Sensitivity of Twentieth-Century Sahel Rainfall to Sulfate Aerosol and CO₂ Forcing

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ABSTRACT

A full understanding of the causes of the severe drought seen in the Sahel in the latter part of the twentieth-century remains elusive some 25 yr after the height of the event. Previous studies have suggested that this drying trend may be explained by either decadal modes of natural variability or by human-driven emissions (primarily aerosols), but these studies lacked a sufficiently large number of models to attribute one cause over the other. In this paper, signatures of both aerosol and greenhouse gas changes on Sahel rainfall are illustrated. These idealized responses are used to interpret the results of historical Sahel rainfall changes from two very large ensembles of fully coupled climate models, which both sample uncertainties arising from internal variability and model formulation. The sizes of these ensembles enable the relative role of human-driven changes and natural variability on historic Sahel rainfall to be assessed. The paper demonstrates that historic aerosol changes are likely to explain most of the underlying 1940–80 drying signal and a notable proportion of the more pronounced 1950–80 drying.

1. Introduction

The Sahel region of Africa experienced a severe reduction in wet season rainfall (July–September) from the 1970s to the 1990s, from which there has been a partial recovery. Numerous studies have found that the direct cause of the drought largely arises from changes in sea surface temperature (SST) patterns, with all ocean basins being important. The role of Atlantic SSTs was revealed first, with a tendency for dry Sahel years to be associated with anomalously cool SSTs in the northern (tropical) Atlantic and anomalously warm SSTs in the South Atlantic, particularly including the Gulf of Guinea (e.g., Lamb 1978; Vizy and Cook 2001; Lau et al. 2006). Warmer SSTs in the Indian Ocean have also been linked with drought over the Sahel (see, e.g., Palmer 1986; Hagos and Cook 2008; Lu 2009), along with a large-scale SST gradient into the western Pacific (Rowell 2001). ENSO events are influential on interannual time scales (e.g., Folland et al. 1991; Janicot et al. 1996; Rowell 2001), and more recently a contribution to drought of decadal variability in the Mediterranean has been confirmed (Rowell 2003; Jung et al. 2006). These anomalous regional SST patterns also contribute to a global “interhemispheric thermal contrast” pattern of SSTs that has been linked with Sahel drought, particularly on decadal time scales (e.g., Folland et al. 1986).

A key question, however, is whether these SST patterns are due to natural variability or anthropogenic forcing. This remains an issue of some debate. Sutton and Hodson (2005), Hoerling et al. (2006), and Ting et al. (2009) suggest that Atlantic SST changes (and hence the drought) were primarily of natural origin, whereas Held et al. (2005), Biasutti and Giannini (2006), Lau et al. (2006), and Kawase et al. (2010) suggest that observed precipitation changes were due to a combination of external anthropogenic forcing and internal (natural) variability. Held et al. (2005) analyzed initial condition ensembles of two versions of a coupled model, and concluded that the drying trend over the Sahel is partly anthropogenically forced (aerosol loadings and greenhouse gas emissions) and partly due to internal variability of the ocean–atmosphere climate system. Biasutti and Giannini

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(2006) use 19 models from the third Coupled Model Intercomparison Project (CMIP3) archive, and reach a similar conclusion, but also estimate that the fraction of the 1930–99 drying trend that was externally forced (primarily by anthropogenic activity) was “at least 30%.” Lau et al. (2006) analyzed the same 19 models, but focused on the period of drought from the 1970s to 1990s, and speculated that the increase in both volcanic and anthropogenic aerosols may have been responsible.

Sulfate aerosol, derived from anthropogenic activity and concentrated mainly in the Northern Hemisphere (NH), increased rapidly from 1940 to 1990 and could well have contributed significantly to the cross-equatorial SST gradient, which substantially altered low-latitude circulation and rainfall (Hulme and Kelly 1993; Williams et al. 2001; Rotstayn and Lohmann 2002; Baines and Folland 2007). However, the cross-equatorial SST changes have been attributed to natural modes of ocean variability by Sutton and Hodson (2005), Hoerling et al. (2006), and Ting et al. (2009), suggesting that anthropogenic activity alone is unlikely to be the sole cause of the Sahel drying. Another process that has been discussed as playing a role in the historical drought is a direct local radiative forcing from dust aerosol, which may have amplified the SST-induced drying in the Sahel (Yoshioka et al. 2007). However, there is currently no multimodel agreement on the sign of the dust impact. This, and other processes (such as carbonaceous aerosol and changing land use), are beyond the scope of this study.

Resolving this discussion on the possible causes of the recent SST variability (that has influenced Sahel precipitation) is important, not only for attributing the prolonged drought of recent decades, but also for improving predictions of future anthropogenic change in this region (see Cook 2008). One possible explanation for the lack of scientific consensus may be due to the limited ensemble sizes of available models to attribute the natural versus forced contributions. Another may be that many of the currently available CMIP3 models do not yet represent the aerosol indirect effects very well [which have been shown to be important in Williams et al. (2001)].

Sulfate aerosols influence climate directly (via the direct radiative effect) where aerosol particles act to scatter or absorb radiation. Aerosol particles also influence the microphysical properties of cloud and induce an indirect radiative effect. There are two indirect radiative effects; the ability of aerosol to change cloud albedo by increasing or decreasing the number of cloud droplets, which can subsequently scatter more or less solar radiation, making the cloud more or less reflective (“the first indirect effect,” assuming a constant cloud liquid water content; see Twomey (1977)), as well as changing the precipitation characteristics of clouds and increasing cloud lifetime [“the second indirect effect”; Albrecht (1989)]. Cooling of the climate system is caused by both the direct and indirect radiative effects of sulfate aerosol, which also contribute to the “global dimming” phenomenon (see Stanhill and Cohen 2001; Romanou et al. 2007). However, modeling work by Jones et al. (2001) suggests that the first indirect effect may well have a greater impact on climate than the direct effect. The first indirect effect is represented in the model by

\[
\text{CDNC} = 3.75 \times 10^8 \left[1 - \exp(-2.5 \times 10^{-9} \text{NCCN}) \right], \tag{1}
\]

where CDNC is the cloud droplet number concentration and NCCN is the number concentration of sulfate aerosol particles (cloud condensation nuclei). Since the relationship in Eq. (1) is nonlinear, the amount of sulfate in the initial (or control) state is important in governing the effects of changes in atmospheric sulfate concentrations on the surface air temperature. Jones et al. (2001) compared these relative aerosol effects in general circulation model (GCM) simulations. They found that by reducing the natural dimethyl sulfide (DMS) emissions in the preindustrial simulation by 50%, the indirect radiative forcing (for present-day relative to preindustrial sulfate concentrations) became 25% stronger. Therefore, the lower the initial concentrations of aerosol present, the more susceptible a cloud is to being influenced by an increase in aerosol concentration and the stronger its influence on climate. The effects of cloud susceptibility can be seen in “ship track” observations by Ackerman et al. (1995).

Despite the strong cooling influence of sulfate aerosols on surface air temperatures, confirmation of their influence on Sahel rainfall by changing hemispheric temperatures is difficult without a large ensemble of models. The ensemble size of the studies cited earlier limits their estimates of the relative importance of different forcings, due to problems in detecting the forced signal against the background of internal variability. The climateprediction.net experiment (Stainforth et al. 2005) uses a distributed computing platform to perform tens of thousands of climate simulations on the home computers of volunteers worldwide. Therefore, the results from the climateprediction.net project are uniquely placed to separate the signal of the sulfate aerosol and greenhouse gas forcing from the effects of natural (and model internal) variability. Furthermore, the wide variety of model climates that are created by the “perturbed physics” approach employed by the climateprediction.net experiment also goes some way to improving on the limited availability of models in earlier studies.

This study concentrates on the impacts of sulfate aerosol contrasted with the response to greenhouse gas forcing.
Two types of experiment are analyzed. In section 2, we investigate the response of a slab ocean GCM to changes in greenhouse gas and sulfur dioxide (SO2) emissions. These provide a conceptual understanding of the relative roles of both aerosols and greenhouse gases in Sahel rainfall, and are used to study climate anomalies in both the present and future climates. The results from these relatively simple experiments are then used to gain insight into the twentieth-century ensemble experiments with a fully coupled ocean–atmosphere GCM, described in section 3. Here, we make use of two sets of ensemble simulations of the twentieth century to explore relationships between changes in aerosol concentration, and impacts on Sahel rainfall using two ensembles of fully coupled ocean–atmosphere GCMs. The first ensemble represents the transient climate response to historical changes in greenhouse gases and sulfate aerosols as well as volcanic and solar forcing. The second ensemble makes use of an ensemble with identical forcings except that the sulfur emissions correspond to a period shifted 70 yr into the future. The combination of all ensembles uniquely enables us to explore the dependence of Sahel drying on sulfate aerosol concentrations (over and above model internal variability) and responses to other forcing mechanisms, such as an increase in greenhouse gases. The conclusions from this study are given in section 4.

2. Slab-ocean GCM experiments

a. Sulfur cycle background and experiments

In these experiments, the third Hadley Centre Atmospheric Model is coupled to both a slab ocean model (HadSM3) and a fully interactive sulfur cycle scheme, as documented in Jones et al. (2001). However, in this study there was no representation of the second indirect effect [unlike in Jones et al. (2001)] since HadSM3 did not have a prognostic ice–cloud scheme. Two gaseous sulfur species are emitted: SO2 and dimethyl sulfide. Sulfur dioxide is derived from anthropogenic activity as well as volcanic emissions within the model, and DMS, emitted from all ocean basins, represents the sulfur produced from the metabolic processes of phytoplankton. DMS is emitted from the surface of the model, whereas anthropogenic SO2 emissions occur from both the surface and between levels 3–5 in the model (to represent “high-level chimney stack” emissions). Volcanic SO2 is emitted into the model between the levels at 870 and 422 hPa.

To form sulfate aerosol, the sulfur gases need to be oxidized. This occurs in cloud-free regions via dry oxidation or within cloud and precipitation droplets via aqueous oxidation. These oxidation processes lead to the formation of three modes of sulfate aerosol, which are Aitken mode (median radius, \( r = 0.024 \) \( \mu m \)), accumulation mode (\( r = 0.095 \) \( \mu m \)), and dissolved sulfate.

Both gaseous sulfur species, as well as the sulfate aerosol, interact with the model meteorology (such as cloud and precipitation) and are transported throughout the domain using the model’s tracer advection scheme. All sulfur species are subjected to removal via dry and wet deposition (“scavenging”), except DMS, which is treated as insoluble. The balance between emissions, conversion, and deposition leads to an atmospheric loading of sulfate known as the “sulfate burden,” which will be used in the rest of this paper [for more detailed information on the climateprediction.net sulfur cycle experiment, see Ackerley et al. (2009)].

An ensemble of 15-yr experiments was run with HadSM3 within the climateprediction.net distributed computing framework. Also, a smaller ensemble of 30-yr experiments was run using the computing resource at the Rutherford Appleton Laboratory (RAL). Table 1 contains the boundary condition forcings applied to each phase of the experiment, with the numbers in parentheses denoting the number of model runs attained. The control runs (to which the subsequent forcing experiments
that precipitation in the Sahel may be influenced by the changes in interhemispheric temperatures and so we start by analyzing the global and hemispheric 1.5-m surface air temperature (SAT) responses to each of the boundary condition forcings (see Table 1), which are given in Table 2.

The results in Table 2 highlight the global mean cooling effects of sulfates and the warming influence of increased CO$_2$ (when perturbed individually). The global mean temperature (GMT) responses in the combined forcing experiments are (approximately) a linear addition of the responses to each individual forcing, which suggests there is little nonlinear interaction resulting from perturbing CO$_2$ and sulfate concentrations simultaneously, in these experiments.

Also included in Table 2 are the hemispheric changes in SAT, which highlight the inhomogeneous response of SAT to a sulfate and/or CO$_2$ forcing. In PD$_{SO2}$, the global mean sulfate cooling is dominated by the response in the NH (also see Williams et al. 2001; Feichter et al. 2004; Kristjánsson et al. 2005), which cools by 1 K compared to 0.3 K in the SH. A similar and opposite response also occurs in PD$_{SO2}$, which displays more warming in the NH than in the SH although the interhemispheric temperature difference is not as large as in PD$_{SO2}$ (see Table 2). The combined forcing experiment (PD$_{all}$) has approximately no global mean change in SAT but the NH cools by 0.3 K and the SH warms by 0.2 K, which is similar to the effect seen in PD$_{SO2}$ (NH cooling more than the SH).

The results are very different in the FUT experiments however. The sulfate cooling in FUT$_{SO2}$ is much smaller than in PD$_{SO2}$ and is similar between the NH and SH, whereas the warming in the FUT$_{CO2}$ experiment is much stronger in the NH than the in SH. The strong NH CO$_2$ warming relative to the SH is also apparent in FUT$_{all}$ but is reduced slightly due to the presence of increased anthropogenic SO$_2$ emissions. The strong NH warming in these runs (despite the globally uniform CO$_2$ concentrations used in this study) is due to a combination of the sea ice–albedo feedback mechanism (see Curry et al. 1995) and strong heating of the NH continental interiors as discussed in Sutton et al. (2007).

The results in this section clearly show that the responses to sulfate and CO$_2$ forcing differ between the NH and the SH, which will have implications for the seasonal migration of the intertropical convergence zone (ITCZ) and may therefore influence Sahel rainfall.

c. Zonal mean surface air temperature response

To show the asymmetrical properties of the surface air temperature response to each of the sulfate and/or
CO₂ forcings further (introduced in Table 2), the zonal, annual ensemble mean temperature responses for Hadley Centre–Climate Research Unit historical temperature data (HadCRU; Brohan et al. 2006) and each experiment are shown in Fig. 3. Figures 3a and 3b show the changes in zonal SAT from the HadCRU dataset, for the period 1860–1980, globally and in the Atlantic basin, respectively. The HadCRU data are not distributed uniformly and so do not provide a direct comparison to the model data but both plots (Figs. 3a and 3b) clearly show a preferential warming of the SH relative to the NH (south of approximately 50°N).
Having identified the larger warming in the SH compared to the NH in HadCRU, we now consider the model SAT changes for each of the forcing experiments globally and within the Atlantic basin. For PD$_{CO2}$ (Figs. 3c and 3d, solid lines), the zonal mean SAT increases at all latitudes but shows a slightly higher increase in the NH, centered around 50°N whereas the zonal mean change in SAT over the Atlantic is reasonably uniform between 650°N and 50°S. The opposite is true for PD$_{SO2}$ (Fig. 3c, dotted–dashed lines) where there is strong cooling of the NH due to the presence of anthropogenic sulfate aerosol. Unlike in PD$_{CO2}$, PD$_{SO2}$ has a stronger cooling in the NH than the SH over the Atlantic (Fig. 3d). In PD$_{all}$ (Fig. 3c, dashed lines), the SH (NH) zonal mean changes in SAT are generally positive (negative) and there is also a warming (cooling) at most latitudes in the SH (NH) Atlantic basin (Fig. 3d). So, despite the warming influence of increased CO$_2$, the NH cools relative to the SH with the presence of sulfate aerosol and more closely resembles the HadCRU data in Figs. 3a and 3b.

The present-day to future hemispheric temperature response therefore is dominated by the CO$_2$-induced warming whereas the preindustrial to present-day response is dominated by the asymmetrical influence of anthropogenic aerosol cooling the NH relative to the SH. The surface air temperature responses observed in this section will help explain the precipitation response to each of the imposed forcings.

d. Precipitation response

Figure 4 shows the zonal annual mean precipitation responses to each of the forcing experiments given in Table 1. The effects of CO$_2$ warming in PD$_{CO2}$ cause an increase in the NH tropical precipitation and a slight decrease in the SH tropical precipitation (in the zonal

![Figure 2](image-url)
Fig. 3. The zonal mean change in SAT (K) from HadCRU between 1860 and 1980 for the average (a) along all meridians and (b) for the Atlantic basin only. The model zonal mean surface air temperature responses (K) for (c) PD relative to PI globally, (d) PD relative to PI in the Atlantic basin, (e) FUT relative to PD globally, and (f) FUT relative to PD in the Atlantic basin. The ensemble mean is taken for experiments consisting of more than one run (also see Table 1).
mean, Fig. 4a). Including anthropogenic sulfates in PD₅₀₂ and PD₅all causes strong drying in the NH tropics and a moistening in the SH tropics (see Fig. 4a). Studies by Rotstayn et al. (2000), Feichter et al. (2004), and Kristjánsson et al. (2005) have also observed similar effects of NH aerosol loading on tropical precipitation and support the results in this study, indicating that the strong cooling in the NH causes a southward shift in the model ITCZ.

The opposite is true in the FUT CO₂ and FUTall experiments, which both display a strong increase in NH tropical precipitation with a smaller decrease in the SH tropical precipitation (see Fig. 4a). Studies by Rotstayn et al. (2000), Feichter et al. (2004), and Kristjánsson et al. (2005) have also observed similar effects of NH aerosol loading on tropical precipitation and support the results in this study, indicating that the strong cooling in the NH causes a southward shift in the model ITCZ.

The results of the simple thermodynamic slab-ocean model demonstrate the differing effects on tropical and Sahel precipitation and give a conceptual background as to how the Hadley Centre's climate model responds to each forcing. The response of these slab models can be seen as a form of idealized climate response, where both the response time scale and the influence of the full ocean-driven natural variability can be neglected. When exploring the implications of drivers of historic Sahel drying however, we need to account for these processes. In the next section we use large ensembles of fully coupled ocean–atmosphere GCMs, which enable the forced climate responses to be analyzed against the background of the ocean-driven variability and response time scale.

The Sahel is defined here to be a rectangle between 10°–18°N and 20°W–22°E (Ruosteenoja et al. 2003) and we focus on the June–July–August (JJA) period when the precipitation is high in this region. The mean change in precipitation associated with each of the forcing experiments in Table 1, relative to their respective control runs in JJA, can be seen in Fig. 5. The hemispheric impacts of increasing the sulfate loading alone from PI in PD₅₀₂ (Fig. 5a) acts to decrease the precipitation throughout the Sahel, which can also be seen in Fig. 5c when CO₂ concentrations are also increased to present-day levels (PD₅all). However, there is little change in precipitation throughout the Sahel when CO₂ alone is increased from PI to PD (PD₅₀₂ (Fig. 5b)), which indicates that sulfate aerosol is having the dominant effect on precipitation in this region, in these experiments.

When considering the “future” forcing scenarios (FUT experiments in Table 1), the precipitation responses are very different. The drying associated with the sulfate cooling in FUT₅₀₂ (Fig. 5d) is closer to the equator, with very little change in Sahel rainfall. However, the effects of doubling CO₂ concentrations with (FUT₅all; see Fig. 5f) and without (FUT₅₀₂; see Fig. 5e) changes in anthropogenic SO₂ emissions cause strong increases in precipitation throughout most of the Sahel and hint at a possible rainfall recovery relative to PD₅all.

The results of the simple thermodynamic slab-ocean model demonstrate the differing effects on tropical and Sahel precipitation and give a conceptual background as to how the Hadley Centre's climate model responds to each forcing. The response of these slab models can be seen as a form of idealized climate response, where both the response time scale and the influence of the full ocean-driven natural variability can be neglected. When exploring the implications of drivers of historic Sahel drying however, we need to account for these processes. In the next section we use large ensembles of fully coupled ocean–atmosphere GCMs, which enable the forced climate responses to be analyzed against the background of the ocean-driven variability and response time scale.
3. Transient GCM experiments

The results of the slab model experiments have provided a useful conceptual platform to help us understand the effects of sulfate and CO₂ forcing on precipitation in the Sahel. However, the slab model uses a rather simple representation of the ocean and gives no information on how the time-evolving climate drivers impact the transient climate response. In this section we make use of a fully coupled ocean—atmosphere GCM to explore some of these transient effects. Utilizing the very large third climate configuration of the Met Office
Unified Model (HadCM3L) ensemble generated within climateteprediction.net (see Frame et al. 2009) enables us to look for a signal of the sulfate aerosol and CO₂ forcings on Sahel precipitation against the background of large model variability (which refers to imposed natural forcing, model internal variability, and differences in model parameter settings).

a. Method

Several thousand simulations of the HadCM3L coupled ocean–atmosphere GCM were completed by volunteers on home computers (Frame et al. 2009). Each of the distributed models had a different combination of model parameters (Stainforth et al. 2005), forcing fields [well-mixed greenhouse gases, anthropogenic sulfate, and solar and volcanic aerosol; see Frame et al. (2009), but all with zero DMS emissions], and initial conditions. There were two ensemble experiments run:

1) standard transient runs [STR, 1920–2000 greenhouse gas concentrations, sulfur emissions, and natural forcings from Nakicenovic et al. (2000)], where the simulations have SO₂ emissions specified according to observations and a total of 1566 simulations are included, which have been selected such that they include the same ranges of model parameter values and forcing fields (solar or volcanic) as in the AATR ensemble (described below), and

2) altered aerosol transient runs [AATR, 1920–2000 greenhouse concentrations and natural forcings with IPCC SRES A1B scenario 1990–2070 sulfur emissions, from Nakicenovic et al. (2000)], where the simulations have SO₂ emissions appropriate for 70 yr ahead of the actual model time (see Fig. 6a) and extend on into the SRES A1B scenario for the future; these runs contain 519 models, which simulated 1920–2000 greenhouse gas and natural forcings with future SO₂ emissions.

In this section, the focus will be on July–September (JAS) precipitation, as this accounts for 70% of the annual total in the Sahel and has changed significantly in the period of interest (Lau et al. 2006). For comparisons with observations, we use the CRU’s 0.5° × 0.5° precipitation dataset (New et al. 2000). We discard results prior to 1940 as there is some evidence that the models were still spinning up. Due to the distributed computing nature of this experiment, we were able to include data from runs that did not have a full dataset uploaded from the volunteer. To account for any missing model data, a “climatologies” was calculated for each run, as the parameter combinations led to some simulations being systematically wetter (or drier) than others. From this a precipitation anomaly can then be calculated relative to the climatology of each ensemble member, which can then be combined with available model data to produce an ensemble of anomalies.

From the slab experiment we would expect models with high initial sulfate loadings to show a smaller-magnitude Sahel precipitation response than in those with low initial sulfate loadings.

b. Results

Figure 6a shows the total sulfur emissions in the AATR (black lines) and STR (red lines) runs. The STR ensemble starts with much lower sulfur emissions than in AATR (approximately a factor of 5) and emissions rapidly increased, between 1940 to 1980, by a factor of 5. With AATR, sulfur emissions are offset 70 yr into the future so that sulfur emissions for the same period in AATR correspond to 2010–50, while all other climate...
For comparison with the slab experiments, note that for the 1940–80 period, the AATR ensemble uses SO$_2$ emissions, which lie closer to the FUT$_{SO_2}$ experiment.

The impacts of the different sulfur emissions in STR and AATR on the radiative forcing were estimated (similar to Forster and Taylor 2006) for the standard model configuration of STR and AATR using the following equation:

$$Q = \lambda \Delta T + \Delta TOA$$  \hspace{1cm} (2)

At a particular point in time, changes to the radiative balance of the earth’s climate (represented by $Q$, the radiative forcing, in W m$^{-2}$) are related to the realized global temperature response ($\Delta T$, K) to these changes and the flux imbalance at the top-of-the-atmosphere ($\Delta TOA$, W m$^{-2}$). These radiative changes, $Q$, may arise from changes in atmospheric composition (e.g., aerosols and greenhouse gases) or incoming solar energy. When these changes are first applied, the surface temperature will not have had time to respond and hence $Q$ will equal $\Delta TOA$. As the climate system equilibrates with any radiative forcing, $\Delta TOA$ approaches zero and $\lambda \Delta T$ approaches $Q$. The magnitude of the $\Delta T$ response will be dependent on the blackbody response to the radiative change and the combined effects of the climate feedback processes represented by $\lambda$.

Using diagnosed, modeled $\Delta TOA$ and $\Delta T$, and a $\lambda$ value representative of the standard model configuration ($\lambda = 1.1$ W m$^{-2}$ K$^{-1}$), the temporal estimate of the radiative forcing of the standard transient simulation can be found. Forcing estimates calculated in this way can be seen as indicative of forcing across the ensemble; though some variation in magnitude could be expected [intermodel forcing ($Q$) differences are expected to be smaller than differences in the magnitude of climate responses ($\lambda$) to these forcings]. This approach has become widely used since it was first demonstrated by Forster and Taylor 2006. The radiative forcing implied for both experiments, estimated using this method, is shown Fig. 6b. The estimates are decadal only (a number of the required diagnostics are only available for decadal means). To illustrate how decadal radiative forcing relates to the annual mean forcing [using Eq. (2)], we have also shown decadal and annual values for a higher-resolution (HadCM3) configuration of this model (as used in Collins et al. 2006) forced by the same emissions as STR, plus the inclusion of background DMS. This shows for example how the radiative forcing associated with individual large explosive eruptions (such as the Agung eruption in 1963) can have significant impacts on the decadal estimates (as can also be seen in STR and AATR).

The impacts of the different aerosol loadings on the radiative forcings in the STR and AATR ensembles are clearly evident (Fig. 6). The STR ensemble has a more negative global forcing relative to AATR, which is consistent with the inferred response from the more rapid increase in sulfur emissions from relatively clean background levels. This estimate corroborates the slab experiment findings presented in section 2.

Also, from the findings of the slab experiment, we would expect the interhemispherical forcing difference (NH – SH) to be weaker in AATR than STR, due to the higher initial sulfate loading in AATR. The hemispheric forcing can be estimated using the following two equations:

$$Q_{NH} = \lambda \Delta T_{NH} + \Delta TOA_{NH}.$$ \hspace{1cm} (3)

where $Q_{NH}$, $\Delta T_{NH}$, and $\Delta TOA_{NH}$ are the Northern Hemispheric radiative forcing, surface temperature change, and change in the top-of-the-atmosphere radiative balance, respectively. Similarly for the Southern Hemisphere (SH),

$$Q_{SH} = \lambda \Delta T_{SH} + \Delta TOA_{SH}.$$ \hspace{1cm} (4)

To do this, we need to make a few assumptions. We assume that there is no net flux of energy between hemispheres and that the $\lambda$ global parameter is representative for the two hemispheres. We do not expect the first of these assumptions to hold in practice as energy is likely to flow from the warmer hemisphere, and hence the values derived are likely to underestimate the true hemispheric forcing somewhat. Nevertheless, the contrasting hemispheric forcing estimates derived using this method are likely to be indicative of the sign of the hemispheric difference (even if they may underestimate the magnitude).

The differences in the hemispheric radiative forcing calculated in this way are illustrated in Fig. 6c. The hemispheric difference in forcing is significantly more pronounced in the STR ensemble relative to AATR. This ties in with the results of the slab model experiments (section 2) and the discussion of the sulfur emission differences shown above. The consequences of this hemispheric forcing difference are likely to lead to a weaker warming trend in the Northern Hemisphere relative to the Southern Hemisphere, which is expected to weaken the northern migration of the ITCZ, particularly in STR.

The precipitation trend in an individual model will depend on both the forced response and on the natural internal variability represented within the model. Although we may expect the difference in hemispheric warming to lead to a drying signal in the Sahel (on the
basis of the slab simulations), this signal in any full GCM may be more than offset (or enhanced) by natural variability (predominately driven by ocean processes). Particularly in a region such as the Sahel, which is prone to high coupled internal variability, it is often difficult if not impossible to tease apart the relative impacts of the forced response from a natural mode of variability. The value of the ensemble presented within this section is that we can look at the precipitation response across a large number of models. Looking at the characteristics of the ensemble as a whole enables us to identify the influence of common drivers over the impact of internal variability.

Looking first at the time period during which the historical SO₂ emissions have increased at the most consistent rate (1940–80; see Fig. 6a), we investigate how the range of trends across the model ensembles compares to the observed rainfall changes. The distribution of modeled precipitation changes during this period is illustrated in Fig. 7a. The impacts of modeled process uncertainty and, perhaps more importantly, the internal climate variability are evident in the widths of the two model distributions.

Differences that arise from the aerosol representation within these two ensembles are evident from the difference in the mean ensemble responses. The vast majority, although not all, of the STR ensemble produced a drying signal during this period. In contrast the AATR ensemble produces both drying and moistening responses, but no net signal emerges across the ensemble. The observed trend based on the CRU data (Fig. 7c) is also plotted in this same figure. Although a small fraction of the AATR simulations reproduce the changes, it is the large number of STR simulations that lie close to the CRU trend. Indeed, the trend in the median STR ensemble response lies close to the observed response, suggesting that much of the long-term (1940–80) change in rainfall can be explained by forced anthropogenic aerosol changes rather than a mode of internal variability.

The largest period of rainfall change, however, occurred as a subset of this longer period. The 1950s were moister in the Sahel, and the drying trend to the 1980s although there remains a substantial fraction of the observed drying that is not explained by the forced response within this modeling framework.

There are a number of possible interpretations for the observational trend lying in the tails of the histogram of modeled trends. Perhaps the simplest of which is that the 1950–80 drying trend was a rare event. In other words, a mode (or modes) of natural variability within the climate system conspired to enhance the 1950–80 drying signal. There is well-established existing literature that supports a role for natural processes to drive the Sahel variability. Knight et al. (2006) and Ting et al. (2009), for example, argue that the Sahel rainfall changes respond to the Atlantic multidecadal oscillation (AMO) through changes in Atlantic SSTs. If we assume that the model captures the correct processes, then the interpretation that a rare mode of natural variability also contributed to the drying trend is consistent with the small fraction of modeled trends agreeing with the observed drying (Fig. 7b).

An alternative interpretation, however, is that the model does not capture the full strength of the precipitation response to the underlying climate drivers. It has previously been suggested by Lau et al. (2006) and Biasutti and Giannini (2006) that HadCM3 (the higher ocean resolution version of this model) reproduces only a weak Sahel response to SST anomalies in the Indian and Atlantic Oceans. Indeed, a weak SST response appears to be a general limitation of current climate models driven by observed SST changes (Scaife et al. 2009). The implication of this would be that the model may underestimate the magnitude of the rainfall changes as a response to either natural variability in SSTs or from SST changes resulting from the forced response. If the model is underrepresenting the rainfall impacts from natural variability, then this would suggest that the width of the distributions of trends (in Figs. 7a and 7b) is too small and hence natural variability is more likely to explain the Sahel 1950–80 drying. If the model is underrepresenting the rainfall impacts from anthropogenically forced SSTs, then this might explain why the ensemble underestimates the 1950–80 drying signal. If we accept this interpretation, a similar experiment with a model, which was able to capture stronger coupling, may conclude that the 1950–80 drying trend is much more consistent with the forced response.
able to resolve this issue here, and this remains a question for the future literature.

A further possibility is that the model may fail to fully capture the strength of some important local feedbacks. These could operate on any time scale from days to years, and might include, for example, dust aerosol emissions (Yoshioka et al. (2007)), soil moisture feedbacks, or vegetation feedbacks (Los et al. (2006)). If this is the case, then improved modeling of these processes would also have the effects of broadening the distributions of trends (due to enhanced natural variability) and enhancing the response to anthropogenically forced SST changes.
4. Conclusions

Simulations with a slab-ocean model have shown that NH tropical precipitation is sensitive to both greenhouse gas and sulfate aerosol forcing. Increases in greenhouse gases cause an increase in precipitation and increases in aerosol loading cause a reduction. We suggest that a hemispherical asymmetry in temperature change leads to a shift in the seasonal migration of the ITCZ.

When the sulfate and greenhouse gas forcings are combined, tropical precipitation depends both on the relative magnitude of the imposed forcings and on the location of the ITCZ. For the sorts of combinations of aerosol and greenhouse gas forcings characteristic of the second half of the twentieth century (such as the 1980s), a reduction in precipitation is seen. The model suggests that, historically, aerosol changes dominate. In contrast, when simulating the change between 1980 and 2050, the future Sahel rainfall increases because the future change is influenced more by greenhouse gas concentrations.

The slab model simulations presented here enable us to identify the model response to changes in atmospheric composition within a framework that minimizes atmospheric variability by explicitly neglecting major modes of ocean internal variability (and using multiyear averages). This enables us to identify hemispheric warming differences as a driver of ITCZ position and seasonal rainfall as an important physical mechanism (within this model structure). The slab experiment presented here provides a conceptual framework against which we can understand the modeled Sahel precipitation changes in ensembles of fully coupled ocean–atmosphere GCMs.

In light of the slab model results, it is not surprising that, in a large fully coupled ocean–atmosphere GCM ensemble simulation (STR) of past climate, we see a fall in Sahel precipitation. This reduction may be attributed to hemispheric forcing differences associated with large increases in historic NH, anthropogenic SO$_2$ emissions over the period, ameliorated by the effect of a rise in greenhouse gas emissions. This drying signal is not evident in a second large ensemble (AATR) with identical model structure. The slab experiment presented here provides a conceptual framework against which we can understand the modeled Sahel precipitation changes in ensembles of fully coupled ocean–atmosphere GCMs.

The differences between historical rainfall trends in the two ensembles illustrate the important role that aerosol changes are likely to have played in the Sahel drought.

Previous studies have also suggested that aerosol changes may have played an important role in governing Sahel precipitation, but the small number of models available in these studies and the large variability inherent in this region has made it difficult to link the observed changes either to variability or the forced response. In this paper we have shown that representing the historical aerosol changes within the models drives a consistent drying trend across most of the ensemble. The absence of this signal in the ensemble with the altered sulfur emissions emphasizes the importance of the historical aerosol changes as part of our understanding of the observed drying trend.

The unique perturbed physics, boundary forcing, and initial condition ensemble approach used here allow us to confirm that historical SO$_2$ emissions are likely to explain most of the 1940–80 rainfall changes and a significant proportion of the more pronounced 1950–80 drying. The use of a perturbed-physics ensemble means that our conclusions are not reliant on any single version of the Hadley Centre climate model, but they do, of course, rely on the underlying model structure. If this structure correctly represents the relevant processes in the real world, it is very unlikely that the observed drought was entirely due to natural climate variability.

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