Anthropogenic aerosols dominate forced multidecadal Sahel precipitation change through distinct atmospheric and oceanic drivers.

Haruki Hirasawa\textsuperscript{1*}, Paul J. Kushner\textsuperscript{1}

Michael Sigmond\textsuperscript{2}, John Fyfe\textsuperscript{2}

Clara Deser\textsuperscript{3}

1 Department of Physics, University of Toronto, Toronto, Ontario, Canada.
2 Canadian Centre for Climate Change Modelling and Analysis, Environment and Climate Change Canada, Victoria, British Columbia, Canada.
3 National Center for Atmospheric Research, Boulder, Colorado, USA

Corresponding Author E-mail:
hirasawa@physics.utoronto.ca
Abstract

Sahel precipitation underwent substantial multidecadal-timescale changes during the 20th century which have had severe impacts on the region’s population. Using initial condition Large Ensembles (LE) of coupled general circulation model (GCM) simulations from two institutions forced multidecadal variability, in which Sahel precipitation declines from the 1950s to 1970s then recovers from the 1970s to 2000s, is found. This signal has similar timing, but considerably smaller magnitude than observed Sahel precipitation variability. Isolating the response using single forcing simulations within the LEs reveals that anthropogenic aerosols (AAs) play a leading role in driving this forced variability.

The relative roles of the direct atmospheric response to forcing versus the ocean-mediated atmospheric response are determined using simulations with the atmosphere-land components of the coupled GCMs used in the LEs. Experiments were conducted using boundary conditions from the LE simulations with the direct atmospheric response being isolated using aerosol and precursor emissions, and the ocean-mediated responses being isolated using LE AA-forced sea surface temperatures (SST) and sea ice anomalies. The direct atmospheric response dominates the anthropogenic aerosol forced 1970s-minus-1950s Sahel drying and the ocean-mediated response to SST and sea ice change dominates the 2000s-minus-1970s wetting. While the responses between the models differ, there is qualitative agreement regarding the relative roles of the direct atmospheric and ocean-mediated responses. Since these effects often compete, the model dependence of these effects and their role in the net aerosol-forced response of Sahel precipitation need to be carefully accounted for in future model analysis.
1. Introduction

Emissions of anthropogenic aerosols (AAs) and their chemical precursors have undergone spatially and temporally complex variations that present an intriguing challenge for understanding anthropogenic influence on multidecadal climate variability. One climate variability signal that has been linked to AA forcing is 20th century multidecadal variability of Sahel precipitation, which includes a wet period in the 1950s, drought conditions that peaked in the 1980s and a subsequent moderate recovery towards the present day [e.g. Dai et al., 2004; Rodríguez-Fonseca et al., 2015]. The coincident timing of Sahel precipitation with variations of sulfur dioxide emissions from North America and Europe provides a compelling reason to suspect a causal link between these precipitation changes and AA forcing. Indeed, the drying effect of AA forcing on the Sahel has been well established through analyses of general circulation models (GCMs) [Held et al., 2005; Kawase et al., 2010; Ackerley et al., 2011; Westervelt et al., 2017; Undorf et al., 2018; Hua et al., 2019]. The analysis of Undorf et al. (2018) found a detectable signature of AA forcing on 20th century West African precipitation variability using an ensemble of CMIP5 models. However, this study also found that the models substantially underestimate observed changes and must be scaled up to match observations. Thus, uncertainty remains regarding the extent of anthropogenic influence on Sahel precipitation due in part to the inability of GCMs to correctly model its climatology and variability [Giannini et al., 2008] and to uncertainty in AA forcing and its forced impact on the climate [Myhre et al., 2013].

Sahel precipitation variability is predominantly driven by variability of the West African Monsoon, which is known to be tightly linked to sea surface temperature (SST) variability [e.g. Giannini et al., 2008; Rodríguez-Fonseca et al., 2015]. Arguments connecting AA forcing to
Sahel precipitation typically invoke the influence of AA-forced SST anomalies, particularly the
effect of AA-forced cooling of the Northern Hemisphere (typically in the North Atlantic sector)
relative to the Southern Hemisphere (South Atlantic sector) causing a southward migration of the
Intertropical Convergence Zone (ITCZ) [Ackerley et al., 2011; Mohino et al., 2011; Hwang et al.,
2013; Wang, 2015; Hua et al., 2019]. However, idealized atmosphere-land GCM (AGCM) studies
have shown that the direct atmospheric response to greenhouse gas (GHG) [Biasutti 2013; Dong
et al., 2015; Richardson et al., 2016; Gaetani et al., 2017] and AA forcing [Dong et al., 2014;
Richardson et al., 2016] can also have a substantial influence on Sahel precipitation, even in the
absence of SST changes.

The rapid atmospheric response directly due to external forcing and the slower response
to forced SST changes are often referred to as the “fast” and “slow” responses of the climate
respectively [e.g. Andrews et al., 2009; Biasutti et al., 2013; Li et al., 2018]. However, due to the
use of these terms to describe a range of time scales from weeks to years vs. centuries to
millennia [e.g. Rugenstein et al., 2019], we opt to use more precise terminology. Thus, we refer to
the “direct atmospheric response” to the forcing and the “ocean-mediated response” of the
atmosphere to forced SST and sea ice anomalies, which are analogous to the “fast” and “slow”
responses respectively of Andrews et al. (2009) and others.

The direct atmospheric and ocean-mediated responses have competing roles in the
response of Sahel precipitation to GHG forcing, exacerbating uncertainties in projected future
changes for the region [Biasutti et al., 2013; Gaetani et al., 2017]. These components of the
response are key for understanding historical changes as well. For example, Dong and Sutton
(2015) found that much of the precipitation recovery since the 1980s is due to the direct
atmospheric response to GHGs, while the direct atmospheric responses to AAs from increasing
precursor emissions of sulfur dioxide from North America, Europe, and Asia has been shown to cause drying in the Sahel [Dong et al., 2014; Richardson et al., 2016]. Model studies of other monsoon systems, namely the Asian Monsoon, have found that AA forcing causes direct atmospheric and ocean-mediated responses that can play separate and regionally competing roles in regional precipitation responses [Ganguly et al., 2012; Li et al., 2018]. Notably, many of these analyses study the difference between two given epochs, which means they cannot resolve the transient increase and subsequent decline of AA emissions from regions like North America and Europe, as well as the distinctive roles of atmospheric and oceanic drivers in determining the response to AA forcing over the last 70 years.

In this study, we utilize single forcing Large Ensembles (LEs) from two coupled climate models consisting of large initial condition ensembles of coupled GCM simulations from which the effect of individual forcing agents can be isolated over the historical period. Using the LEs, we are able to isolate the multidecadal AA-forced regional climate changes from internal variability over maritime and terrestrial regions. By comparing these simulations to corresponding all-forcing historical LEs, we can determine the role of AAs in the overall forced climate response. We then conduct AGCM experiments focused on three periods in the latter half of the 20th century. These AGCM experiments use the LEs to obtain AA-forced perturbations to the atmospheric and oceanic boundary conditions. These experiments allow us to investigate the roles of the transient multidecadal direct atmospheric and ocean-mediated responses in AA-forced variability since the 1950s. Section 2 provides details on the Large Ensemble Coupled GCM and AGCM simulations analysed in this study. In section 3, we analyse the LE simulations and the coupled response to AA forcing in the Sahel. In section 4, we analyse the AGCM
simulations and the resultant direct atmospheric and ocean-mediated responses. Concluding remarks are made in section 5.

2. Methods

2.1 Coupled General Circulation Model Large Ensemble Simulations

Analysis of the Sahel precipitation response to AA forcing is first performed using large initial condition ensembles (LE) of coupled ocean-atmosphere GCM simulations from the National Center for Atmospheric Research-Department of Energy Community Earth System Model 1 (NCAR-DOE CESM1) [Kay et al., 2015] and the Canadian Centre for Climate Modeling and Analysis Canadian Earth System Model 2 (CCCma CanESM2) [Kirchmeier-Young et al., 2017; Kushner et al., 2018]. These each include a set of simulations that are configured to isolate the effects of AA forcing. AA levels in both models are set by prescribing emissions of aerosols (i.e. black carbon) and aerosol precursors (i.e. sulfur dioxide) of anthropogenic origin. This may include emissions from biomass burning (BMB), such as forest and grass fires, however they are considered separately in some protocols. The simulations used in this study are summarized in Table 1.

From the CESM1 LE we primarily analyze the 35 All forcing (ALL) and 20 All-but-Aerosol forcing (XAER) simulations which are run at a nominal 1° resolution. In the XAER simulation, the external forcings follow their historical trajectory except for AA and their precursor emissions, which are held fixed to 1920 conditions [Pendergrass et al., 2019]. The impact of AA forcing is determined by taking the difference of the ALL minus the XAER simulations. In order to obtain the ensemble spread of “Aerosol only” (ALL - XAER) simulations for CESM1, individual “AER” ensemble members are calculated as the difference of the ensemble mean of the ALL simulations minus each XAER ensemble member. Notably, BMB
aerosol emissions are not grouped with AA in the XAER simulations. Thus, we also analyze a set of 15 All-but-Biomass Burning aerosol (XBMB) simulations from the CESM1 LE to assess the impact of BMB aerosols. These simulations are analogous to the XAER simulations, but for BMB aerosol emissions. From the CanESM2 LE we analyze 50 All forcing (ALL) simulations and 50 AA-only simulations (AER) which are run at T63 resolution (~1.8°x1.8°). In contrast to the XAER simulations, in the AER simulations all external forcings except for AA are set to pre-industrial conditions, while AAs and their precursor emissions vary along their historical trajectories. Unlike the CESM1 LE, BMB emissions are included as AAs in the CanESM2 LE AER simulations.

Both the ALL and the XAER LEs in CESM1 were initialized using round-off level atmospheric perturbations from a single all-forcing historical simulation at 1920 which was in turn initialized at 1850 from a long 1850 control run [Kay et al., 2015; Pendergrass et al., 2019]. As a result, all the simulations in the CESM1 ALL and XAER ensembles have the same ocean conditions at their initialization in 1920. The ALL and AER CanESM2 LE simulations are initialized by using atmospheric perturbations to branch 10 simulations at 1950 from each simulation in two smaller 5-member ALL and AER forcing ensembles. These original simulations are initialized at 1850 by randomly sampling ocean-atmosphere states from a pre-industrial control simulation [Kirchmeier-Young et al., 2017; Kushner et al., 2018].

If the forced responses to different external forcings were approximately additive, the two simulation protocols for isolating the AA impact on climate would be approximately equivalent. We test potential non-additivity using a small three member ensemble of historical AER simulations in CESM1 (Fig. 3). However, the emission changes differ in the two CESM1 ensembles, as BMB aerosols are included in the small Aerosol-only CESM1 ensemble (Table 1).
Additionally, the three member ensemble was conducted in the CMIP5 CESM1-CAM5 version of CESM1 which differs slightly from the CESM1-LE version and does not include diagnostic ocean-atmosphere biogeochemistry calculations. Precipitation variability in the LEs are evaluated against the Global Precipitation Climatology Centre precipitation analysis [Schneider et al., 2011], the Climate Research Unit Time-Series (CRU TS) [University of East Anglia Climatic Research Unit, 2013], the University of Delaware Terrestrial Precipitation (U. Delaware) [Wilmott and Matsuura, 2001], and the National Oceanic and Atmospheric Administration PRECipitation REConstruction over Land (PREC/L) [Chen et al., 2002].

Both CESM1 and CanESM2 include prognostic aerosol schemes which differ in their representation of aerosol processes such as formation, deposition, and cloud interactions. CESM1 uses the Modal Aerosol Model 3 [Liu et al., 2012] and includes representations of both the aerosol-cloud albedo (Twomey) and aerosol-cloud lifetime effects [Neale et al., 2012]. CanESM2 uses a bulk aerosol model and only represents the aerosol-cloud albedo effect [Von Salzen et al., 2013]. This results in differing aerosol effective radiative forcings (ERFs) with CESM1 at -1.37 Wm\(^{-2}\) and CanESM2 at -0.84 Wm\(^{-2}\) [Zelinka et al., 2014]. It has been found that CESM1 overestimates the enhancement of cloud liquid water path in response to aerosol perturbations from naturally occurring tropospheric volcanic aerosols [Malavelle et al., 2017], which suggests the model may overestimate the negative radiative forcing due to the aerosol-cloud lifetime effect [Toll et al., 2019]. We note in passing that aspects of Sahel precipitation response to radiative changes in these models might be similar. In particular, CESM1-CAM5 and CanESM2 were identified in Monerie et al. (2017) as having similar patterns of Sahel precipitation response to projected GHG forcing. However, as GHG radiative forcing largely affects the longwave and the
spatial patterns of the GHG and AA forcings differs, this finding does not necessarily apply to the historical response to aerosol forcing.

We focus our analysis and additional simulations on three periods during the late 20th century: 1950-1959 (1950s), 1970-1979 (1970s), and 2000-2009 (2000s or near present day). The 1950-1959 period is the initial decade of the CanESM2 LE simulations and represents a time of relatively weak AA forcing and high Sahel precipitation. The 1970-1979 period marks the peak of sulfur dioxide emissions in North America and Europe (NA/EU), and thus strong AA forcing in those regions. Sulfur dioxide emissions in these regions subsequently declined due to air quality regulations [Smith et al., 2011] towards the near present-day 2000-2009 period while emissions from Asia continued to increase. Thus, we study two distinct regimes in the history of sulfur dioxide emissions (and thus sulphate aerosol burdens): 1) the 1970s minus the 1950s (drying period), which is a period of drying in the Sahel region along with strong increases in AA related emissions from NA/EU and moderate increases from Asia; 2) and the 2000s minus the 1970s (recovery period), which is a period of precipitation recovery in the Sahel along with a decline in emissions from NA/EU and continued emissions increases from Asia.

2.2 Atmospheric General Circulation Model Simulation

In order to separate the roles of the direct atmospheric response and the ocean-mediated response in the total coupled response to AA forcing we perform a set of “time slice” AGCM experiments using the atmospheric and land components of the LE models: the Community Atmosphere Model 5 (CAM5) for CESM1 [Neale et al., 2012] and the Canadian Atmosphere Model 4 (CanAM4) for CanESM2 [von Salzen et al., 2013]. Time slice AGCM simulations use boundary conditions that vary seasonally, but do not vary from year to year. The simulation descriptions are summarized in Table 2. Similar simulation protocols have been used in Ganguly
et al. (2012) and Li et al. (2018) to study the Asian Monsoon response to AA forcing. The simulations described here provide two advantages over previous work:

- They provide more information about the multidecadal transient evolution of the precipitation signal by comparing three decadal periods rather than the effect of change between two epochs such as the pre-industrial to present day change.
- They utilize initial condition large ensembles to filter internal climate variability in SST and sea ice. This means that the SST and sea ice perturbations extracted from the coupled models and used to force the AGCMs result in a model-consistent estimate of the forced atmospheric response to AA-forced SST and sea ice anomalies in the coupled GCMs.

The control simulation (EXP 1) is set to seasonally varying observed conditions averaged over 2000-2009 globally, with the SST and sea ice concentration (SIC) climatology calculated using the Hadley Centre Global Sea Ice and Sea Surface Temperature dataset [Rayner et al., 2003]. AA emissions are averaged over 2000-2010 with data pre-2005 using historical emissions and post-2005 using RCP8.5 projected emissions. Other forcing conditions, such as GHG and land use, are set to year 2000 conditions. The other experiments are then perturbed away from this control simulation to give the responses for the 2000s minus the 1950s and the 2000s minus the 1970s. The response for the 1970s minus the 1950s is then obtained by taking the difference of the two sets of experiments.

In the AGCM simulations, the direct atmospheric response is obtained by modifying the aerosol and aerosol precursor emissions to the levels of the target decade. For CAM5, we do not modify BMB aerosol emissions as they are not included as AA in the CESM1 LE XAER simulations. For CanAM4, BMB is included as they are also included in the CanESM2 LE AER
simulations. The ocean-mediated response is obtained by calculating the SST and SIC anomalies due to AA forcing in the LEs for each month of the year for the 2000s minus the target decade then subtracting these anomalies from the observed climatology. The SST and SIC fields are then adjusted in sea ice regions to ensure physical consistency between the two fields, such as reducing SIC in regions where SSTs exceed the melting temperature (Hurrell et al., 2008). Sea ice thickness is not modified in these simulations. By using the anomalies from each set of LE simulations as perturbations in their respective atmosphere components, we obtain a combined direct atmospheric plus ocean-mediated response that is in principle equivalent to the total coupled response of the LE simulations. The ability of the AGCM experiments to replicate the LE response is evaluated in section 4.

The two boundary condition perturbations are applied separately and combined in both models for 2000-2009 minus 1950-1959 giving a total of four 100-year time slice simulations (including the control) for each period and model (Table 2). For 2000-2009 minus 1970-1979, this is the case for CanAM4, but one of the simulations is omitted for CAM5 (EXP 3) owing to computational resource limitations. Where all four simulations were completed for a given period, both the direct atmospheric and ocean-mediated responses can be calculated in two ways. For example, if we consider the direct atmospheric response for the 2000s minus 1970s, it can be calculated as EXP 1 minus EXP 2 in which we change AA emissions with a background of control SST/SIC, as well as EXP 3 minus EXP 4 in which we change AA emissions with a background of SST/SIC perturbed with the AA-forced anomalies. Thus, we obtain the direct atmospheric response with two different SST/SIC background conditions. We find that there is some non-additivity in the precipitation response, though the patterns of response are similar regardless of the background conditions (Fig. S1). For Figs. 6 to 11 both combinations are
averaged together in order to maximize the statistical sampling of the responses. The full list of combinations used is listed in Supplementary Table 1.

2.3 Moisture Budget Analysis

Analysis of the atmospheric moisture budget is conducted following the moisture convergence (MC) method of Li and Ting (2017), wherein a detailed derivation of the following equations can be found. The steady state balance of atmospheric moisture is approximated as

$$\Delta(P - E) \approx \delta MC + \delta ED = \delta TH + \delta DY + \delta ED$$

where precipitation $P$ minus evaporation $E$ anomaly is approximated as the sum of a mean flow ($\delta MC$) and an eddy ($\delta ED$) components of the moisture convergence change, with the mean flow component further divided into the mean flow thermodynamic ($\delta TH$) and dynamic ($\delta DY$) terms. These two terms are calculated as

$$\delta TH + \delta DY = \frac{-1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} u_{k,c} \delta q_k \Delta p_k - \frac{1}{g \rho_w} \nabla \cdot \sum_{k=1}^{K} \delta u_k q_{k,c} \Delta p_k$$

with gravitational acceleration $g$, density of water $\rho_w$, vertical pressure level $k$, total vertical levels $K$, $\Delta p_k$ the pressure thickness of level $k$, $u_k$ the horizontal wind vector at level $k$, and $q_k$ the specific humidity at level $k$. All variables represent the climatological average of monthly mean values, ($)_c$ indicates the time averaged value in the control simulation and the $\delta$ symbol indicates the response calculated as $\delta() = ()_r - ()_c$, where $()_r$ is the time averaged value in the perturbed simulation.

In these equations, responses to different forcings in the long-term mean of the time-slice AGCM simulations are interpreted as anomalies with respect to a control climatology. The mean flow thermodynamic term ($\delta TH$) represents the contribution of anomalous specific humidity ($\delta q$)
being advected by the climatological winds ($u_c$) and the mean flow dynamic term ($\delta DY$) represents the contribution of anomalous winds ($\delta u$) advecting the climatological specific humidity ($q_c$). We calculate $\delta TH$ and $\delta DY$ and assume the residual between these two terms and the $P - E$ is dominated by the $\delta ED$ term, thus neglecting the quadratic term associated with covarying $u$ and $q$ anomalies. That is, products of the response are neglected in this simple budget.

3. Results from Coupled GCM Large Ensembles

July-August-September (JAS) precipitation over the Sahel undergoes significant forced multidecadal changes over the late 20th century in the CESM1 (Fig. 1) and CanESM2 (Fig. 2) LEs. In both models, the ensemble mean of the ALL forcing simulations exhibits a JAS drying across much of Northern Hemisphere Africa in the drying period (1970s minus 1950s) (Fig. 1a, 2a) and a wetting over the Sahel and drying over the Guinea coast in the recovery period (2000s minus 1970s) (Fig. 1c, 2c). Comparison of the ALL forcing ensembles to the AA forcing simulations shows a close correspondence between the response to ALL and AA forcings. This indicates that AA forcing is the predominant cause of the 1970s-minus-1950s forced drying (Fig. 1b, 2b) and a substantial contributor to the 2000s-minus-1970s forced recovery (Fig. 1d, 2d). The ALL forcing response sees larger increases in precipitation compared to the AA forcing response in the recovery period in both models, suggesting that strengthening GHG forcing contributes to the recovery of precipitation as discussed in previous studies [e.g. Dong and Sutton, 2015; Giannini and Kaplan, 2019]. This GHG effect on the recovery is stronger in CanESM2 (Fig. 2c versus 2d) than in CESM1 (Fig. 1c versus 1d), especially in the Sahel. Both models simulate similar patterns of precipitation response to AA forcing over the region, although the local maximum of the CanESM2 response occurs south of the Sahel. Analysis of the signal to noise of
the precipitation change suggests that the minimum ensemble size required to detect a statistically
significant response to AA forcing using the t-test requires as many as 10 to 20 ensemble
members over much of the Sahel for the 1970s minus 1950s (Fig. S2), reinforcing the need for
LEs to identify forced signals on multidecadal time scales.

The transient behaviour of this climate signal is shown in Fig. 3 which displays the 10-
year rolling average of JAS precipitation averaged over the Sahel as defined by the blue boxes in
Figs. 1a, 1b and 2a, 2b. There is close correspondence of the ALL and AER ensemble means in
both CESM1 (Fig. 3a) and CanESM2 (Fig. 3b) with a good degree of agreement between the two
coupled GCMs in the timing and amplitude of the forced variability. CanESM2 has relatively
small intra-ensemble variability compared to CESM1 in this region, which is in part due to the
largest precipitation anomalies occurring south of the Sahel in CanESM2 (Fig. 2). These results
are consistent with the analysis Undorf et al., 2018 who detected an AA-forced signal in West
African Monsoon precipitation in using the CMIP5 multi-model ensemble. The CanESM2 ALL
and AER ensemble means diverge in the 2000s, suggesting an increased wetting effect due to
GHG. Nonetheless, AA forcing has the largest contribution to the recovery in both models, in
contrast to the findings of Dong and Sutton, 2015 who saw a dominant role of the direct
atmospheric response to GHG in the recovery of Sahel precipitation in their AGCM simulations.

In CESM1, the AA forcing response drives a -0.11 mm day$^{-1}$ drying from the 1950s to the 1970s
and a 0.15 mm day$^{-1}$ wetting from the 1970s to the 2000s, which are 102% and 120% of the ALL
forcing response in their respective periods. In CanESM2, the AA forcing response drives a -0.11
mm day$^{-1}$ drying from the 1950s to 1970s and a 0.17 mm day$^{-1}$ wetting from the 1970s to 2000s,
which are 91% and 94% of the ALL-forcing response in their respective periods.
The timing of the forced precipitation signal is similar to the observed multidecadal variability. Calculating the minimum precipitation year between 1950 and 2020 in the 10-year rolling averaged JAS precipitation, we find the average year of peak drought conditions among members of the LEs is 1976 for CESM1 and 1977 for CanESM2 in their respective ALL forcing ensembles with standard deviations of 12 and 10 years respectively. Thus, the observed peak drought timing of ~1986 (with some spread among the observational datasets) is consistent with the internal variability of the LEs. However, a comparison to the observed precipitation records shows that both CESM1 and CanESM2 substantially underestimate the magnitude of the multidecadal variability (Fig. 3), as the precipitation anomalies for all four observation datasets fall outside the range given by the ensembles. The inability of the CESM1 LE to capture this multidecadal variability is also shown by McKinnon and Deser (2018) who found that the CESM1 LE underestimates the variability of 50-year JJA precipitation trends throughout Sub-Saharan Africa relative to an observational LE derived estimate of internal variability. This is a deficiency that is common among CMIP5 models [Biasutti, 2013; Undorf et al., 2018] and complicates quantitative assessment of the relative roles of internal and forced variability in the observed historical record.

In the case of CESM1, we are able to compare the three member ensemble of AER simulations to the ALL minus XAER ensemble mean. The two protocols have similar Sahel precipitation trajectories, suggesting there is reasonable additivity between the response to AA forcing and other external forcings such as GHG for this variable and region. In addition, we find that the regional drying does not appear to be amplified in the ALL simulations (where GHG and AA are both present) relative to the AER simulations. Thus, these models do not show direct
evidence for enhanced drying in response to combined GHG and AA forcing as proposed by

Fig. 4 displays the annual mean sulphate (SO4) burden anomalies from the CESM1 and
CanESM2 LEs. The 1970s minus 1950s period is characterized by widespread increases in
sulphate, especially in the North Hemisphere near and downwind of European and North
American industrial regions (Figs. 4a, b). There is a local increase of sulphate over the Sahel
during this period due in part to emissions being transported from Europe. The sulphate
anomalies are transported further in CanESM2, resulting in greater anomalies in the North
Atlantic, North Pacific, and Arctic relative to CESM1. For the 2000s minus 1970s, there are
decreases in sulphate burdens in much of the North Hemisphere extratropics associated with
decreasing North American and European emissions. Continued increases occur in East Asia,
South Asia, and the Middle East. Locally, there is little change over the Sahel which may be a
combination of decreasing transport from Europe and increasing local emissions and transport
from the Middle East.

Black carbon (BC) burdens increase locally over the Sahel in both periods (Fig. S3), with
much larger increases in CanESM2 as a consequence of the inclusion of BMB emissions. BC
burdens increase over Asia for both periods and decrease over Eastern North America and
Western Europe for both periods. The effect of BMB emissions is evaluated using the CESM1 LE
XBMB simulations. BMB changes result in large BC burden increases over Central Africa in the
drying period (Fig. S4b). However, unlike CanESM2, there is no decline in BC burdens in the
recovery period (Fig. S4d), suggesting some discrepancy in emissions between the models. BMB
aerosol changes result in increased precipitation in the Guinea region in both periods (Fig. S5a,
5c), with a small contribution to the aerosol-forced drying and recovery of Sahel precipitation.
Fig. 5 displays the annual mean AA-forced SST anomalies for the LEs in the drying and recovery periods. These are representative of the boundary conditions used to force the AGCM simulations described in section 2 (see Tables 2 and S1). Both CESM1 and CanESM2 have broadly similar SST anomalies. In the drying period, there is widespread cooling that is strongest in the Northern Hemisphere, apart from the subpolar North Atlantic which shows a “cooling hole” similar to the “warming hole” seen in response to GHG warming [e.g. Menary and Wood, 2018]. In the recovery period, the SST anomalies change sign in the North Hemisphere, with both models showing warming in the North Atlantic and the subpolar North Pacific. However, there is continued cooling in the Indian Ocean and tropical West Pacific, downwind of Asian emission regions that undergo continued AA emission increases. Arctic sea ice cover responses follow the Northern Hemisphere SST response, with an increase in Arctic sea ice cover in the drying period and a decline in the recovery period. However, we do see a small increase in regional ice cover during the recovery period in CESM1 in parts of the Arctic. This appears to be an artifact of subtracting anomalies from the ALL and XAER simulations which both lose sea ice but do so in different regions due to their differing sea ice covers. Changes to Antarctic sea ice cover and Southern Ocean SST are small as AA forcing is weak in this region.

The AA-forced multidecadal SST change in the North Atlantic in the LEs (Fig. 5) results in a reversal of the hemispheric asymmetry of SST anomalies between the two periods. Based on previous arguments (e.g. Ackerley et al., 2011), this reversal of the hemispheric asymmetry of SST anomalies in the Atlantic would appear to be responsible for the reversal of AA-forced Sahel precipitation anomalies. However, a key takeaway from the AGCM experiments discussed below is that the direct atmospheric response to AA forcing plays a substantial and at times competing
role with the ocean-mediated response. In particular, it is the direct atmospheric response rather than the ocean-mediated response that drives the precipitation reduction during the drying period.

4. Direct Atmospheric and ocean-mediated Responses

For our prescribed SST/sea ice AGCM simulations, it is hypothesized that the total AA response (direct atmospheric plus ocean-mediated response) from the AGCM simulations is close to the coupled response to AA forcing from the LE simulations as the atmosphere in both cases experience the same perturbations. However, the match between the AGCM and coupled simulations will be imperfect due to missing nonlinear feedbacks from the lack of ocean-atmosphere coupling in the AGCM experiments, the use of time-slice rather than transient forcing, and the non-additive or state-dependent effects introduced by the construction of the SST/SIC boundary forcing (in which perturbations are applied to the observed SST/SIC climatology rather than the model climatology).

The coupled and AGCM responses are compared for the 1970s minus 1950s in Fig. 6. We find that the CAM5 simulations (Fig. 6c) can reasonably reproduce the JAS African precipitation response from the CESM1 LE (Fig. 6a), while the CanAM4 (Fig. 6d) simulations only capture part of the drying seen in the CanESM2 LE (Fig. 6b). The spatial correlation between the LE and AGCM simulations over global land (African land) is 0.39 (0.37) for CAM5 and 0.35 (0.14) for CanAM4. The agreement between the LE and AGCM improves for the 2000s minus 1970s (Fig. S6), with spatial correlations for global land (African land) being 0.62 (0.76) for CAM5 and 0.50 (0.63) for CanAM4. This may be due to stronger precipitation changes and thus higher signal to noise ratio for this period. The precipitation changes are stronger in CAM5 compared to CESM1 in this recovery period. Overall, while there are errors arising from the idealized nature of the AGCM simulations, the CAM5 simulations can capture the broad characteristics of the response.
However, the CanAM4 simulations can only do so in the recovery period. The subsequent discussion first centres on the CAM5 AGCM simulations, which provide the clearest signals that are most consistent with the coupled model response. We then return to the CanAM4 simulations which provide intriguing points of comparison with CAM5.

The CAM5 experiments reveal that distinct mechanisms drive the precipitation changes during the drying and recovery periods in the Sahel (Fig. 7). For the 1970s minus 1950s the drying seen across much of NH Africa in the total response is dominated by the direct atmospheric response (Fig. 7b), with minimal change over continental Africa due to the ocean-mediated response (Fig. 7c). This pattern of direct atmospheric drying is consistent with other studies of the direct atmospheric response to sulphate aerosol forcing [Dong et al., 2014; Richardson et al., 2016], however the dominance of the direct atmospheric response contrasts with the view that AA-forced Sahel drying is primarily caused by a southward shift of the ITCZ in response to hemispherically asymmetric SST anomalies [e.g. Biasutti and Giannini, 2006; Ackerley et al., 2011; Hua et al., 2019]. One potential explanation for the lack of ocean-mediated precipitation response in this period may be cancelling effects from SST anomalies in different basins. Cool North Atlantic SST are associated with drying in the Sahel while cool Indian ocean, cool Tropical Atlantic, and warm East Tropical Pacific SST are associated with wetting (Fig. 5a) [Giannini et al., 2008, Mohino et al., 2011; Dyer et al., 2017].

For the 2000s minus 1970s, the CAM5 ocean-mediated response (Fig. 7f) causes increased precipitation across much of Africa, particularly in the Sahel and Congo basin with further direct atmospheric drying (Fig. 7e) on the Guinea Coast. In this period, the SST anomalies across the basins tend to be those that are associated with increased precipitation in the Sahel, with warmer North Atlantic SST and cooler Indian Ocean, Tropical Atlantic, and East Tropical
Pacific SST (Fig. 5c). Despite the relatively small change in sulphate burden in the region and the
decline in emissions from North America and Europe, the direct atmospheric response continues
to cause drying in parts of West Africa, though this signal does not extend as far north into the
Sahel as in the earlier drying period with some small precipitation increases occurring in the
Sahara. The drying response may be a remote impact of increasing AA emissions from Asia
similar to that reported by Dong et al. (2014). Thus, what appears in the total response to be a
coherent northward shift of precipitation over Africa extending from the shift in the Tropical
Atlantic ITCZ (Fig. 1c, 7d) actually results from a combination of opposing but spatially offset
direct atmospheric and ocean-mediated responses.

The CAM5 direct atmospheric and ocean-mediated responses of zonally averaged vertical
velocity over Africa align with the precipitation responses in the model (Fig. 8). The direct
atmospheric response causes a weakening of tropical upwelling for both periods (Fig. 8b, e) and
over the Sahel and Sahara north of 15N there are downwelling anomalies during the drying
period which change to upwelling anomalies during the recovery period. This change in the
response aligns with the changing sign of direct atmospheric precipitation response in these
regions. The ocean-mediated response causes upwelling anomalies in the mid troposphere of the
ITCZ for both periods (Fig. 8c, f). In the recovery period, the upwelling anomalies strengthens
and extend to the Sahara north of 20N corresponding to the expansion of the rain band. These
responses are consistent with competing and latitudinally separated direct atmospheric and ocean-
mediated precipitation responses remarked on above. Thus, the strengthening of the ocean-
mediated response is the main driver of the changing total response over the Sahel between the
two periods, with a minor contribution from the changing sign of direct atmospheric response
north of 15N.
Further analysis of the AGCM response is conducted using moisture convergence analysis to decompose the $P-E$ change into mean flow dynamic ($\delta DY$) and thermodynamic ($\delta TH$) components (section 2.3). Over the Guinea Coast and Sahel region south of 15N, the total mean flow component explains most of the change in $P-E$, while the eddy component plays a more pronounced role in the Sahara north of 15N (Fig. S7 for CAM5; Fig. S8 for CanAM4). In order to determine the relative roles of the dynamic and thermodynamic moisture convergence, we calculate the spatial covariance fraction over NH Africa (0 to 20N, 15W to 35E) (Fig. 8) as

$$\frac{cov(\delta MC, P - E)}{var(P - E)}, \frac{cov(\delta DY, P - E)}{var(P - E)}, \frac{cov(\delta TH, P - E)}{var(P - E)}$$

Thus, higher fractional covariance scores indicate greater contributions to the overall change in $P-E$. We find that for CAM5 (Fig. 9a, c), $\delta DY$ explains most of the $P-E$ change for the total, direct atmospheric, and ocean-mediated responses in both periods, with $\delta TH$ accounting for a relatively small portion of the change. These results suggest that AA-forced precipitation changes over continental Africa are largely dominated by changes to circulation in the region, which aligns with similar analysis performed for changes to Asian monsoon circulation [Li et al., 2018].

Although the precipitation response in CanAM4 is noisier and weaker than in CAM5 (Fig. 10), aspects of direct atmospheric and ocean-mediated responses appear to be qualitatively similar. The generally weak signals reflect overall weaker AA effective radiative forcing. For the drying period, there is drying in the Sahel that, while statistically insignificant in the total response (Fig. 10a), is caused by a direct atmospheric drying (Fig. 10b) that has some statistical significance and is partially cancelled by ocean-mediated wetting (Fig. 10c). In the recovery period, there is a statistically significant increase in precipitation over the Sahel in the total response (Fig. 10d) that is dominated by the ocean-mediated response (Fig. 10f), but also has a
contribution from a direct atmospheric wetting response (Fig. 10e). Thus, CanAM4 resembles CAM5 in that drying seen for the 1970s minus 1950s is dominated by the direct atmospheric response, while the wetting seen for the 2000s minus 1970s is dominated by the ocean-mediated response. However, the models differ in that CanAM4 sees ocean-mediated wetting in the drying period and more direct atmospheric wetting in the recovery period.

Like CAM5, CanAM4 shows a direct atmospheric downwelling response (Fig. 11b) and an ocean-mediated upwelling response (Fig. 11c) in the ITCZ in the drying period. In the recovery period, the direct atmospheric vertical velocity response is diminished and reverses sign (Fig. 11e) while the ocean-mediated response continues to cause an upwelling anomaly (Fig. 11f). Thus, the models have qualitative agreement on the sign of the circulation response to the AA-forced direct atmospheric and ocean-mediated responses, except in the case of the recovery period direct atmospheric response. However, the weaker response in CanAM4 results in a lower signal to noise ratio for the response, which may contribute to the discrepancy between the models. Analysis of the P-E response using moisture convergence shows that like CAM5, the dynamic component dominates over the thermodynamic component (Fig. 9b, d). However, we see that the total mean flow moisture convergence has a low fractional covariance with the P-E for some of the responses, with covariance fraction of less than 0.4 for the direct atmospheric and ocean-mediated responses in the drying period. Again, this may be due to the noisy nature of the precipitation response as the covariance tends to be lower for the experiments with weaker responses.

5. Conclusions

By using the single forcing Large Ensembles in CESM1 and CanESM2 to isolate externally forced Sahel precipitation variability from internal variability, we find that it is AA
forcing rather than GHG forcing that dominates the externally forced component of both the
Sahel drought and recovery in the late 20th century in both models. However, while the forced
precipitation variability in the LEs has similar timing to that of the observed Sahel precipitation
variability, the LEs substantially underestimate the magnitude of the variability. The single
forcing LEs prove particularly important for deciphering the influence of AA forcing on regional
climate due to the spatially non-uniform and temporally non-monotonic changes in aerosol and
precursor emissions. As a result, AA can have impacts on multidecadal time scales that might be
difficult to robustly identify with smaller ensembles.

Using AGCM experiments designed to capture the transient multidecadal behaviour of the
coupled GCMs, we find that the direct atmospheric and ocean-mediated responses have distinct
and at times competing roles in driving the multidecadal AA-forced Sahel climate variability. The
CAM5 AGCM simulations reproduce the LE response, while the CanAM4 AGCM simulations
only do so for the recovery period. In the drying period (1970s minus 1950s), AA-forced drying
is largely driven by the direct atmospheric response with either no ocean-mediated response (for
CAM5) or an ocean-mediated wetting response (for CanAM4). This apparently contrasts with the
hypothesis that forced Sahel precipitation variability is predominantly a response to hemispheric
asymmetries in SST anomalies [Held et al., 2005; Ackerley et al., 2011; Mohino et al., 2011;
Wang, 2015]. Although we cannot determine why the CAM5 ocean-mediated component of the
response is so weak, we hypothesize that it may be a result of competing influences from
different ocean basins. In the recovery period (2000s minus 1970s), it is the ocean-mediated
response that drives the increased precipitation with the direct atmospheric response either
continuing to dry (for CAM5) or having moderate wetting effect (for CanAM4). The two models
have qualitatively similar direct atmospheric and ocean-mediated vertical velocity responses in
the two periods, but CAM5 has stronger and more coherent responses. The AA-forced response
of Sahel precipitation therefore cannot be solely understood in terms of SST anomalies feeding
back on the atmosphere. Instead, the direct atmospheric response to the forcing plays a key role
in the forced response. As a result, correlations between multidecadal SST variability and Sahel
precipitation derived from observations and coupled ocean-atmosphere GCM simulations may
not necessarily imply a direct causal link. Rather, they may covary in part due to common
influence from external forcing.

Because anthropogenic aerosol and precursor emissions are expected to decline into the
future [Gidden et al., 2019], we expect further AA-forced increases in Sahel precipitation. Indeed,
analysis of the CESM1 XAER Large Ensemble suggests that AA will continue to have a
dominant role in driving Sahel precipitation increases up to the mid-21st century. However, the
ALL and AER simulations in CanESM2 begin to diverge in the 2000s, indicating that GHG
dominate future changes in this model. Thus, there appears to be model uncertainty regarding the
role of AA forcing into the future. The direct atmospheric response to AA will likely contribute to
the continued recovery of Sahel precipitation as emissions from Asia and Africa begin to decline,
reversing the sign of the response. On the other hand, it is not clear what the role of the ocean-
mediated response will be based on the AGCM simulations in this study, as it is the result of a
combined response to anomalies in different ocean basins.

The contrasting roles of direct atmospheric and ocean-mediated responses for the drying
and recovery periods emphasizes the importance of considering the transient response when
studying the influence of AA on regional climate. Assessing the full pre-industrial to present day
response can give an incomplete picture of the effect of AA on Sahel precipitation by obscuring
aspects of the transient response such as the timing of direct atmospheric and ocean-mediated
influences on regional climate. Analysis of transient AGCM simulations from CMIP6 endorsed
MIPs such as AerChemMIP [Collins et al., 2017] will provide an opportunity to gain further
insight into the role of direct atmospheric and ocean-mediated responses in driving the regional
climate response to AA forcing.

Acknowledgments

We thank Thomas Oudar for his helpful input to the study. This work was supported by
funding from Environment and Climate Change Canada, the National Science and Engineering
Research Council of Canada, and the Ontario Graduate Scholarship Program. Computations were
performed on the Niagara supercomputer at the SciNet HPC Consortium. SciNet is funded by the
Canada Foundation for Innovation; the Government of Ontario; Ontario Research Fund -
Research Excellence; and the University of Toronto.

This material is based in part on work supported by the National Center for Atmospheric
Research, which is a major facility sponsored by the National Science Foundation (NSF) under
Cooperative Agreement 1852977. CESM1 LE data are available from ESG at
http://www.cesm.ucar.edu/projects/communityprojects/LENS/datasets.html. The CESM project is
supported primarily by the NSF.

CanESM2 LE data is available from the Environment Canada and Climate Change
Delaware, PREC/L, and GPCC precipitation data provided by the NOAA/OAR/ESRL PSD,
Boulder, Colorado, USA, from their Web site at https://www.esrl.noaa.gov/psd/.

Acknowledgments

We thank Thomas Oudar for his helpful input to the study. This work was supported by
funding from Environment and Climate Change Canada, the National Science and Engineering
Research Council of Canada, and the Ontario Graduate Scholarship Program. Computations were
performed on the Niagara supercomputer at the SciNet HPC Consortium. SciNet is funded by the
Canada Foundation for Innovation; the Government of Ontario; Ontario Research Fund -
Research Excellence; and the University of Toronto.

This material is based in part on work supported by the National Center for Atmospheric
Research, which is a major facility sponsored by the National Science Foundation (NSF) under
Cooperative Agreement 1852977. CESM1 LE data are available from ESG at
http://www.cesm.ucar.edu/projects/communityprojects/LENS/datasets.html. The CESM project is
supported primarily by the NSF.

CanESM2 LE data is available from the Environment Canada and Climate Change
Delaware, PREC/L, and GPCC precipitation data provided by the NOAA/OAR/ESRL PSD,
Boulder, Colorado, USA, from their Web site at https://www.esrl.noaa.gov/psd/.
References


Schneider, Udo; Becker, Andreas; Finger, Peter; Meyer-Christoffer, Anja; Rudolf, Bruno; Ziese, Markus (2011): GPCC Full Data Reanalysis Version 6.0 at 0.5°: Monthly Land-Surface Precipitation from Rain-Gauges built on GTS-based and Historic Data. DOI: 10.5676/DWD_GPCC/FD_M_V7_050


### Tables

<table>
<thead>
<tr>
<th>Model</th>
<th>Simulation Name</th>
<th>Anthro. Aerosol Emissions</th>
<th>Biomass Burning Emissions</th>
<th>Other External Forcings</th>
<th>N</th>
<th>Years</th>
</tr>
</thead>
<tbody>
<tr>
<td>CESM1</td>
<td>ALL</td>
<td>Historical</td>
<td>Historical</td>
<td>Historical</td>
<td>35</td>
<td>1920-2080</td>
</tr>
<tr>
<td>CESM1</td>
<td>XAER</td>
<td>1920</td>
<td>Historical</td>
<td>Historical</td>
<td>20</td>
<td>1920-2080</td>
</tr>
<tr>
<td>CESM1</td>
<td>XBMB</td>
<td>Historical</td>
<td>1920</td>
<td>Historical</td>
<td>15</td>
<td>1920-2029</td>
</tr>
<tr>
<td>CESM1</td>
<td>AER</td>
<td>Historical</td>
<td>Historical</td>
<td>Pre-industrial</td>
<td>3</td>
<td>1850-2005</td>
</tr>
<tr>
<td>CanESM2</td>
<td>ALL</td>
<td>Historical</td>
<td>Historical</td>
<td>Historical</td>
<td>50</td>
<td>1950-2020</td>
</tr>
<tr>
<td>CanESM2</td>
<td>AER</td>
<td>Historical</td>
<td>Pre-industrial</td>
<td>Pre-industrial</td>
<td>50</td>
<td>1950-2020</td>
</tr>
</tbody>
</table>

**Table 1.** A summary of the coupled ocean-atmosphere GCM simulations used in this study.

### EXP #

<table>
<thead>
<tr>
<th>EXP #</th>
<th>Anthropogenic Aerosol Emissions</th>
<th>SST/Sea Ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2000s</td>
<td>2000s</td>
</tr>
<tr>
<td>2</td>
<td>1970s</td>
<td>2000s</td>
</tr>
<tr>
<td>3*</td>
<td>2000s</td>
<td>2000s+(1970s - 2000s SST/Sea Ice from LENS)</td>
</tr>
<tr>
<td>5</td>
<td>1950s</td>
<td>2000s</td>
</tr>
<tr>
<td>6</td>
<td>2000s</td>
<td>2000s+(1950s - 2000s SST/Sea Ice from LENS)</td>
</tr>
<tr>
<td>7</td>
<td>1950s</td>
<td>2000s+(1950s - 2000s SST/Sea Ice from LENS)</td>
</tr>
</tbody>
</table>

*For CanAM4 only

**Table 2.** Descriptions of the AA and precursor emission and SST/SIC conditions in the AGCM simulations. All simulations were run in both CanAM4 and CAM5 except EXP 3, which was only run in CanAM4.
Figure 1: Ensemble mean JAS precipitation anomalies from the CESM1 Large Ensemble simulations for the 1970s minus 1950s (top row) and 2000s minus 1970s (bottom row) in CESM1 ALL (left column) and ALL-XAER (right column). Stippling surrounded by a thin red contour indicates grid points that are statistically significant at the 95% level. A black dashed contour indicates the 2000-2009 climatological 4 mm per day JAS precipitation region. The blue box indicates the averaging region used in Fig. 3, which includes notches on the western edge as a result of masking out ocean grid boxes.
Figure 2: As in Fig. 1, but for the CanESM2 Large Ensemble’s ALL (left column) and AER (right column) simulations.
Figure 3: 10-year rolling averaged July-August-September (JAS) Precipitation anomalies relative to the 1950-1999 mean regionally averaged over the Sahel (10N to 20N and 20W to 35E) from the (a) CESM1 and (b) CanESM2 Large Ensembles. The ensemble means of the LEs are shown in solid red (All forcing response) and blue (AA forcing response) lines where the AA forcing response is calculated as the ensemble mean of ALL-XAER for CESM1 and ensemble mean of AER CanESM2. Shading indicating the 5-95 ranges of each large ensemble. The ensemble mean of a smaller 3-member ensemble of AA-only simulations in CESM1-CMIP5 (cyan) is plotted in (a). 10-year rolling averaged observed JAS precipitation anomalies from CRU TS (dash), NOAA PREC/L (dash-dot), GPCC (solid), and U. Delaware (dot) are overlaid on the modeled anomalies in black.
**Figure 4:** Annual mean sulphate burden anomalies for 1970s minus 1950s (top row) and 2000s minus 1970s (bottom row) from CESM1 <ALL> - XAER LE (left column) and CanESM2 AER LE (right column).
Figure 5: As in Fig. 4 for the AA-forced SST anomalies. Sea ice concentration anomalies are displayed in purple contours on 2.5% intervals.
Figure 6: Comparison of the AA-forced 1970s minus 1950s JAS precipitation anomalies for the coupled LE CESM1 (a) and CanESM2 (b) compared to the total (direct atmospheric + ocean-mediated) response in their respective atmosphere models: CAM5 (c) and CanAM4 (d). Note that (a) and (b) are the same as Fig. 1b and Fig. 2b respectively. Stippling and thin red contour indicates responses that are significant at the 95% level using a point-wise t-test.
Figure 7: Total (left column), Direct Atmospheric (middle column) and Ocean-Mediated (right column) components of the JAS precipitation response to AA forcing in CAM5 for the 1970s minus 1950s (top row) and 2000s minus 1970s (bottom row). Stippling and thin red contour indicates responses that are significant at the 95% level using a point-wise t-test.
Figure 8: As in Fig. 7 for JAS vertical velocity ($\omega$) zonally averaged over Africa (15W to 35E).

Black contours indicate the climatological vertical velocity on intervals of 0.1 Pa s$^{-1}$. 
Figure 9: Covariance fraction discussed in the text, comparing the P-E anomaly field to the
Mean Flow Total (δMC), Dynamic (δDY), and Thermodynamic (δTH) Moisture Convergence
anomalies over NH Africa (0N to 20N, 20W to 35E) for the 1970s minus 1950s (top row) and
2000s minus 1970s (bottom row) in CAM5 (left column) and CanAM4 (right column). For each
case, the Total (TOT/green), Direct Atmospheric (ATM/orange), and Ocean-Mediated
(OCN/blue) response bar plots are shown with the total (δMC), dynamic (δDY), and
thermodynamic (δTH) mean flow moisture convergence from left to right.
Figure 10: As in Fig. 7 for CanAM4 precipitation. Note that the color scale is halved relative to Fig. 7.

Figure 11: As in Fig. 8 for the CanAM4 response. Note that the color scale is halved relative to the CAM5 response in Fig. 8.