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Key Points:

- In our simulations, geoengineering effectively maintains global mean soil moisture under a high CO₂ scenario, but some regional changes remain
- Summertime soil moisture is reduced in India and the Amazon under geoengineering; this change is dominated by the change in precipitation
- Geoengineering using multiple injection locations yields smaller magnitude of changes in soil moisture than equatorial-only injection

Supporting Information:

- Supporting Information S1
- Table S1
- Table S2
- Table S3
- Table S4
- Table S5

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Soil Moisture and Other Hydrological Changes in a Stratospheric Aerosol Geoengineering Large Ensemble

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Abstract Stratospheric sulfate aerosol geoengineering has been proposed as a potential strategy to reduce the impacts of climate change. Here we investigate the impact of stratospheric aerosol geoengineering on the terrestrial hydrological cycle. We use the Geoengineering Large Ensemble, which involves a 20-member ensemble of simulations using the Community Earth System Model with the Whole Atmosphere Community Climate Model, in which sulfur dioxide (SO₂) was injected into the stratosphere at four different locations, to maintain global mean surface temperature, and also the interhemispheric and equator-to-pole temperature gradients at values representative of 2020 (“baseline”) under the Representative Concentration Pathway 8.5. In our simulations, annual mean land precipitation and evapotranspiration (ET) increase by 12% each under Representative Concentration Pathway 8.5. Under the Geoengineering Large Ensemble, the hydrological cycle is suppressed compared to the baseline, with end-of-century decreases of 1.4% (12 ± 5 mm/year) and 3.3% (18 ± 2 mm/year) in global mean, annual mean precipitation, and ET over land, respectively. Geoengineering effectively maintains global mean soil moisture under a high CO₂ scenario, although there is significant regional variability. Summertime soil moisture is reduced by 42 ± 11 kg/m² (3.5%) and 27 ± 16 kg/m² (2.1%) in India and the Amazon, respectively, which is dominated by the decrease in precipitation. We also compare these regional changes in soil moisture under the Geoengineering Large Ensemble with an equatorial-only SO₂ injection case and find a similar sign in residual changes, although the magnitude of the changes is larger in the equatorial run.

1. Introduction

Stratospheric sulfate aerosol geoengineering has been suggested as a possible option to counteract some of the impacts of climate change caused by rising greenhouse gas (GHG) concentrations (e.g., Crutzen, 2006; MacMartin et al., 2018; National Research Council, 2015). This idea takes its inspiration from the stratospheric aerosol layer created from large volcanic eruptions, which results in a cooling of Earth's surface lasting several years (Robock, 2000). Intentional continuous sulfur dioxide (SO₂) injections into the stratosphere would result in an increase in stratospheric sulfate aerosols, which would scatter some of the incoming solar radiation back to space.

Global warming leads to an intensified hydrological cycle due to the increases in lower tropospheric water vapor and net downward radiative flux (Held & Soden, 2006). One of the concerns identified with using stratospheric aerosol geoengineering to completely offset this positive forcing from GHGs is a reduction in the hydrological cycle. A robust conclusion that was first exposed through solar reduction modeling is that geoengineering overcompensates global mean precipitation relative to global mean temperature (Bala et al., 2008; Kleidon et al., 2015; Kravitz, Rasch, et al., 2013), with potentially significant changes in regional precipitation patterns and weakening of monsoonal precipitation (Dagon & Schrag, 2017; Niemeier et al., 2013; Robock et al., 2008; Tilmes et al., 2013). Similar conclusions hold for stratospheric aerosol geoengineering (Xia et al., 2017), with potentially even larger regional decreases in precipitation due to the warming of the stratosphere (Niemeier et al., 2013).

However, these past modeling studies have mostly examined only equatorial or tropical injection scenarios, as in the Geoengineering Model Intercomparison Project scenarios G3 and G4 (Kravitz et al., 2011). For

some desired climate objectives, the climate response to geoengineering can be at least partially tailored by choosing the amount of injection, the latitude of injection, altitude of injection, and the time of year of injection (Dai et al., 2018; MacMartin et al., 2017; Tilmes et al., 2017; Tilmes, Richter, Mills, et al., 2018). Since the spatial pattern of aerosol optical depth can be at least partially controlled, stratospheric aerosol geoengineering using multiple injection locations can improve how well geoengineering compensates for the spatial pattern of climate changes due to increased atmospheric CO₂ concentrations (MacMartin et al., 2017).

This method has been demonstrated by Kravitz et al. (2017). The objectives chosen for this study were to maintain 2020 levels not only for the global mean surface temperature (T_0) but also the interhemispheric (south-to-north) temperature gradient (T_1), and the equator-to-pole temperature gradient (T_2). Based on departures of the simulation from chosen objectives from the previous years, the amount of stratospheric sulfur injection at each of four latitudes, 30°N, 30°S, 15°N, and 15°S, was annually adjusted using a feedback-control algorithm, which effectively manages temperature objectives despite uncertainty and non-linearities. Simulations were performed using the Community Earth System Model, version 1, with the Whole Atmosphere Community Climate Model as its atmospheric component, CESM1(WACCM): a coupled atmosphere-ocean general circulation model with fully interactive stratospheric chemistry, dynamics, and a sophisticated treatment of sulfate aerosol formation, microphysical growth, and deposition (Mills et al., 2017). Using a nearly-identical setup to Kravitz et al. (2017), the 20-member stratospheric aerosol Geoengineering Large Ensemble (GLENS) of simulations has been conducted (Tilmes, Richter, Kravitz, et al., 2018). This allows for a more robust assessment of the resulting regional climate changes in the presence of natural variability, which is particularly important for many hydrological fields with high interannual variability and thus lower signal-to-noise ratios (MacMartin et al., 2019).

Maintaining temperature gradients can reduce some side effects that result from only offsetting global mean temperature change. Controlling the interhemispheric temperature gradient is expected to reduce shifts in the latitude of the Intertropical Convergence Zone (ITCZ) and associated tropical precipitation that could occur if one hemisphere is cooled more than the other (Haywood et al., 2013; Jones et al., 2017), while controlling the equator-to-pole gradient is motivated by the overcooling of the tropics relative to the poles that is seen in many previous studies (e.g., Bala & Caldeira, 2000; Kravitz et al., 2016; Kravitz, Caldeira, et al., 2013). Managing three degrees of freedom of temperature response, rather than just one, should improve compensation of the changes due to increases in GHG concentrations. However, there are still some potentially significant regional changes in surface temperature under such a strategy (Kravitz et al., 2017; Tilmes, Richter, Kravitz, et al., 2018). While the objectives were specified in terms of temperature, this strategy might be expected to also result in reduced changes in regional precipitation patterns relative to equatorial injection.

As noted earlier, previous simulations show that geoengineering sufficient to completely offset the positive forcing from increasing GHG concentrations would generally weaken precipitation on land, particularly in monsoon regions (Tilmes et al., 2013). However, these decreases in precipitation could also be accompanied by a decline in evaporation that could result in net increases in runoff and soil moisture in many regions even in the presence of reduced precipitation (Glienke et al., 2015; Tilmes, Richter, Kravitz, et al., 2018). Vegetation also plays an important role in the hydrological cycle because of the physiological effect of CO₂ on vegetation and vegetation-climate feedback that reduce evapotranspiration (ET) with increasing CO₂ (Dagon & Schrag, 2016; Glienke et al., 2015; Tilmes et al., 2013), although there is not always a reduction in ET in some water-limited regions due to increased vegetation productivity (Ukkola et al., 2016). Plant water use is influenced by elevated CO₂ concentrations, leading to lower stomatal conductance and increased leaf photosynthetic rates (Lammertsma et al., 2011). On the other hand, increased scattering with aerosol geoengineering could also increase transpiration due to enhanced photosynthesis, somewhat counteracting the reduction in ET from elevated CO₂ (Xia et al., 2017).

Rather than only considering precipitation or ET , changes in soil moisture have the potential to influence weather and are more directly related to climate change impacts on agriculture, ecosystems, drought risk, and forest fires. We focus on soil moisture in the top 10 cm, top 0.8 m, and top 3.8 m of soil. Soil moisture in the top 10 cm is more relevant for changes in the frequency and intensity of heat waves (Dagon & Schrag, 2017; Hirschi et al., 2011). For vegetation, deeper soil moisture matters, with the relevant depth depending on the plant. Soil moisture in the top 0.8 m is more relevant for some agricultural plants. Hydrologically active soil layers span a depth of 3.8 m in our simulations, although some trees extend far

below this depth. Soil moisture memory, the integrative nature of soil moisture in storing water, has the potential to affect and amplify droughts and floods (Koster & Suarez, 2001; McColl et al., 2017). The simulated soil moisture memory decreases from the bottom to top of the soil due to the influence of strong precipitation and evaporation forcing from above (Lawrence et al., 2007). Moreover, soil moisture constrains plant transpiration and photosynthesis in several regions of the world, with consequent impacts on water, energy, and biogeochemical cycles (Seneviratne et al., 2010). Soil moisture is the main variable used to define and identify agricultural drought, which is a major cause of limited agricultural productivity and of crop yield uncertainty throughout the world (Martínez-Fernández et al., 2016). The occurrence of forest fires depends on fuel mass and structure, the presence of an ignition source, wind speed, and fuel moisture (Bartsch et al., 2009). Wet surface soil moisture conditions limit the extent of burned area and prevent the outbreak of fires due to the delayed volatilization of the fuel and the increased heat required for ignition (Nelson, 2001).

Regionally averaged soil moisture integrates changes in the water cycle and vegetation. When soil moisture is close to saturation, small variations in precipitation may result in large variations in runoff. Changes in the frequency of precipitation extremes could affect the storage of soil moisture and the formulation of runoff. On the other hand, in dry soil moisture regimes, *ET* is very sensitive to variations in soil moisture. Changes in soil moisture would alter numerous processes and feedback loops in the climate system. Higher precipitation leads to higher soil moisture, which leads to higher *ET*. The higher *ET* then leads to higher precipitation, amplifying the feedback (Seneviratne et al., 2010). However, there is also a potential negative feedback: increasing *ET* decreases the available soil moisture. When available water is decreasing, plants can preserve soil water by shutting their stomata (Hetherington & Woodward, 2003). Whenever soil moisture limits the total energy transferred by latent heat flux, more energy is available for sensible heating, inducing an increase in near-surface air temperature. Increased temperature leads to a higher vapor pressure deficit and evaporative demand, possibly leading to a further decrease in soil moisture.

Here we focus on soil moisture changes and the associated driving factors under stratospheric sulfate aerosol geoengineering, using the GLENS 20-member ensemble of simulations with a state-of-the-art climate model to estimate the forced response in the presence of variability, and using a multiple-latitude injection strategy deliberately designed to reduce residual temperature changes (Tilmes, Richter, Kravitz, et al., 2018). In addition, we also include results from three ensemble members where equatorial injection is used to maintain only the global mean temperature and not temperature gradients (Kravitz et al., 2019). This comparison can help assess how managing the additional temperature degrees of freedom affects the response of the hydrological cycle.

The rest of the paper is structured as follows. In section 2, we describe the Community Earth System Model and simulations. In section 3, we describe the results of the simulations, including global features, soil moisture regional changes, and seasonal changes. In section 4, we discuss regional changes in climate and the hydrological cycle, the differences between only equatorial and multiple location injections, and mechanisms of hydrological responses. In section 5, we summarize our study and provide conclusions.

2. Model and Simulations

Simulations were conducted with CESM1(WACCM), with horizontal resolution of 0.9° in latitude by 1.25° longitude, and 70 vertical layers in the atmosphere up to 140 km. The model is fully coupled to land, ocean, and sea ice components, with fully interactive stratospheric chemistry, dynamics (including an internally generated quasi-biennial oscillation), and a sophisticated treatment of sulfate aerosol formation, microphysical growth, and deposition. Model description and validation are given by Mills et al. (2017), with the exception that the version used here has a new version of the land model CLM4.5, as described by Tilmes, Richter, Kravitz, et al. (2018). Relative to the earlier version CLM4, CLM4.5 includes some modifications of hydraulic properties of frozen soils and shows some seasonal changes in surface temperature over the Arctic (Mills et al., 2017; Oleson et al., 2013; Swenson et al., 2012).

We describe the CLM4.5 land model in more detail, particularly as it relates to modeling of soil moisture. The land model with fixed distributions of vegetation includes interactive carbon and nitrogen cycles, strongly limiting the CO_2 fertilization effect compared to the models without nitrogen cycle (Glienke et al., 2015). There is a seasonal cycle of leaf area index, which is calculated prognostically. CLM4.5 allows leaf area index to respond to climate changes. The stomatal conductance is directly influenced by the relative

humidity at the leaf surface, the CO₂ concentration at the leaf surface, and the soil water. Vegetation responds to changes in diffuse radiation, but enhanced UV and ozone-mediated impacts on vegetation are not included. The crop model is not turned on.

In CLM4.5, the liquid precipitation reaching the groundwater and meltwater from snow is then partitioned between infiltration into the soil, surface runoff, and surface water reservoirs. The parameterization of *ET* accounts for soil resistance, litter resistance, and canopy stability which leads to a better agreement with observations (Barbour et al., 2005). Surface runoff consists of overland flow due to saturation excess and infiltration excess mechanisms, which is determined by the topographic characteristics and soil moisture state of a grid cell (Niu et al., 2005). Subsurface drainage is lateral flow calculated based on the SIMTOP scheme (Niu et al., 2005) with a modification to account for reduced drainage in frozen soils. Glacier runoff is calculated from the residual of the water balance, but CLM4.5 has a somewhat unrealistic treatment of accumulation and melting for standard glacier land units (e.g., when glacier ice melts, the meltwater is assumed to remain in place until it refreezes in CLM; Oleson et al., 2013).

Soil moisture is simulated based on the 10 hydrologically active layers spanning a depth of 3.8 m, in which the vertical soil moisture transport is governed by infiltration, surface and subsurface runoff, gradient diffusion, gravity, canopy transpiration through root extraction, and interactions with groundwater (Oleson et al., 2013). Soil water from each grid cell to its downstream neighboring grid cell is routed by River Transport Model using a linear transport scheme (Branstetter, 2001). CLM4.5 includes a bulk aquifer model to simulate saturated water storage dynamics below the model soil column. Aquifer storage increases when it receives recharge from the overlying soil column, and decreases due to lateral flow (i.e., subsurface drainage) and capillary rise (Lawrence et al., 2011). The water table is simulated based on work by Niu et al. (2007). However, CLM4.5 aquifer model simulates unrealistic long-period behavior relative to total water storage observations, especially in the large wetting events, due to the lack of a finite lower boundary in the bulk aquifer model (Swenson & Lawrence, 2015), and thus, modeled aquifer storage is not necessarily meaningful. In at least a few places, the simulated soil moisture has been shown to explain much of the observational total water storage variability (Swenson & Lawrence, 2015). CLM4.5 exhibits biases in soil evaporation when the canopy is sparse or absent, which reduces moisture inputs into the ground in semiarid regions (Lawrence et al., 2011; Swenson & Lawrence, 2014).

All simulations use prescribed GHG concentrations following the Representative Concentration Pathway 8.5 (RCP8.5) (Meinshausen et al., 2011). A scenario with more mitigation, such as RCP4.5, would not require as much cooling to maintain the same temperature. Choosing to simulate with RCP8.5 forcing provides a high signal-to-noise ratio for assessing changes in the climate system response. A 20-member ensemble of RCP8.5 simulations was conducted from 2010 to 2030, with three ensemble members extended until 2097. The GLENS simulations were branched from the RCP8.5 simulations in 2020 and continued until 2099. In subsequent analyses we use the average over 2010–2030 of the RCP8.5 simulations as the baseline for comparisons, and consider the change averaged over 2075–2095 relative to this baseline. Differences are quoted below together with a $\pm$ range that indicates the standard error in estimating the change. The simulations are described in detail in Tilmes, Richter, Kravitz, et al. (2018).

Global mean temperature is maintained at the target value (with standard deviation across ensemble members of ± 0.1 °C) throughout the simulation, resulting in 4.05 ± 0.06 °C cooling averaged over the 2075–2095 period, and requiring 42 Tg SO₂/year injection (Tilmes, Richter, Kravitz, et al., 2018). There is some small residual bias in *T*₂ (high latitudes are warmer than the equator). The objectives are defined in terms of annual mean temperature, but there are potentially important seasonal effects. In particular overcooling during summer and undercooling during winter at high latitudes, likely a result of both more sunlight in the summer to reflect, and dynamical effects from stratospheric heating (Driscoll et al., 2012).

To understand the changes in soil moisture, some useful insight can be obtained by considering a water balance approach (Huntington & Billmire, 2013; Oleson et al., 2013) where the difference between precipitation over land (*P*), evapotranspiration (*ET*), and runoff (*R*) can be described by the following basic conservation equation:

$$P - ET - R = \frac{ds}{dt}$$

Figure 1. Annual, land mean time evolution of model output for the RCP8.5 (red), GLENS (blue), and equatorial injection (green) simulations. Single ensemble members are shown in lighter colors, the ensemble mean in darker colors. The top 3.8 m of soil is the hydrologically active part of the soil column (Oleson et al., 2013). Total runoff includes total surface runoff and subsurface drainage.

where $\frac{ds}{dt}$ is the rate of change in water storage, which includes soil moisture, surface water reservoirs (e.g., canopy intercepted water, surface water, and snow water), and unconfined aquifer water. Soil moisture changes could be affected by not only changes in precipitation, ET , and runoff but also changes in the distribution among surface water reservoirs. In turn, changes in soil moisture will affect the ground evaporation, aquifer recharge, runoff, and consequently precipitation, until the system reaches a new steady state. The GLENS simulations are never in equilibrium as GHG concentrations are continuing to rise, but changes in $P - ET - R$ are still small (Figure S1a) relative to the changes in precipitation, ET , and runoff, individually. In the quasi-equilibrium state, ET (for example) is not only responding to changes in net surface shortwave flux or changes in CO_2 concentration, but it is also responding to changes in soil moisture and surface water reservoirs, making it difficult to sort out cause and effect.

When looking at regional changes, we focus on a number of specific regions, indicated in Figure 3 and listed in Tables S1–S4. There are the same areas chosen by Dagon and Schrag (2016), with four additional regions chosen because of interesting features in the hydrological responses.

3. Simulation Results

Here we describe some global features of the hydrological response from the GLENS geoengineering simulations, focusing in particular on soil moisture regional changes (section 3.2), drivers of soil moisture changes (section 3.3), and seasonal changes (section 3.4). In Figures 1 and S1, we show the annual mean time evolution of various model outputs averaged over land. In Figures 2–7 we show two comparisons: the climate changes under RCP8.5 (three-member ensemble mean of RCP8.5 simulations during 2075–2095) relative to the baseline (20-member ensemble mean of RCP8.5 simulations during 2010–2030), and the climate

Figure 2. Maps showing simulated (a and b) annual mean, (c and d) JJA average, and (e and f) DJF average changes in surface air temperature (°C) for the (left column) RCP8.5 ensemble mean (2075–2095) and the (right column) GLENS ensemble mean (2075–2095) compared with RCP8.5 ensemble mean during 2010–2030 (baseline), respectively. Hatched areas are regions where changes are not statistically significant at the 5% level using the Student's *t* test.

changes under GLENS (20-member ensemble mean of geoengineering simulations during 2075–2095) relative to the baseline, with our primary focus being the latter comparison.

3.1. Global Features

Relative to the baseline (RCP8.5 simulations during 2010–2030), the RCP8.5 simulations during 2075–2095 exhibited a 5.4 ± 0.1 °C warming averaged over land, where the indicated range represents the standard error (after averaging over 21 years, each available ensemble member is treated as an independent sample; Figure 1a). Under RCP8.5, the water cycle intensifies and both *ET* and precipitation increase as a result of

Figure 3. Same as in Figure 2 but for soil moisture (liquid and ice) in the top 3.8 m (kg m^{-2}). The boxed regions in (a) show 10 regions of interest: the Congo, the Amazon, India, Western Russia, Western Europe, the U.S. Central Plains, the U.S. Southwest, Australia, northwest China, and North Africa.

increased atmospheric water vapor content and vertical motion associated with temperature-dependent climate feedback; precipitation and ET are also affected by the “fast” response to all forcings, including CO_2 , in which the atmosphere regains its energy balance (Andrews et al., 2010; Cao et al., 2015; Chou et al., 2011). Under RCP8.5, annual mean ET and precipitation averaged over land both increase by roughly 12% (see Table 1 for details and Figures 1b, 1c, and S2). The responses of runoff to increases in precipitation are thus counteracted by the increasing ET (Figure S1 b), although changes in glaciers may contribute a significant portion of changes in runoff (Huntington & Billmire, 2013). Global annual mean total runoff (including total surface runoff and subsurface drainage) increases by 9.5% almost completely contributed by subsurface drainage, owing to the increase in global annual mean precipitation and

Figure 4. Annual mean percentage change in soil moisture (liquid and ice) in the top 3.8 m under RCP8.5 ensemble mean (2075–2095) and GLENS ensemble mean (2075–2095) relative to baseline ensemble mean (RCP8.5 runs during 2010–2030), respectively. Error bars represent standard error. The soil moisture (liquid and ice) in the top 3.8 m (kg m^{-2}) of baseline is listed under each region name. The text under each bar gives the label of each region: CON (the Congo), AMA (the Amazon), IND (India), W_RUS (Western Russia), W_EUR (Western Europe), US_C (the U.S. Central Plains), US_S (the U.S. Southwest), AUS (Australia), N_CHI (northwest China), and N_AFR (North Africa).

permafrost thaw under RCP8.5 (Lawrence et al., 2015; Figures 1b, 1d, S1c, S1d, and S3). Annual mean surface water reservoirs (including canopy intercepted water, surface water, and snow water) decrease by 28.1%, which is dominated by snow melting and the increases in ET (Figures 1c, 1e, S1e, and S1f). Soil moisture (liquid water and ice) in the top 10 cm, 0.8 m, and 3.8 m (the hydrologically active part of the soil column; see Oleson et al., 2013) decreases by 7.2%, 4.0%, and 0.9%, respectively, with the change ratio in the upper layers much more than that of the entire modeled soil column (Figures 1f, S1g, and S1h). Under RCP8.5 there is a significant reduction in soil ice in permafrost regions without as large an increase in soil liquid content (Figures 1g, 1h, and S3). This is responsible for the overall decline in land-averaged soil moisture, despite the increase in precipitation minus ET ($P - ET$) under RCP8.5 (Lawrence et al., 2015; Figures 1f, S1g, and S1h). Under the newer version of the land model, CLM5, the decreased soil ice in permafrost regions under climate change is compensated by an increase in soil liquid water (Lawrence et al., 2019), but it is not known which behavior is more realistic.

Geoenvironment as simulated in GLENS successfully maintains broad features of the annual mean surface temperature over land at the goal of the control period (2010–2030); however, there are still residual regional changes relative to the baseline period (Figures 1a and 2). Relative to baseline, the surface net shortwave radiation averaged over land decreases by $5.4 \pm 0.1 \text{ W m}^{-2}$ (3.9%) in GLENS due to the stratospheric sulfate aerosol layer; however, the surface net longwave radiation increases by $2.6 \pm 0.1 \text{ W m}^{-2}$ (4.1%; positive values indicating increases in the downward direction) owing to the increasing GHG concentrations and adjustments of clouds (Figures S1i and S1j). Annual mean ET averaged over land decreases by $18 \pm 2 \text{ mm/year}$ (3.3%) under geoenvironment in GLENS during 2075–2095 compared with baseline (Figure 1c). Ground evaporation and transpiration are the main contributors to the total ET changes; however, canopy evaporation increases by $7 \pm 1 \text{ mm/year}$ (7.4%), which could be related to the increase in canopy intercepted water (Figures S1e and S1k–S1m). Annual mean precipitation averaged over land decreases by $12 \pm 5 \text{ mm/year}$ (1.4%), which is consistent with the decrease in surface total (shortwave plus longwave) radiative flux (Figures 1b, S1i, and S1j), because mean precipitation is governed by the availability of energy. Total runoff is almost unchanged, because the decrease in ET counteracts the precipitation change averaged over land (Figures 1d and S1b). However, total surface runoff decreases by $4.1 \pm 1.3 \text{ mm/year}$ (2.3%) and subsurface drainage increases by $7.3 \pm 3.1 \text{ mm/year}$ (3.9%; Figures S1c and S1d). Annual mean surface water

Figure 5. Same as in Figure 2 but for precipitation (mm/day).

reservoirs decrease by $10.2 \pm 0.4 \text{ kg/m}^2$ (7.5%) due to loss of snow under residual warming at high latitudes during winter (Figures 1e, 2, and S1f). There are small increases of $0.22 \pm 0.05 \text{ kg/m}^2$ (0.8%), $0.9 \pm 0.3 \text{ kg/m}^2$ (0.4%), and $1.3 \pm 0.9 \text{ kg/m}^2$ (0.1%) for soil moisture in the top 10 cm, 0.8 m, and 3.8 m, respectively (Figures 1f, S1g, and S1h). RCP8.5 results in significant decrease in permafrost (soil ice) that is roughly maintained under GLENS (Figure 1h).

Geoengineering as simulated in GLENS approximately maintains the global mean soil moisture at baseline values, although there are some potentially important residuals (Figure 3). We calculated an area-weighted root-mean-square error metric via

Figure 6. Same as in Figure 2 but for ET (mm/day).

$$\text{RMSE}_A = \sqrt{\frac{\sum (S - S_{\text{base}})^2 dA}{\sum dA}}$$

where S is the soil moisture (liquid and ice) in any grid cell averaged over 2075–2095 (ensemble average), S_{base} is the soil moisture (liquid and ice) in the baseline (2010–2030; ensemble average) in that grid cell, dA is the area of that grid cell, and summation is over all grid cells. RMSE_A for RCP8.5 is 4.0, 23.2, and 81.1 kg/m² for soil moisture in the top 10 cm, 0.8 m, and 3.8 m, respectively. For GLENS, the values are 2.7, 15.9, and 52.5 kg/m², respectively. On average, the residual changes in soil moisture under GLENS are smaller (relative to the baseline) than the changes in RCP8.5. Some of this is driven by the changes in

Figure 7. Same as in Figure 2 but for total liquid runoff (mm/day). Total runoff includes total surface runoff and subsurface drainage.

permafrost regions in RCP8.5; excluding these regions, then for the top 3.8-m soil moisture the $RMSE_A$ under RCP8.5 is 63 kg/m^2 , while GLENS is only slightly smaller at 51.9 kg/m^2 .

We next discuss annual mean regional changes in soil moisture and drivers of soil moisture changes. While many of the changes are not strongly seasonally dependent, some exceptions are described in section 3.4.

3.2. Soil Moisture Regional Changes

Soil water content responds rapidly to changes in the amount and distribution of precipitation. Temperature changes or radiative forcing changes (e.g., with cloud cover changes) also affect soil moisture by influencing ET , and plant water use is further influenced by the physiological effect of CO_2 on vegetation and other

Table 1
Annual Mean Changes From GLENS Ensemble Mean (2075–2095) and RCP8.5 Ensemble Mean (2075–2095), Relative to Baseline Ensemble Mean (RCP8.5 Runs During 2010–2030)

Land-averaged annual mean	Surface air temperature (°C)	Soil moisture (liquid and ice) in the top 10 cm (kg/m ²)	Soil moisture (liquid and ice) in the top 3.8 m (kg/m ²)	Precipitation (mm/day)	ET (mm/day)	Total runoff (mm/day)	Soil liquid water in the top 3.8 m (kg/m ²)	Soil ice in the top 3.8 m (kg/m ²)
RCP8.5	5.38 ± 0.10 (54.1%)	−1.90 ± 0.08 (−7.2%)	−9.5 ± 1.8 (−0.9%)	0.29 ± 0.04 (−0.9%)	0.18 ± 0.01 (−0.9%)	0.10 ± 0.02 (−0.9%)	60.4 ± 1.9 (−0.9%)	−69.9 ± 0.7 (−0.9%)
GLENS	−0.05 ± 0.06 (−0.5%)	0.22 ± 0.05 (0.8%)	1.3 ± 0.9 (0.1%)	−0.03 ± 0.01 (−1.4%)	−0.05 ± 0.00 (−3.3%)	0.01 ± 0.01 (0.9%)	3.3 ± 1.0 (0.9%)	−2.0 ± 0.6 (0.9%)

factors (e.g., increased diffuse light under aerosol geoengineering) which affect plant productivity. While the global averaged changes in soil moisture under GLENS are relatively small, there are regional differences. To explore regional changes, we focus in particular on 10 specific areas indicated in Figure 3. Six of these regions are chosen following Dagon and Schrag (2016) and four additional regions are chosen due to their significant changes in soil moisture under GLENS.

A rise in GHG concentrations under the RCP8.5 scenario is associated with reduced annual mean soil moisture in the top 3.8 m in most land areas, except equatorial North Africa, India, northwest China, Western Russia, and Southern Canada (Figure 3). Soil moisture in the top 3.8 m generally decreases by over 30 kg/m² in the Northern Hemisphere 30°–40° and 50°–70° latitude bands, and Southern Hemisphere 30°–40° latitude band by the end of the century. In most regions, the changes do not have a strong seasonal dependence. However, during boreal summer June–July–August (JJA), soil moisture in Western Russia decreases, owing to the reduced precipitation during this season. This is in contrast to the strong increase in December–January–February (DJF) which dominates the annual mean in that region. Soil moisture changes in the top 10 cm and top 0.8 m reflect a similar pattern (Figure S4). The sign of annual mean changes in soil moisture relative to baseline does not generally change among these different depths, although a different sign of changes under RCP8.5 appears in some mid-high-latitude regions (e.g., Canada and Russia), which could be caused by the permafrost thaw-induced transitions in soil hydrologic conditions (Lawrence et al., 2015; Figure S3).

Figure 4 shows the annual mean percentage change in soil moisture (liquid and ice) in the top 3.8 m under RCP8.5 ensemble mean (2075–2095) and GLENS ensemble mean (2075–2095) relative to baseline ensemble mean (RCP8.5 runs during 2010–2030). Under RCP8.5, there is a significant increase in soil moisture in India, Western Russia, northwest China, and North Africa. However, under GLENS, the soil moisture decreases by 14 ± 9 kg/m² (1.1%), 37 ± 7 kg/m² (3.2%), 10.0 ± 1.8 kg/m² (1.7%), and 17.0 ± 0.6 kg/m² (3.8%) in the Amazon, India, northwest China, and North Africa, respectively (Figures 3 and 4 and Table S1). For India, northwest China, and North Africa, the soil moisture under GLENS is much less than either baseline or RCP8.5 conditions. For the Amazon, there is a decrease in soil moisture relative to baseline; however, it is still wetter than without geoengineering. In the U.S. Central Plains, the U.S. Southwest, and Australia, the soil moisture increases by 31 ± 7 kg/m² (3.0%), 27 ± 8 kg/m² (2.7%), and 25 ± 11 kg/m² (2.8%) compared with baseline, respectively. In addition, the soil moisture in the Congo, Western Russia, and Western Europe is almost unchanged and not significantly different from zero in GLENS.

3.3. Drivers of Soil Moisture Changes

All of the changes are interrelated; that is, changes in precipitation will lead to changes in soil moisture that will lead to changes in ET that will alter precipitation. In most of the regions considered (e.g., Amazon, India, northwest China, U.S. Southwest, and Australia), the dominant factor of changes in soil moisture is precipitation changes. The decreases in soil moisture are mainly due to DJF averaged 0.6 ± 0.4 mm/day (8.0%), JJA averaged 1 ± 0.5 mm/day (16.6%), and DJF averaged 0.04 ± 0.02 mm/day (15.5%) decreases in precipitation in the Amazon, India, and northwest China, respectively, and ET is expected to decrease owing to water availability and the reduced incoming shortwave radiation from GLENS (Figures 5, 6, and S5 and Tables S1–S3). The decreases in precipitation lead to the 0.3 ± 0.2 mm/day (23.6%) and 0.3 ± 0.1 mm/day (31.9%) decrease in summertime total runoff in the Amazon and India, respectively (Figures 7 and 8). These results are consistent with many previous results that show a weakening of the summer monsoon circulation, as well as observations after the 1991 Pinatubo eruption (Kravitz, Rasch, et al., 2013; Tilmes et al., 2013; Trenberth & Dai, 2007). In regions where soil moisture increases, it is mainly due to increased precipitation, with annual mean 0.1 ± 0.1 mm/day (6.6%) for the U.S. Southwest and 0.5 ± 0.2 mm/day (62.2%) averaged during JJA for Australia (Figure 5). In Australia, the annual average ET increases by 0.17 ± 0.09 mm/day (12.0%) due to the increased water availability and the increase in surface longwave radiative flux from cloud adjustment; however, annual average ET is almost unchanged in the U.S. Southwest because the decrease in ground evaporation is balanced by increases in canopy evaporation and transpiration (Figures 6, S5, and S6). While

Figure 8. Distribution of regional average annual values of soil moisture (liquid and ice) in the top 3.8 m (kg m^{-2}), precipitation (mm/day), ET (mm/day), total runoff (mm/day), and surface air temperature ($^{\circ}\text{C}$) under GLENS (2075–2095) and baseline (2010–2030). The distributions shown consist of 420 years (20 members $\times$ 21 years).

increases in ET (e.g., ground evaporation) are dominant for decrease in soil moisture in North Africa, the soil moisture in the top 10 cm is almost unchanged relative to baseline. In the same region there is a significant decrease in soil moisture in the top 3.8 m, because more water in the lower layers can be used to evaporate in arid regions (Figures 3, 6, S4, and S6). In North Africa and the Middle East, $P-ET$ is already negative under the baseline (2010–2030) conditions, and thus, soil moisture would continue to decrease to meet this evaporative demand even if baseline conditions were maintained, that is, even with no change in $P-ET$. That is, the baseline state we are comparing with is not in steady state.

The change in soil moisture is usually consistent with the change in total runoff, but there are some cases where this is not true. In Western Canada, Alaska, and some areas of Russia, more permafrost thaw can decrease soil ice and increase total runoff, while the enhanced ET can result in an overall decrease in soil moisture under RCP8.5 (Figures 3, 5–7, and S3). In the Western Russia and Eastern Canada, there is a significant decrease in runoff which is caused by the decrease in precipitation under GLENS; however, there is a slight increase in soil moisture in top 3.8 m as the decrease in precipitation is counteracted by the decrease in ET and loss of snow (Figures 3, 5–7, and S7). The 20 ensemble members and 21 years give 420 possible differences. These are used in Figure 8 to show the distribution of regional average annual values of soil moisture in the top 3.8 m, precipitation, ET , total runoff, and temperature under GLENS and baseline, demonstrating that these changes can be characterized as a shift in the mean, with little change in the variability.

Stratospheric sulfate geoengineering results in a decrease in shortwave radiative flux at the surface (Figure S5). However, there are increases in shortwave radiative flux in some regions due to adjustments of clouds (decrease in low clouds; e.g., North Atlantic, North Pacific, Eastern Pacific) and surface albedo from changes in sea ice (Figure S5; Kravitz et al., 2017). The increase in surface shortwave radiative flux is consistent with the increase in ET in North Atlantic and Southern Ocean (Figure 6). Under GLENS, increasing GHG concentrations and cloud adjustments affect the change in increasing longwave radiation. The increase in total cloud fraction is consistent with the increase in precipitation in the tropical Eastern Pacific, western South America, Western Asia, southern China, and Australia (Figure 5). Reduced cloud cover in GLENS results in a net positive forcing, meaning that the injection amount must be further increased to prevent surface air temperature increases. The globally averaged total (shortwave + longwave) radiative flux is negative; the decrease in solar irradiance has a stronger effect on the surface than the increase in longwave radiative flux.

ET mostly decreases over land under GLENS, because the decrease in solar radiation acts to limit surface energy available for evaporation, and the increase of atmospheric CO_2 has a direct effect on plant physiology through stomatal conductance and photosynthesis (Figures 6, S5, and S6; Lammertsma et al., 2011). In the zonal mean, the largest decreases in transpiration and thus ET occur in the equatorial region (Amazon, Congo, and Southeast Asia; Figure S6). Ground evaporation shows decreases over most all latitudes, but the canopy evaporation shows increases over most all latitudes which could be related to the increase in canopy intercepted water, with additional influences from the more frequent occurrence of medium precipitation (Figures S1e, S1l, S1n, S6, and S8–S10). While zonal mean transpiration decreases, an increase in plant water use efficiency could help explain the accompanying increase in gross primary productivity (GPP; Figure S6). This increase in GPP could also be driven by an increase in diffuse radiation due to enhanced aerosol scattering (Xia et al., 2016, 2017).

Figure 9 shows the relationship between regional and annual ensemble mean changes in total water storage in soil and surface (including soil moisture (liquid and ice) in the top 3.8 m, canopy intercepted water, surface water, and snow water), and changes in precipitation, ET , and $P-ET$ under GLENS (2075–2095) relative to baseline (RCP8.5 runs during 2010–2030). The changes in water storage in soil and surface are strongly correlated with the changes in precipitation and $P-ET$, with correlation coefficient 0.76 and 0.80, respectively. Here the changes in total water storage in soil and surface are more consistent with the changes in $P-ET$ rather than the changes in precipitation or ET alone. The regional averaged amounts of canopy intercepted water and surface water both are small, although there are some significant changes in some regions considered (Figure S9). The change in snow water is only significant for Western Russia (Figure S9); this effect is a result of winter temperatures in this region remaining higher than the baseline values (Figure 2).

Figure 9. Scatterplots of regional, annual ensemble mean changes in total water storage in soil and surface (including soil moisture (liquid and ice) in the top 3.8 m, canopy intercepted water, surface water, and snow water; kg m^{-2}), precipitation (mm/day), ET (mm/day), and $P-ET$ (mm/day). Change is relative to the baseline period (RCP8.5, 2010–2030 average). Text in each panel gives the label of each region: CON (the Congo), AMA (the Amazon), IND (India), W_RUS (Western Russia), W_EUR (Western Europe), US_C (the U.S. Central Plains), US_S (the U.S. Southwest), AUS (Australia), N_CHI (northwest China), and N_AFR (North Africa).

3.4. Seasonal Changes

There are important seasonal changes in temperature, ET , precipitation, and soil moisture. While increased GHGs tend to warm winter more than summer, geoengineering as simulated in GLENS tends to have the opposite effect, cooling summer more than winter. This effect is a result of both more sunlight in the summer to reflect, and dynamical changes from stratospheric heating (Driscoll et al., 2012).

The objectives for GLENS are defined in terms of annual averages, which does not guarantee zero change in the seasonal response, even for the three temperature objectives that the feedback algorithm manages. There are seasonal differences in temperature as shown in Figure 2. In particular, one of the dominant patterns is overcooling during summer and undercooling during winter (Kravitz et al., 2017). The winter warming at high latitudes, that is, Western Russia, is likely more strongly a result of dynamical effects from stratospheric heating (Driscoll et al., 2012). Although there is $2.6 \pm 0.8^\circ\text{C}$ (40.2%) warming in DJF under GLENS relative to baseline in Western Russia, it still offsets 66.2% of the winter warming under RCP8.5.

The interhemispheric surface temperature ($T1$) goal is met but contains a slight positive bias (stronger warming in the Northern than Southern Hemisphere) for the ensemble mean after 2040 (Tilmes, Richter, Kravitz, et al., 2018). The positive anomaly in $T1$ is greater during DJF than the annual, as Northern Hemisphere winter warming and Southern Hemisphere summer cooling at high latitudes contributes 0.3°C difference in temperature between northern and southern hemispheres (Figure S11). However, the anomaly in $T1$ during JJA is slightly negative (stronger warming in the Southern than Northern Hemisphere) because of Southern Hemisphere winter warming.

ET is generally suppressed over land in GLENS relative to baseline, except in Western Russia during DJF because of winter warming and loss of snow (Figures 6 and S9 and Tables 1 and S3). During JJA, there are increases in ET in the U.S. Southwest, the central region of South America, Australia, and North Africa, which is mainly due to the increase in precipitation during summer (Table S2). The response of zonal mean precipitation under GLENS is negative in the northern and southern latitudes of 40° – 60° , in alignment with the decrease in intensity of the Ferrel cell. In the tropical Pacific, there is a slight bias toward a more El Niño-like mean state (blue values in Figure 10) which has a seasonally dependent influence on hydrological changes, such as in the U.S. Southwest and the central region of South America. In Australia and North Africa, the changes in precipitation during JJA are affected by the atmospheric moisture transport from oceanic sources and seasonal migration of the ITCZ, respectively (Figures 5, S11, and S12).

On seasonal and longer time scales, the ITCZ migrates, typically toward a warming hemisphere (Schneider et al., 2014). We use the centroid of global average precipitation as a proxy for the position of the ITCZ (see Kravitz et al. (2016) for details on the calculation), although the shift in ITCZ depends on the metric. Under RCP8.5 the annual mean value of the precipitation centroid shifts further north by $1.0 \pm 0.1^\circ$, while under GLENS the annual mean precipitation centroid is approximately maintained. Under RCP8.5, there is a 41.3% suppression of seasonal ITCZ migration relative to baseline (the JJA and DJF precipitation centroids are both further north by 0.60 ± 0.16 and $1.4 \pm 0.1^\circ$, respectively; Figure S11). In GLENS, the annual mean precipitation centroid is approximately maintained at the baseline value, compensating for the

Figure 10. Changes in zonally averaged monthly mean soil moisture (liquid and ice) in the top 3.8 m (kg m^{-2}), precipitation over land (mm/day), *ET* over land (mm/day), and precipitation over ocean (mm/day) for (left) GLENS ensemble mean and (right) equatorial ensemble mean. The change is calculated as the difference from an average of the RCP8.5 ensemble mean over years 2010–2030.

northward shift under RCP8.5, which is consistent with the constraint of $T1$ (Figure S11). One result of the overcooling of the summer landmass is that there is less seasonal variation in the ITCZ latitude under GLENS than baseline. The JJA precipitation centroid is further south and the DJF centroid further north by 0.18 ± 0.08 and $0.08 \pm 0.07^\circ$ under GLENS relative to baseline, respectively, so the seasonal extent thus decreases by 13.4%.

The seasonal variation of soil moisture is mainly controlled by seasonal variations in precipitation and ET , and in some regions can also be affected by snow cover in winter (Nie et al., 2008). There is higher seasonal variation in top 10 cm than in the top 3.8 m (Figures 3 and S4). Relative to baseline, the decreases of DJF soil moisture under GLENS in the Amazon and Southern Africa are both much larger than those of JJA, which is consistent with seasonal precipitation cycles (Figures 3 and 5). In central Europe and Southeast Asia, geoengineering as simulated in GLENS leads to decreases in JJA soil moisture, but soil moisture increases in DJF relative to baseline because of seasonal circulation shifts. In these four regions, the seasonal means of total runoff show almost the same trends as seasonal soil moisture variation (Figures 3 and 7). In addition, there is an increase in DJF runoff under GLENS compared with baseline in central Europe, primarily due to the decrease in snowpack under winter warming (Figures 3 and S7).

4. Discussion

4.1. Comparison With Solar Reduction Geoengineering

While a number of studies have considered the impact of geoengineering on precipitation, few have considered the response of soil moisture. One exception is Dagon and Schrag (2016), who examined the impact of SRM on the terrestrial hydrologic cycle, using CLM4 coupled to the Community Atmosphere Model, version 4. The doubling of CO_2 concentrations in Dagon and Schrag (2016) results in 3.1°C global mean warming while the stratospheric aerosols in GLENS are offsetting 4.1°C warming (Table S5). The reduction in total solar irradiance by 2% in Dagon and Schrag (2016) roughly compensates for this change in global mean temperature (referred to below as 2%-SR), with 0.4°C of residual global mean warming relative to present day. Under 2%-SR, the efficacy in global mean temperature (ratio of geoengineering—high GHGs and high GHGs—baseline) is -0.9 , that is, a slight under compensating of the effects of elevated GHGs (Table S5). We do not expect the results to compare exactly, because of the residual temperature response, different GHGs scenarios, and differences in absorbing and scattering aerosols relative to a solar reduction. Global mean ET over land decreased by 25.6 mm/year (4.6%) in 2%-SR, with canopy transpiration changes being the largest contributor to total ET changes. This overall change in ET is roughly consistent with our GLENS results in which land-averaged ET decreased by 17.9 mm/year (3.3%); however, in the GLENS simulations, the ground evaporation and transpiration changes are the two main contributors to total ET changes over land. Precipitation averaged over land decreases by 12.3 mm/year (1.4%) for GLENS, however, increases by 3.7 mm/year (0.6%) for 2%-SR in part due to not fully cancelling the rise in global mean temperature (if the response were linear, a solar reduction sufficient to cancel the global mean temperature change would have resulted in a decrease in land-average precipitation of roughly 5.5 mm/year). Soil moisture in the top 10 cm increases by 0.3 kg/m^2 (0.9%) under 2%-SR, which is closer to the 0.2 kg/m^2 (0.8%) increase under GLENS. But soil moisture of the entire modeled soil column increases by 15.6 kg/m^2 (1.1%) and 1.3 kg/m^2 (0.1%) under 2%-SR and GLENS, respectively. There is substantial disagreement between the two sets of simulation on the sign of the regional changes, for example, the Amazon, Western Russia, the U.S. Southwest, and North Africa.

In principle the differences might result from the use of different climate model configurations (CESM1 (WACCM) with CLM4.5 as the land model in GLENS; CLM4 coupled to Community Atmosphere Model, version 4 including the lack of a fully coupled ocean in Dagon and Schrag (2016)), the difference between stratospheric aerosols versus a solar reduction (e.g., the effect of stratospheric heating), better tailoring of the temperature gradients in our experimental setup, or the different GHG and aerosol emissions scenarios (RCP8.5 was used in GLENS and a fixed CO_2 concentration of 800 ppm was used in 2%-SR), along with not fully cancelling the resulting temperature rise in Dagon and Schrag (2016). For a similar reduction in global temperatures, stratospheric aerosol injection is expected to lead to a greater decrease in the hydrological cycle than sunshade geoengineering (Niemeier et al., 2013; Xia et al., 2017). The absorption of solar and thermal radiation by a stratospheric aerosol layer would increase downwelling thermal radiation that

would result in additional warming of the troposphere, which would then need to be balanced by a greater reduction in downwelling solar radiation. The combination of these two effects would result in increased stability of the troposphere that would lead to less precipitation in certain regions (Andrews et al., 2010; Ferraro & Griffiths, 2016). Xia et al. (2016, 2017) showed increases in transpiration accompanied by increasing GPP (albeit without explicit nutrient limitation) under stratospheric aerosol geoengineering relative to solar geoengineering with the same negative radiative forcing at the top of the atmosphere, partially explained by changes in diffuse radiation (from increased scattering of aerosols). However, Dagon and Schrag (2019) found decreases in transpiration and GPP under solar geoengineering relative to high CO₂, although their simulations differ as they include nutrient limitation but do not include the diffuse effect.

4.2. Comparison With Equatorial Injection Geoengineering

To better understand the effect of managing temperature gradients with very similar sulfur injection amount on the hydrological cycle response, we also compare equatorial-only injection stratospheric sulfate aerosol geoengineering (Kravitz et al., 2019) with GLENS which uses SO₂ injection at four different latitudes. The model and simulation setup are identical for the equatorial geoengineering and GLENS, except that SO₂ was injected into only one grid box at the equator (also at 180°E) to maintain only the global mean temperature. There are two main differences between GLENS and equatorial geoengineering: (1) equatorial geoengineering overcools the tropics and undercools at higher latitudes, which might affect precipitation either directly from the residual temperature change or indirectly due to the change in the meridional heat transport, and (2) equatorial geoengineering requires larger rates of SO₂ injection, and results in more stratospheric heating and hence tropospheric impact of this heating. The “winter warming” effect at high latitudes is stronger in the equatorial run than in GLENS (Figures 2 and S13). This is because 14% more sulfur had to be injected to reach the same temperature reduction resulting in more sulfate mass and more tropical stratospheric heating (Kravitz et al., 2019). Zonally averaged changes in 3.8-m soil moisture, precipitation over land, *ET* over land, and precipitation over ocean are similar in sign between the two simulations; however, the magnitude of changes is generally much larger in the equatorial run than in the multiple injection geoengineering (Figure 10), because of the larger injection amount and more nonuniform aerosol distribution (Kravitz et al., 2019). In India, Australia, and North Africa, the soil moisture in the top 3.8 m is over 11.8 kg/m² (1.8%) wetter under the equatorial run than the GLENS run; however, in North America, the Amazon, and Europe, it is over 12.5 kg/m² (1.1%) drier (Figure S13 and Table S4). The soil moisture changes are associated with statistically significant precipitation differences between GLENS and the equatorial runs in these regions, except for Europe (Figures 3 and S13). In Europe, the drier soil moisture in the equatorial run is mainly influenced by much more *ET* and the stronger winter warming effect (Figures 3, 6, and S13). Both the changes in meridional temperature gradients and the change in stratospheric heating could contribute to the resulting changes in hydrological cycle response; additional simulations are required to disentangle these effects.

5. Conclusion

In our simulations, strategic geoengineering as simulated in GLENS successfully maintains three temperature features at their 2020 values under an RCP8.5 scenario. However, the residual spatial and seasonal temperature changes, along with the stratospheric heating from aerosols, the direct effects of CO₂ on the stability of the troposphere and on vegetation, and the reduction in net surface shortwave could bring about the suppression in the hydrological cycle. The land-averaged soil moisture in the top 3.8 m is nearly unchanged under GLENS relative to baseline, with an increase of 0.1% (1.3 ± 0.9 kg/m²), but there are significant regional changes. In India and the Amazon, soil moisture is reduced by 42 ± 11 kg/m² (3.5%) and 27 ± 16 kg/m² (2.1%) during the summertime, respectively. In most of the regions considered, the dominant cause of soil moisture changes is precipitation changes, while changes in *ET* or snowpack are dominant in a few regions (i.e., North Africa and Western Russia, respectively).

Changes in soil moisture will depend on the scenario, both in terms of emissions (e.g., RCP8.5) and amount of geoengineering. We choose the geoengineering to maintain three temperature features constant at 2020, instead of a hydrologic constraint such as soil moisture. As a result soil moisture does change regionally under GLENS. In addition, changes in soil moisture will be contingent on how geoengineering is deployed, that is, using different latitudes of injection or different aerosols.

The geoengineering approach used in GLENS does not result in zero change in all locations or seasons, as the objectives chosen for GLENS were defined in terms of annual averages. It will be important to understand whether these patterns of precipitation and *ET* changes are robust in different models, to identify physical mechanisms behind them, and testable hypotheses. In addition, more regionally refined models will be needed for the impacts of these changes to society and ecosystems. These results suggest potential benefits, but also residual changes in the hydrological cycle using a stratospheric sulfate aerosol strategy deliberately designed to manage temperature gradients. These results could be used to motivate different strategies, that is, seasonal stratospheric aerosol injection to alleviate hydrological cycle changes in a geoengineered world.

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