2018: Drivers of precipitation change: An energetic
16 understanding. *J. Climate*, **31**, 9641-9657, doi:10.1175/JCLI-D-17-0240.1.
- 17 Samset, B. H., and Coauthors, 2016: Fast and slow precipitation responses to individual
18 climate forcers: a PDRMIP multimodel study. *Geophys. Res. Lett.*, **43**, 2782-2791,
19 doi:10.1002/2016GL068064.
- 20 Samset, B. H., and Coauthors, 2018: Weak hydrological sensitivity to temperature
21 change over land, independent of climate forcing. *npj Climate and Atmospheric*
22 *Science*, 1:20173, doi:10.1038/s41612-017-0005-5.
- 23 Shiogama, H., and Coauthors, 2010: Emission scenario dependency of precipitation on
24 global warming in the MIROC3.2 model. *J. Climate*, **23**, 2404-2417,

doi:10.1175/2009JCLI3428.1.

Stjern, C. W., and Coauthors, 2017: Rapid adjustments cause weak surface temperature response to increased black carbon concentrations. *J. Geophys. Res.*, **122**, 11462-11481, doi:10.1002/2017JD027326.

Stephens, G. L., and T. D. Ellis, 2008: Controls of global-mean precipitation increases in global warming GCM experiments. *J. Climate*, **21**, 6141-6155, doi:10.1175/2008JCLI2144.1.

Stephens, G. L., and Y. Hu, 2010: Are climate-related changes to the character of global-mean precipitation predictable? *Environ. Res. Lett.*, **5**, 025209, doi:10.1088/1748-9326/5/2/025209.

Suzuki, K., G. L. Stephens, and J.-C. Golaz, 2017: Significance of aerosol radiative effect on energy balance control on global precipitation change. *Atmos. Sci. Lett.*, **18**, 389-395, doi:10.1002/asl.780.

Suzuki, K., and T. Takemura, 2019: Perturbations to global energy budget due to absorbing and scattering aerosols. *J. Geophys. Res. Atmos.*, **124**, 2194-2209, doi:10.1029/2018JD029808.

Takemura, T., T. Nozawa, S. Emori, T. Y. Nakajima, and T. Nakajima, 2005: Simulation of climate responses to aerosol direct and indirect effects with aerosol transport-radiation model. *J. Geophys. Res.*, **110**, D02202, doi:10.1029/2004JD005029.

Takemura, T., M. Egashira, K. Matsuzawa, H. Ichijo, R. Oishi, and A. Abe-Ouchi, 2009: A simulation of the global distribution and radiative forcing of soil dust aerosols at the Last Glacial Maximum. *Atmos. Chem. Phys.*, **9**, 3061-3073, doi:10.5194/acp-9-3061-2009.

- Takemura, T., and K. Suzuki, 2019: Weak global warming mitigation by reducing black carbon emissions. *Scientific Reports*, **9**:4419, doi:10.1038/s41598-019-41181-6.
- Watanabe, M., and Coauthors, 2010: Improved climate simulation by MIROC5: mean states, variability, and climate sensitivity. *J. Climate*, **23**, 6312-6335, doi:10.1175/2010JCLI3679.1.
- Watanabe, M., Y. Kamae, H. Shiogama, A. M. DeAngelis, and K. Suzuki, 2018: Low clouds link equilibrium climate sensitivity to hydrological sensitivity. *Nature Climate Change*, doi:10.1038/s41558-018-0272-0.
- Wu, P., N. Christidis, and P. Stott, 2013: Anthropogenic impact on Earth's hydrological cycle. *Nature Climate Change*, **3**, doi:10.1038/NCLIMATE1932.

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Fig. 1: Scatter plot of fractional magnitudes of the (a) total and (b) slow responses of the global-mean precipitation (in %) as a function of the global-mean surface air temperature (T_s) change. The red, blue and black lines represent the linear fittings to responses to BC, SF and all the forcings, respectively. The leftmost blue point and rightmost red point represent the SFx10 and BCx10 cases, respectively.

Fig. 2: (Left) Scatter plot of the fast response of (a) latent plus sensible heat and (c) latent heat alone as a function of the effective radiative forcing for atmosphere. (Right) Scatter plot of the slow response of (b) latent plus sensible heat and (d) latent heat alone as a function of the slow-response component of the atmospheric radiative cooling. The dashed lines represent the one-to-one relationship. In (a) and (c), the rightmost blue points and leftmost red points represent SFx10 and BCx10 cases,

respectively. In (b) and (d), the leftmost blue points and rightmost red points represent SFx10 and BCx10 cases, respectively.

Fig. 3: Scatter plot of the (a) slow-response component of the clear-sky atmospheric radiative cooling as a function of the column water vapor (W) change and of the (b) column water (W) change as a function of the global-mean surface air temperature (T_s) change. The dashed lines represent the linear fittings to all the data with slopes of $-0.34\text{Wm}^{-2}\%^{-1}$ (a) and $7.6\%\text{K}^{-1}$ (b), respectively. The leftmost blue point and rightmost red point represent the SFx10 and BCx10 cases, respectively.

Fig. 4: (Left) Scatter plot of the fast response of (a) effective radiative forcing for atmosphere and (c) sensible heat as a function of the global-mean surface air temperature change. (Right) Scatter plot of the slow response of (b) atmospheric radiative heating and (d) sensible heat as a function of the global-mean surface air temperature change. The red, blue and black lines represent the linear fittings to responses to the BC, SF and all the forcing agents, respectively. The leftmost blue point and rightmost red point represent the SFx10 and BCx10 cases, respectively.

Fig. 5: Contributions from different energy components to AHS for the SF, BC, Sol and CO₂ scenarios. The “Net tendency” represents the combined contributions from all the components. The “AHS” denotes the AHS estimated from global-mean responses of precipitation and temperature.

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