Inferring Aerosol Forcing from Surface Temperature Record

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Abstract. The Earth's surface temperature has risen by about 0.8 K over the last 100 years partly due to human activities. If not for the poorly understood radiative effects of aerosols, one would have been able to better use the observed warming to constrain the all-important transient climate response (TCR, the pace of the warming caused by a continuous buildup of greenhouse gases). Here we show that it is feasible to infer the historical aerosol forcing and subsequently TCR from the surface temperature record by exploiting the seasonality difference between the aerosol and greenhouse gas forcings. Our analysis suggests that the seasonally-varying aerosol forcing may have played a crucial role in causing the Northern Hemisphere (NH) winter to warm at a faster pace than the NH summer – a distinct feature of the observed warming. The estimated 25% to 75% range is -1.7 – -0.7 W m^{-2} for the aerosol forcing, and is 0.9 – 1.7 K (at the time of CO_2 doubling) for TCR. The respective 5% to 95% ranges are -2.5 – 0.1 W m^{-2} and 0.7 – 3.8 K. The median TCR of 1.3 K is considerably smaller than those projected with state-of-the-art climate models or derived in attribution studies. The results are highly relevant to shaping the climate outlook for the coming decades.

1. Introduction

As the near-term climate change is strongly tied to the transient climate response (TCR), the inability of climate models to converge on TCR hinders a reliable projection of future climate. (In this work, TCR is defined as the surface temperature change in response to a 1% year^{-1} increase of CO_2 at the time of doubling.) The minimum and maximum values among the nineteen Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) models are 1.2 and 2.6 K, respectively, with a median of 1.6 K [*Randall et al.*, 2007]. (Note that this model ensemble does not provide an adequate sampling of model-simulated TCR, and the resulting range is particularly sensitive to outliers.) The inter-model spreads in cloud feedback and ocean heat uptake are among the main causes [*Soden and Held*, 2006; *Winton et al.*, 2001].

The alternative of inferring TCR from the past warming is plagued by the large uncertainty in the historical aerosol forcing [*Andreae et al.*, 2005]. The greenhouse gas forcing is 2.8 W m^{-2} in the year 2010 (http://www.esrl.noaa.gov/gmd/aggi/), and the forcing associated with doubling CO_2 is 3.7 W m^{-2}. One can write TCR approximately as $3.7\delta T/(2.8 + F_a\epsilon)$, where δT is the observed surface temperature change, and F_a and ϵ are the aerosol forcing and its efficacy [*Joshi et al.*, 2003] (defined as the ratio of the climate sensitivity for a non-CO_2 forcing to that for CO_2), respectively. [Note that this expression does not explicitly consider forcings other than greenhouse gases and aerosols (e.g., ozone, land use and solar), but their effects are likely to be small [*Forster et al.*, 2007].] The forward model estimates of a subset of known aerosol effects range from -2.2 to -0.5 W m^{-2} [*Forster et al.*, 2007]. The implied TCR could vary considerably depending

on the estimated aerosol forcing. Unfortunately, process-level studies cannot rule out the possibility of a strong aerosol cooling and thus a large TCR [*Lohmann et al.*, 2010].

Is it possible to constrain the aerosol forcing and TCR simultaneously with the surface temperature record? If the surface temperature responses to aerosols and greenhouse gases possess sufficiently different patterns, their magnitudes can be quantified with the optimal fingerprinting approach by regressing the observed temperature anomalies on the changes caused by individual forcings [*Stott et al.*, 2006]. Hence the effectiveness of the approach hinges on one's ability to realistically simulate the latter. Particularly challenging for climate models is the surface temperature response to the aerosol forcing, which, due to its spatially and temporally inhomogeneity, is more effective at inducing circulation changes than well-mixed greenhouse gases [*Ming and Ramaswamy*, 2011], an issue that we will return to in the discussion section. This paper presents an attempt at separating the climate impacts of the two forcings by exploiting their difference in seasonality.

2. Seasonality of Aerosol Forcing

We use a modified version of the GFDL-AM2.1 model [*Ming and Ramaswamy*, 2009] and the GFDL-AM3 model [*Donner et al.*, 2011] to calculate the pre-industrial to present-day anthropogenic aerosol forcing, evaluated as radiative flux perturbation [*Haywood et al.*, 2009]. Both direct (via scattering/absorbing solar radiation) and indirect effects (via enhancing cloud albedo and lifetime) are included. Two models are different in cumulus parameterization, aerosol transport, aerosol optical properties and aerosol-cloud interactions. GFDL-AM2.1 is also used to compute the greenhouse gas forcing.

Fig. 1 shows the normalized monthly mean aerosol forcing simulated with two climate models. (Note that the normalized monthly mean forcing or insolation is calculated as

$(X_i - \bar{X})/\bar{X}$, where X_i is the monthly mean for the i-th month and $\bar{X}$ is the annual mean ($\sum_{i=1}^{12} X_i/12$).) The seasonal cycle of the aerosol forcing – peaking in the boreal summer and tapering off to a wintertime minimum – is remarkably consistent between the models in spite of the numerous differences in model physics. Furthermore, it tracks closely the seasonal cycle of the NH insolation. This is because anthropogenic aerosols are situated primarily in the NH, and interact mainly with the shortwave (or solar) radiation. In contrast, the greenhouse gas forcing, which affects the longwave (or terrestrial) radiation, does not exhibit any seasonality. Assuming everything else being equal, one may speculate that the seasonally-varying aerosol forcing would cool the NH summer more than the NH winter. In the presence of a background greenhouse gas warming, aerosols would cause the NH winter to warm more (or, in a transient sense, faster) than the NH summer. The seasonality difference represents a physical mechanism through which the climate impacts of aerosols and greenhouse gases can be potentially separated from each other.

3. Surface Temperature Record

This thinking leads us to examine the hemispheric seasonal warming trends using the Goddard Institute for Space Studies (GISS) surface temperature analysis (GISTEMP) [*Hansen et al.*, 2010]. We choose the analysis period of 1911-2010 for two reasons. First, the data prior to 1911 may not be as reliable. Second, the 100-year span is long enough to keep the role of natural variability relatively small. [Analyses for other periods (e.g., 1881-2010 and 1941-2010) yield the same basic findings.] The least-squares linear trends of the NH winter (December-January-February, denoted as δT_w) and summer (June-July-August, denoted as δT_s) surface temperature are 0.89 and 0.61 K (100 years)$^{-1}$, respectively (Fig. 2), both of which are statistically significant at the 95% confidence level

based on the two-tailed Student's t-test [*Woodward and Gray*, 1993; *Santer et al.*, 2000]. Stronger NH wintertime warming was also seen for other observational datasets [*Wallace et al.*, 1995; *Balling et al.*, 1998; *Stine et al.*, 2009].

Can one explain the seasonal contrast inherent in the observed warming by invoking the natural fluctuations intrinsic to the climate system? We construct the cumulative probability distribution of the seasonal difference in the unforced 100-year NH temperature trend ($\delta T_w - \delta T_s$, denoted as $\mathcal{T}$) from randomly sampling a 4000-year control simulation performed with the GFDL-CM2.1 model [*Delworth and Coauthors*, 2006] (a Bootstrap method [*Vecchi et al.*, 2006]). The distribution is shown in the upper panel of Fig. 3. The observed $\mathcal{T}$ of 0.27 K (100 years)$^{-1}$ lies outside the 95% confidence interval [-0.24 – 0.24 K (100 years)$^{-1}$], suggesting that it is most likely forced by external forcings.

4. Role of Greenhouse Gases

The more rapid NH wintertime warming has been discussed in the general context of global warming [*Balling et al.*, 1998]. This is, to a certain extent, consistent with the IPCC AR4 model simulation of the climate response to a 1 % year^{-1} increase of CO_2, which are archived at the Program for Climate Model Diagnosis and Intercomparison (PCMDI) of the US Department of Energy Lawrence Livermore National Laboratory (DOE/LLNL). All the models agree that a steady rise in the greenhouse gas concentration would cause the NH winter to warm at a faster pace than the NH summer. They, however, diverge substantially in the strength of the effect. The normalized $\mathcal{T}$ [$2(\delta T_w - \delta T_s)/(\delta T_w + \delta T_s)$, denoted as $\tilde{\mathcal{T}}$] due to CO_2 has a mean of 0.22 with a variance of 0.09 (the lower panel of Fig. 3). The observed $\tilde{\mathcal{T}}$ (0.37) deviates from the ensemble-mean by 0.15, which is only slightly less than two variances. In addition, only one out of a total of seventeen models

yields a $\tilde{\mathcal{T}}$ that is greater than the observation. These are strong indications that forcing agents other than greenhouse gases may have also contributed to the seasonal difference in the past warming. It is plausible to assume that, in light of its strong seasonality, the aerosol forcing is responsible for the fraction of the observed seasonal difference that cannot be accounted for by greenhouse gases.

5. A Conceptual Model of Aerosol Forcing

This argument motivates us to derive a mathematical expression for the historical aerosol forcing. If the observed warming is driven entirely by greenhouse gases, the seasonal difference in the NH warming ($\mathcal{T}_g$) would be approximated as $0.5\tilde{\mathcal{T}}_g(\delta T_w + \delta T_s)$, where $\tilde{\mathcal{T}}_g$ is the normalized $\mathcal{T}$ in response to greenhouse gases alone and is estimated at 0.22 ± 0.09 (one variance) from the IPCC AR4 model ensemble (the lower panel of Fig. 3). The influence of aerosols ($\mathcal{T}_a$) is calculated as the residual of subtracting $\mathcal{T}_g$ from the observed $\mathcal{T}$.

The seasonal cycle of the solar input is the main driving force behind that of the surface temperature. In the present climate state, the seasonal difference in the climatological NH top-of-the-atmosphere solar absorption ($\delta\mathbf{F}$) gives rise to a seasonal difference in the NH surface temperature ($\delta\mathbf{T}$). In an idealized climate state with time-invariant solar input, the surface temperature would not vary temporally either. We assume that the climate system, when being subjected to a change in the seasonal cycle of the solar input, would behave linearly by following the slope set by the two states. It then follows that $\mathcal{F}_a$, which is the seasonal difference in the NH aerosol forcing ($F_{a,w} - F_{a,s}$), and can be thought of as a perturbation to the solar input, is equal to $\mathcal{T}_a \delta\mathbf{F}/\delta\mathbf{T}$. (Because the phase of the seasonal cycle of the surface temperature lags that of the solar input roughly by one month as

a result of thermal inertia, both $\mathcal{F}_a$ and $\delta\mathbf{F}$ contrast November-December-January with May-June-July.) $\delta\mathbf{F}$ is 153.3 W m^{-2} according to the 5-year Clouds and Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) dataset [*Loeb et al.*, 2009], and $\delta\mathbf{T}$ is 12 K according to the 30-year surface temperature climatology constructed by the University of East Anglia Climate Research Unit (CRU) [*Jones et al.*, 1999]. The respective variances owing to natural variations are estimated at 0.4 W m^{-2} and 0.1 K (see SOM).

Because $F_{a,s}$ is about 2.2 times of $F_{a,w}$, $\mathcal{F}_a$ is equal to $1.2F_{a,w}$. The NH annual mean aerosol forcing, which can be evaluated as $0.5(F_{a,w} + F_{a,s})$ or $1.6F_{a,w}$, is about $1.3\mathcal{F}_a$. Since anthropogenic aerosols are mainly in the NH, the global mean aerosol forcing (F_a) is approximately half of the NH forcing (i.e., $0.65\mathcal{F}_a$). We then arrive at a one-line equation for F_a, which is $0.65[(1 - 0.5\tilde{\mathcal{T}}_g)\delta T_w - (1 + 0.5\tilde{\mathcal{T}}_g)\delta T_s]\delta\mathbf{F}/\delta\mathbf{T}$. The implied TCR can be computed from F_a using the aforementioned equation $[3.7\delta T/(2.8 + F_a\epsilon)]$. The forcing efficacy ($\epsilon$) is estimated at 0.7 with a variance of 0.05 [*Forster et al.*, 2007]. The observed 1911-2010 global mean surface temperature change (δT) is 0.76 K based on GISTEMP.

Our method for determining the aerosol forcing and TCR is deterministic in nature. The mean values of δT_w, δT_s and δT are based on GISTEMP. Their natural variations are estimated from randomly sampling the GFDL-CM2.1 control simulation described above. The probability distributions of $\tilde{\mathcal{T}}_g$ and ϵ are assumed to be Gaussian. Monte Carol simulations are performed to generate the probability distributions of the aerosol forcing and TCR.

Fig. 4 contains the probability distributions of the aerosol forcing (the upper panel) and TCR (the lower panel) generated from propagating the natural variations in the surface

temperature changes (δT_w, δT_s and δT) and the estimated uncertainties in $\tilde{\mathcal{T}}_g$, $\delta\mathbf{F}$, $\delta\mathbf{T}$ and ϵ through the equations derived above. The median value of F_a is -1.2 W m^{-2}, with a 5% to 95% range of -2.5 – 0.1 W m^{-2}. The values are generally consistent with the prevailing view that aerosols pose a net cooling effect on the climate. Our method for inferring F_a does not explicitly preclude the possibility of an aerosol cooling that is strong enough to overwhelm the greenhouse gas warming. In that case, the net historical forcing would be negative, thereby implying a negative climate sensitivity. The method is well-behaved in the sense that no solution resides in this unphysical regime. It is also reassuring to see that our results agree broadly with the forward climate model calculations (a median of -1.3 W m^{-2} and a 5% to 95% range of -2.2 – -0.5 W m^{-2}) [*Forster et al.*, 2007]. Nonetheless, this agreement should be interpreted with caution as the two sets of results are not directly comparable. For example, the former implicitly accounts for all aerosol effects, while the latter considers only a subset.

The probability distribution of TCR is skewed heavily toward the low end (less than 1 K), with a long tail extending to 3 – 4 K. This distribution reflects the fact that TCR is inversely proportional to the net historical forcing, and the latter (or the aerosol forcing) is determined independently of TCR in our method. The 90% confidence interval of 0.7 – 3.8 K is slightly wider than the likely range of 1 – 3.5 K given in AR4 [*Hegerl et al.*, 2007], and encompasses those of individual attribution studies [*Stott et al.*, 2006; *Forest et al.*, 2006; *Knutti and Tomassini*, 2008; *Gregory and Forster*, 2008; *Gillett et al.*, 2011; *Libardoni and Forest*, 2011; *Padilla et al.*, 2011]. The probability distributions, however, are very different. Our median TCR of 1.3 K is significantly smaller than the previously reported values (i.e., 2.1 K [*Stott et al.*, 2006], 1.9 K [*Forest et al.*, 2006] and 1.6 K [*Knutti*

and Tomassini, 2008]). It is even smaller than the 5% percentiles of most existing studies [*Stott et al.*, 2006; *Forest et al.*, 2006; *Gregory and Forster*, 2008; *Gillett et al.*, 2011; *Padilla et al.*, 2011]. A series of sensitivity experiments are used to assess how the results may vary with the estimated natural variations in the surface temperature trends and the estimated uncertainties in $\tilde{\mathcal{T}}_g$ and ϵ. Halving the estimated natural variations in δT_w and δT_s (Sensitivity Case #1) reduces the 95% percentile to 2.5 K, while halving the estimated uncertainty in $\tilde{\mathcal{T}}_g$ (Sensitivity Case #2) decreases it to 2.8 K. (Note that the natural variations in the UKMO-HadCM3 model are about half of those in the GFDL-CM2.1.) Neither the 5% percentile nor the median is substantially changed in these sensitivity experiments. This indicates that the low end of the estimated TCR is rather robust.

6. Discussion

Our results indicate that the magnitudes of the aerosol forcing and TCR may have been systematically overestimated in the attribution studies. As explained before, the fingerprinting approach employed in these studies relies heavily on the model-simulated surface temperature response to aerosols. The inter-hemispheric asymmetry in the aerosol forcing tends to prompt a net cross-equatorial energy transport, which acts to compensate for the NH radiative cooling, and in doing so, spread its impact to the Southern Hemisphere [*Ming and Ramaswamy*, 2011]. It has been shown, however, that the degree of the compensation varies considerably with some highly uncertain aspects of model physics such as cumulus parameterization and cloud feedback [*Kang et al.*, 2008]. In some of the AR4 historical simulations, the NH warms much less than the SH. This is contrary to the observations, and appears to indicate that the aerosol-induced cooling is to a large extent

confined within the NH due to rather weak compensation in those models. Given that the NH/SH contrast in warming has been deemed particularly important for constraining the aerosol forcing [*Stott et al.*, 2006], the discrepancies among models in the response to an asymmetric forcing would render some of them unfit for being used to detect aerosols.

In comparison, our method for separating the climate impacts of aerosols and greenhouse gases is based primarily on a physically sound mechanism, namely the seasonality difference between two forcings. Unlike the fingerprinting approach, it does not use the model-simulated response to aerosols. Adding to the evidence of serious model deficiency in simulating the response to aerosols is the fact that the AR4 models generally cannot capture the observed changes in the amplitude and phase of the seasonal cycle [*Stine et al.*, 2009]. This raises the doubt that the attribution studies based on these models may be unable to take into account the mechanism identified here.

We conclude by emphasizing that our method, which is broad-brush in nature, involves a number of assumptions. Chief among them is that the aerosol-induced change in the seasonal cycle would follow the same proportionality constant as the full seasonal cycle itself. One cannot discount the possibility that the seasonal cycle may respond differently to the inhomogeneous aerosols and to the relatively smooth insolation. Unfortunately, we cannot use existing climate models to test this assumption due to their aforementioned deficiencies in capturing many aspects of the climate response to aerosols.

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Figure 1. Normalized monthly mean forcing or insolation.

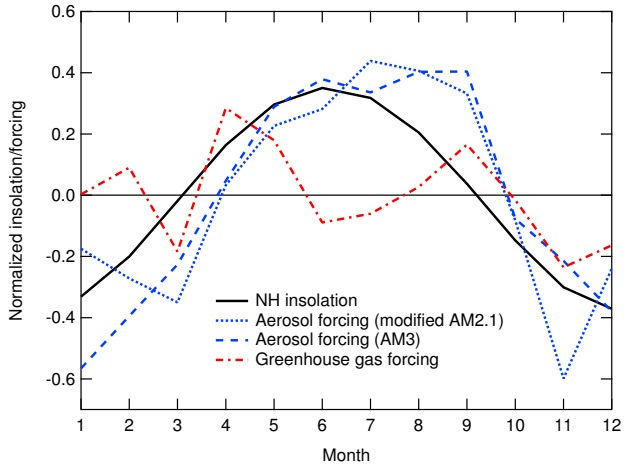

Figure 2. 5-year running means of the NH seasonal surface temperature anolmalies (solid lines) and the corresponding least-squares linear trends (dotted lines).

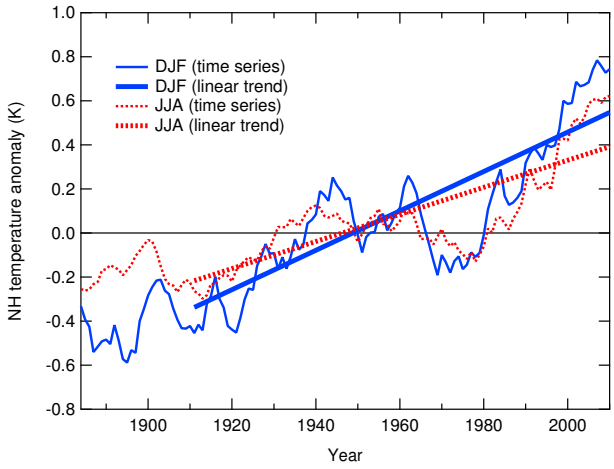

Figure 3. Cumulative probability distribution of the seasonal difference in the unforced 100-year NH temperature trend created from randomly sampling a 4000-year control simulation (the upper panel) and the normalized seasonal difference in the CO_2-induced warming (the lower panel).

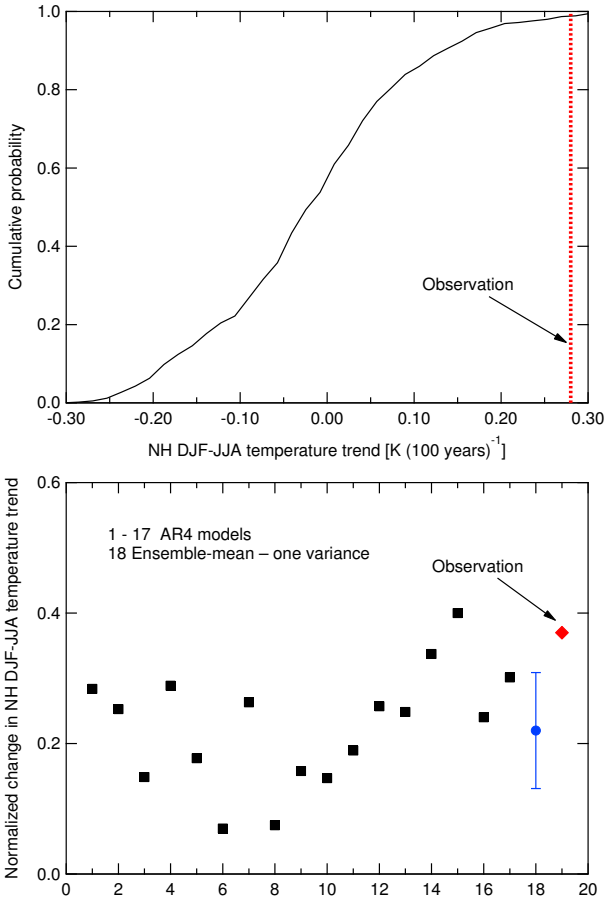

Figure 4. Probability distributions of the inferred aerosol forcing (the upper panel) and transient climate response (the lower panel). See the text for the description of the sensitivity cases. The horizontal bars represents the 5% to 95% ranges.