



1 **Spatially uneven impacts of stratospheric aerosol**
2 **injection on compound dry-heat wave events in the**
3 **GeoMIP G6sulfur experiment**

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16



17 Abstract

18 Compound dry-heat wave (CDHW) events, characterized by the joint occurrence of extreme
19 heat and dry conditions for at least three consecutive days, pose substantial risks to ecosystems,
20 agriculture, and human societies. However, the implications of stratospheric aerosol injection (SAI)
21 for such compound extremes, particularly at regional scales, remain poorly understood. Here, we
22 investigate changes in CDHW frequency, duration, and severity under the GeoMIP G6sulfur
23 scenario relative to the high-emission SSP5-8.5 scenario using simulations from three Earth system
24 models. We find that SAI substantially offsets the projected intensification of CDHWs over global
25 land, with marked reductions in their frequency, duration, and severity. However, this mitigation is
26 highly heterogeneous: Central Africa emerges as a prominent regional hotspot where CDHWs
27 become more persistent in all three models and, while their frequency also increases under the
28 percentile-based dry-day criterion, primarily in MPI-ESM1-2-LR. The enhancement is driven
29 mainly by the dry side of the compound: dry spells lengthen in all three models (by 0.2–0.9 days
30 per event) and the number of hot-and-dry days per summer rises, even though the total number of
31 heat wave days falls in two of the three models, whereas the intensity of the individual compound
32 events is reduced, consistent with the overall cooling. Comparison with SSP2-4.5 scenario, which
33 has a global-mean radiative forcing comparable to that G6sulfur, indicates that the enhanced CDHW
34 response over Central Africa is not driven by the greenhouse gas forcing component alone. The
35 reduction in greenhouse gas forcing alone produces little change in CDHW exposure over Central
36 Africa, whereas SAI reduces surface shortwave radiation and evaporation over tropical land,
37 promoting mid-tropospheric subsidence and reductions in cloud cover, precipitation, and soil
38 moisture over the region. The contrasting response over the Amazon Basin, where CDHWs decrease,
39 is largely explained by the greenhouse gas component. Reduced greenhouse gas forcing
40 substantially alleviates the projected drying over the Amazon but provides no comparable benefit
41 over Central Africa, where aerosol-specific drying dominates. These hydroclimatic changes are
42 further coupled with declining soil moisture and reduced latent heat flux, partly offsetting the SAI-
43 induced cooling and favoring persistent heat extremes. Our results demonstrate that reducing global
44 mean warming through SAI does not necessarily translate into uniform reductions in compound



45 climate extremes, highlighting the importance of distinguishing the removal of greenhouse gas

46 forcing from aerosol-specific effects, and of land-atmosphere interactions when assessing the

47 climate risks and regional trade-offs of solar climate intervention.

48



49 **1 Introduction**

50 Anthropogenic increases in greenhouse gas concentrations have driven sustained global
51 warming, which has significantly exacerbated the frequency and intensity of extreme weather events
52 (Clarke et al., 2022; Fischer and Knutti, 2015), which are expected to continue intensifying over the
53 coming decades (Seneviratne et al., 2021; Tebaldi et al., 2021). Severe heat waves and extreme
54 droughts are among the clearest signs of this trend (Perkins-Kirkpatrick and Lewis, 2020; Dai, 2013).
55 Prolonged and deadly heat extremes such as the 2003 European heat wave (Stott et al., 2004), the
56 catastrophic 2010 Russian summer (Barriopedro et al., 2011), the 2021 Pacific Northwest heat dome
57 (Philip et al., 2022), and the record European and Chinese summers of 2022 have become more
58 frequent and intense, a trend robustly supported by event attribution studies and coupled model
59 projections (Fischer et al., 2021; Perkins, 2015). Concurrently, the combination of elevated
60 evapotranspiration and shifting precipitation patterns under continued warming has exacerbated soil
61 moisture deficits, driving more frequent and intense dry conditions across numerous regions (Cook
62 et al., 2020; Dai, 2013). These escalating hazards impose devastating burdens on socioeconomic
63 and ecological systems (Rahman and Lateh, 2016).

64 Beyond isolated hazards, compound extreme events have garnered increasing attention.
65 Characterized by the co-occurrence or sequential occurrence of multiple climate drivers
66 (Zscheischler et al., 2018, 2020), these compound hazards are often more destructive than their
67 isolated counterparts (Barichivich et al., 2019; Brás et al., 2021). In particular, compound dry-heat
68 wave (CDHW) events, defined as the simultaneous occurrence of extreme heat and drought, have
69 emerged as a critical area of focus. Such events exert profound adverse impacts on ecosystems,
70 agriculture, water resources, and human health (Hao et al., 2020; Ribeiro et al., 2020; von Buttlar et
71 al., 2018). For example, when isolated dry or heat stress transitions into CDHW events, the risk of
72 corn yield losses is amplified by approximately 20% (Feng et al., 2019), and the non-accidental
73 mortality risk nearly triples (He et al., 2024).

74 Given that emission reductions act slowly on a climate system already committed to further
75 warming, solar climate intervention strategies have attracted growing interest to counteract the
76 escalating risks of extreme weather events (Seneviratne et al., 2021; Wells et al., 2024). Specifically,



77 stratospheric aerosol injection (SAI) is the most widely studied solar radiation modification (SRM)
78 approach to induce surface cooling and alleviate the severe impacts of regional climate hazards
79 (Crutzen, 2006; MacMartin et al., 2018). Mechanistically, SAI mimics the cooling effects of major
80 explosive volcanic eruptions: sulfur dioxide (SO₂) injected into the lower stratosphere forms sulfate
81 aerosols that predominantly scatter incoming solar radiation, reducing the solar energy reaching the
82 surface (Robock, 2000; Tilmes et al., 2018). Sulfate is not a purely scattering aerosol, however. It
83 also absorbs longwave and near-infrared radiation and thereby warms the lower stratosphere, an
84 effect that is one of the pathways by which SAI alters the tropical circulation (Vioni et al., 2021)
85 and one motivation for the continuing interest in more reflective solid particles (Keith et al., 2016).
86 Although idealized simulations indicate that SAI can offset a substantial fraction of global mean
87 warming and attenuate these extreme hazards (Kravitz et al., 2013; Irvine et al., 2019), its
88 implementation entails profound risks and regional trade-offs (Haywood et al., 2013; Kravitz et al.,
89 2019). Notable concerns include the disruption of global precipitation patterns (Tilmes et al., 2013),
90 stratospheric ozone depletion (Pitari et al., 2014), interhemispheric climate asymmetries (Haywood
91 et al., 2013), and the severe "termination shock" that would ensue if injections were abruptly
92 discontinued (Jones et al., 2013). At the same time, many of the systems potentially affected by
93 these side effects are already being disrupted by greenhouse-gas-driven climate change itself, so the
94 risks of deploying SAI must be weighed against the risks of a world without it. Evaluating such
95 risk–risk trade-offs is a central task for SAI assessment (MacMartin et al., 2018; Irvine et al., 2019).

96 To systematically evaluate these multidimensional effects across Earth system models, the
97 Geoengineering Model Intercomparison Project (GeoMIP) established standardized protocols
98 (Kravitz et al., 2011). Within the CMIP6 framework, the G6sulfur experiment specifically prescribes
99 continuous SO₂ injections to reduce the radiative forcing of the high-emission SSP5-8.5 scenario to
100 match that of the intermediate SSP2-4.5 scenario (Kravitz et al., 2015). The underlying SSP
101 scenarios are those of O'Neill et al. (2016), providing a consistent multi-model framework for
102 assessing the efficacy and climate impacts of SAI (Vioni et al., 2021). Building upon these
103 standardized frameworks, existing research has characterized the climate responses to SAI. While
104 simulations consistently demonstrate that SAI effectively reduces global mean surface temperature,



105 they also reveal profound regional disparities within the hydrological cycle (Kravitz et al., 2013;
106 Simpson et al., 2019). Specifically, SAI-induced stratospheric heating modifies the meridional
107 temperature gradient, altering the strength and latitudinal extent of the Hadley circulation, with
108 direct consequences for the distribution of tropical and subtropical precipitation (Ferraro et al., 2014;
109 Staten et al., 2018; Lu et al., 2007). In addition, hemispheric asymmetries in aerosol loading can
110 further displace the intertropical convergence zone (ITCZ), suppressing monsoon rainfall over sub-
111 Saharan Africa and South Asia, regions characterized by high ecological sensitivity, strong
112 agricultural dependence on rainfall, and limited adaptive capacity (Haywood et al., 2013; Robock
113 et al., 2008). Importantly, these hydrological and circulation responses are strategy-dependent. The
114 interhemispheric asymmetries described above arise mainly from single-hemisphere or otherwise
115 unbalanced injection rather than from equatorial injection as such. Single-point injection
116 experiments indicate that injection at the equator produces a weaker tropical circulation and ITCZ
117 response than injection at off-equatorial latitudes. What distinguishes a fixed equatorial strategy
118 such as G6sulfur is instead that it has no latitudinal degrees of freedom with which to compensate
119 residual regional changes, whereas feedback-controlled, multi-latitude strategies (e.g., GLENS and
120 ARISE-SAI) can reduce interhemispheric asymmetries and some of the associated regional
121 precipitation changes by design (Kravitz et al., 2017; Richter et al., 2022; MacMartin et al., 2022).

122 Although the responses of individual heat and precipitation extremes to SAI have been
123 examined (Tilmes et al., 2013; Curry et al., 2014), the compound response of CDHWs has not been
124 systematically evaluated across global land regions. This gap is consequential because CDHWs
125 emerge from the joint probability structure of thermal and hydrological extremes (Zscheischler et
126 al., 2020). Given that SAI may simultaneously suppress surface temperatures and modify
127 precipitation regimes through distinct physical mechanisms, these competing effects could yield
128 highly non-linear outcomes. Therefore, addressing this gap is essential both for advancing
129 mechanistic understanding of SAI's compound climate consequences and for regional risk
130 assessments that any credible governance of SRM must incorporate.

131 In this study, we evaluate how SAI (as implemented in the GeoMIP G6sulfur experiment)
132 modifies the frequency, duration, and severity of CDHWs over global land, with particular attention



133 to the pronounced regional heterogeneity of these impacts. Because the climate outcomes of SAI
134 depend on the injection scenario and strategy (MacMartin et al., 2022), our findings should be
135 interpreted as specific to the G6sulfur implementation, continuous tropical injection against a high
136 emission background, rather than as generic properties of SAI. The remainder of the paper is
137 organized as follows: Sect. 2 presents the datasets and methodology; Sect. 3 presents the projected
138 changes in CDHW characteristics and investigates their regional heterogeneity and underlying
139 physical processes; and Sect. 4 provides the concluding remarks and discussion.

140 **2 Data and methods**

141 **2.1 CMIP6 multi-model simulations**

142 GeoMIP is a structured scientific initiative established in 2010, comprising a comprehensive
143 set of standardized experiments intended to characterize both the physical dynamics and the
144 potential climatic impacts of solar climate intervention (Kravitz et al., 2011). In G6sulfur, the global
145 mean radiative forcing of the high emission SSP5-8.5 pathway is reduced to the level of the
146 moderate SSP2-4.5 scenario by an SO₂ injection, or prescribed aerosol loading, that increases over
147 2020–2100 (Kravitz et al., 2015; Vioni et al., 2021; the SSP scenarios themselves are defined by
148 O'Neill et al., 2016). We investigate the impacts of SAI using three models participating in the
149 G6sulfur protocol (Table 1; further elaborated in Vioni et al., 2021). Daily maximum near-surface
150 air temperature (tasmax, hereafter Tmax) and precipitation, together with the monthly variables
151 required to examine the underlying physical mechanisms, are available for these three models and
152 for UKESM1-0-LL, which was not included in the present analysis; CESM2(WACCM) and MPI-
153 ESM1-2-HR lack part of the required daily or land-surface output. Because the original protocol
154 was non-prescriptive to accommodate varying model complexities (Kravitz et al., 2015), the
155 integration of SAI bifurcates into two fundamental approaches. Specifically, IPSL-CM6A-LR
156 simulated the complete gas-to-aerosol conversion interactively by injecting SO₂ along the
157 Greenwich meridian (10°N–10°S) at altitudes of 18–20 km. In contrast, the remaining two models
158 relied on prescribed aerosol optical depth (AOD) forcing fields. CNRM-ESM2-1 utilized the
159 GeoMIP G4 Specified Stratospheric Aerosol (G4SSA) distribution, which corresponds to an
160 injection of 8 Tg SO₂ yr⁻¹ (Tilmes et al., 2015), rescaled to meet the G6sulfur forcing target, whereas



MPI-ESM1-2-LR prescribed sulfate distributions simulated by ECHAM-HAM for equatorial SO₂ injection (Niemeier and Schmidt, 2017; Niemeier et al., 2020), with the loading rescaled each year to meet the forcing target (Visioni et al., 2021). All model output is obtained from the Earth System Grid Federation (ESGF), with one ensemble member (r1ilp1f1 or r1ilp1f2 as available) used per model for the historical, SSP5-8.5 and G6sulfur simulations alike. CNRM-ESM2-1 and MPI-ESM1-2-LR provide two and three G6sulfur members with daily output, but IPSL-CM6A-LR provides one, and a single member per model keeps the three models equally weighted. The G6sulfur simulation of MPI-ESM1-2-LR ends in 2099. For this model, all analyses of the 2071–2100 period therefore cover 2071–2099. In addition, the SSP2-4.5 simulations of the same ensemble members, whose global-mean radiative forcing G6sulfur is designed to reproduce, are analyzed in Sect. 3.4 in order to separate the effect of the reduced greenhouse-gas forcing from effects specific to the stratospheric aerosol layer. The daily and monthly SSP2-4.5 output was obtained from the ESGF cloud-hosted CMIP6 archive on Google Cloud Storage and conservatively regridded onto a common 1° × 1° grid for subsequent analysis.

Events are identified from daily Tmax (tasmax) and precipitation (pr). The mechanism analysis of Sect. 3.3 additionally uses monthly horizontal winds (ua, va), pressure velocity (wap), specific and relative humidity (hus, hur), total cloud cover (clt), surface soil moisture (mrsos), the four surface radiative flux components (rsds, rsus, rlds, rlus), and the surface latent and sensible heat fluxes (hf1s, hfss).

Table 1. Overview of the CMIP6 climate models and ensemble members used in this study.

Model	Institute	G6sulfur ensemble	SSP5-8.5 ensemble	historical ensemble
CNRM-ESM2-1	CNRM-CERFACS	r1ilp1f2	r1ilp1f2	r1ilp1f2
IPSL-CM6A-LR	IPSL	r1ilp1f1	r1ilp1f1	r1ilp1f1
MPI-ESM1-2-LR	MPI-M	r1ilp1f1	r1ilp1f1	r1ilp1f1

2.2 Identification of CDHW events

In this study, climatic extremes are identified relative to the 1985–2014 reference period. All analyses are restricted to the local summer season, defined as June–August (JJA) in the Northern



184 Hemisphere and December–February (DJF) in the Southern Hemisphere. This hemispheric
185 definition is applied at all latitudes, including the tropics; for the Central African region (5°S–5°N)
186 analyzed below, it therefore combines JJA north of the equator with DJF south of it.

187 A heat wave (HW) event is defined as a period during which T_{max} exceeds the grid-cell-
188 specific local summer 95th percentile threshold for at least three consecutive days, while a dry event
189 is identified when daily precipitation falls below the corresponding local summer 10th percentile
190 threshold for a minimum of three consecutive days (Perkins, 2015; Zscheischler and Seneviratne,
191 2017). Following the compound event framework proposed by Zscheischler and Seneviratne (2017),
192 a CDHW event was identified when both the heat wave and dry event criteria were satisfied
193 simultaneously for at least three consecutive days during the local summer seasons. All percentile
194 thresholds are calculated separately for each model from the local summer days of its historical
195 simulation during 1985–2014 and are determined at each grid cell. This approach enables a model-
196 dependent and spatially consistent identification of extreme events, thereby minimizing potential
197 biases associated with differences in the simulated climatological mean state among models
198 (Seneviratne et al., 2021; Perkins, 2015).

199 The characteristics of HWs, dry events, and CDHW events were quantified using event-based
200 metrics. For all three event types, frequency was defined as the mean annual number of events
201 (events yr^{-1}), while duration was calculated as the mean length of individual events (days event^{-1})
202 (Perkins and Alexander, 2013; Wang et al., 2017).

203 Because the intensity of different event types cannot be represented by a common metric,
204 severity was evaluated separately for HWs and CDHWs. For HWs, severity was defined as the mean
205 exceedance of daily T_{max} above the local 95th percentile threshold during heat wave days and was
206 expressed in °C, representing the average thermal intensity of heat wave events (Perkins and
207 Alexander, 2013). Because no accepted severity metric exists for percentile-based dry spells
208 (Zscheischler and Seneviratne, 2017; Hao et al., 2018), dry events are characterized by frequency
209 and duration only.

210 Following the standardized compound event index framework used in previous studies (Hao
211 et al., 2019; Wu et al., 2020), this study adopts a standardized additive metric combining the thermal



212 and hydrological dimensions of compound dry-heat conditions to quantify the severity of CDHWs.

213 For each day d belonging to an identified CDHW event (i.e., only days within runs of at least
214 three consecutive hot-and-dry days; isolated exceedance days are not scored), the thermal and
215 hydrological severity components were calculated as:

$$216 \quad S_{thermal,d} = \frac{T_{max,d} - T_{95}}{\sigma_T}$$
$$217 \quad S_{hydro,d} = \frac{P_{10} - P_d}{\sigma_P}$$

218 For each grid cell, the 95th percentile of daily maximum temperature (T_{95}) and the 10th
219 percentile of daily precipitation (P_{10}) were calculated based on daily values during the local summer
220 season of the 1985–2014 reference period. The standard deviations of daily maximum temperature
221 (σ_T) and precipitation (σ_P) were derived from the same baseline-period summer samples. All daily
222 precipitation values, including zero-precipitation days, were retained when calculating σ_P . These
223 definitions were applied consistently across all models and grid cells.

224 The total daily CDHW severity ($S_{day,d}$) is the sum of these two dimensionless components.
225 This formulation places the temperature and precipitation anomalies on a common dimensionless
226 scale, preserving their combined effect while allowing each component to be evaluated
227 independently. Both components are standardized by the historical day-to-day variability of the
228 corresponding variable. The difference in Tmax between G6sulfur and SSP5-8.5 is large compared
229 with that variability, whereas the corresponding precipitation difference is small compared with the
230 much larger variability of daily precipitation. Changes in the total severity are therefore dominated
231 by the thermal term (Sect. 3.2), and we interpret them accordingly. To distinguish event intensity
232 from duration, severity was calculated at the event level. Specifically, the severity of a single CDHW
233 event is defined as the arithmetic mean of the daily severities during the event. For a given grid cell
234 and analysis period, the overall CDHW severity (S_{CDHW}) is the arithmetic mean of the severities of
235 all individual events. This equal-weighting approach, where each event is given the same weight
236 regardless of its duration, is consistent with the common practice of treating heat wave intensity and
237 duration as distinct properties of extreme events (Perkins-Kirkpatrick and Lewis, 2020).

238 Because daily CDHW severity is additive, the event-based averaging procedure applies equally
239 to the thermal and hydrological components. Therefore, both the overall CDHW severity and its



240 change between the G6sulfur and SSP5-8.5 scenarios (ΔS_{CDHW}) can be decomposed into the sum
241 of their corresponding thermal and hydrological components. This decomposition effectively
242 characterizes the relative contributions of the thermal and dry dimensions to the changes in severity.
243 Notably, because the number and composition of CDHW events may differ between scenarios,
244 differences in the component metrics reflect both changes in the intensity of the individual
245 dimensions and the impacts of altered event characteristics. Thus, they represent components of the
246 changes in severity, rather than causal factors driving changes in event frequency or duration.

247 Additionally, to examine the sensitivity of the results to the temporal weighting scheme, an
248 alternative day-weighted severity was calculated by directly averaging S_{day} over all CDHW days
249 during the analysis period. The resulting spatial patterns were then compared with those obtained
250 using the primary event-equally weighted approach.

251 In this work, the least squares method is applied to fit the linear trend. Two analysis frames are
252 used throughout. Time-series figures show trends over the full 2020–2100 period, which measures
253 how fast the two scenarios diverge. Spatial maps compare 2071–2100 climatological means, which
254 measures how far apart they end up at the end of the century. In all spatial maps, stippling marks
255 grid cells where every model with a defined value agrees on the sign of the G6sulfur minus SSP5-
256 8.5 difference (i.e., all three models, or both remaining models where the metric is undefined in
257 one). Because precipitation is less than 0.1 mm day^{-1} at 81–90% of land grid cells during the local
258 summer, depending on the model, the percentile-based dry day criterion distinguishes between very
259 small precipitation amounts across much of the land surface. To assess the sensitivity of the CDHW
260 results in Fig. 2 to the dry day definition, the analysis was repeated using two alternative criteria,
261 including a fixed threshold of 1 mm day^{-1} and the 10th percentile of wet day precipitation, with wet
262 days defined as days with precipitation $\geq 1 \text{ mm day}^{-1}$, during the local summer of the reference
263 period (Fig. S5).

264 **2.3 Hadley circulation**

265 The Hadley circulation (HC) represents a crucial large-scale atmospheric mechanism. It plays
266 a pivotal role in shaping regional hydroclimates by concentrating moisture near the equator and
267 transferring both energy and angular momentum from tropical to subtropical zones (Lionello et al.,



268 2024). The HC is defined as the zonal mean meridional streamfunction (Ψ , Oort and Rasmusson,
269 1970; Oort & Yienger, 1996):

$$270 \quad \Psi(p, \phi) = \frac{2\pi a \cos(\phi)}{g} \int_0^p \bar{v}(p, \phi) dp$$

271 where ϕ is the latitude, p is the pressure, a is the radius of the Earth, g is the gravitational
272 acceleration, and $\bar{v}$ is the zonal-mean meridional wind velocity. The strength and extent of the HC
273 are quantified by the cross-section of the Ψ at 500 hPa (Ψ_{500}); that is, the intensity of the Northern
274 Hemisphere HC is represented by the maximum value of Ψ_{500} over the latitudes between 15°S–
275 30°N, and the Southern HC intensity is represented by the minimum value over the latitudes 15°N–
276 30°S. The extent of the Northern Hemisphere HC is defined by the latitude of sign reversal in Ψ_{500}
277 over the latitudes 20°N–50°N and the extent of the Southern hemisphere HC is 20°S–50°S.

278 3 Results

279 3.1 Global mitigation and regional enhancement of CDHWs under SAI

280 The temporal evolution of global-land summer Tmax and precipitation under the G6sulfur and
281 SSP5-8.5 scenarios over the period 2020–2100 is shown in Fig. 1, which compares the summer
282 means with the corresponding averages during CDHWs. All three models demonstrate that SAI
283 substantially mitigates the projected warming, consistent with Visionsi et al. (2021). Under the SSP5-
284 8.5 scenario, the multi-model mean (MME) exhibits a pronounced warming trend of +0.71 °C
285 decade⁻¹. SAI effectively suppresses this overall trend to +0.33 °C decade⁻¹, with this cooling effect
286 becoming particularly pronounced after 2050. This mitigation effect is equally evident during
287 CDHWs (Fig. 1b), where SAI nearly halves the rate of Tmax increase from +0.50 °C decade⁻¹ (under
288 SSP5-8.5) to +0.26 °C decade⁻¹. These results suggest that the reduced intensity of extreme heat
289 under SAI largely reflects the mitigation of greenhouse-gas-induced background warming (Tilmes
290 et al., 2020; Visionsi et al., 2021).

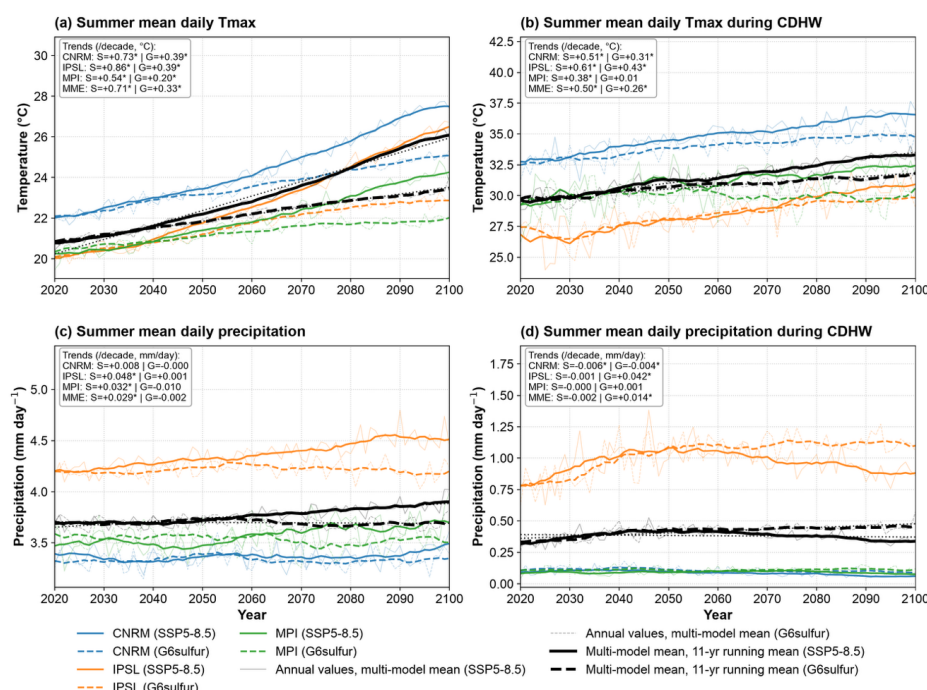


Figure 1. Temporal evolution of summer climate conditions over global land under G6sulfur and SSP5-8.5 during 2020–2100. (a) Summer mean daily maximum near surface air temperature (Tmax), (b) summer mean daily Tmax during CDHW days, (c) summer mean daily precipitation, and (d) summer mean daily precipitation during CDHW days. Thick solid and dashed black lines show the MME under SSP5-8.5 and G6sulfur, respectively. Thin coloured lines show the individual models (CNRM-ESM2-1, IPSL-CM6A-LR, MPI-ESM1-2-LR). Thick curves are 11-year running means and the faint curves behind them are the annual values. Dotted lines are least-squares linear trends over 2020–2100. In the panel insets, S and G are the linear trends under SSP5-8.5 and G6sulfur, respectively, expressed per decade ($^{\circ}\text{C decade}^{-1}$ in a–b, $\text{mm day}^{-1} \text{decade}^{-1}$ in c–d). Asterisks mark trends significant at the 95% confidence level.

Compared with the robust temperature response, precipitation exhibits a substantially weaker and more model-dependent response to SAI (Figs. 1c&1d). For the climatological summer mean (Fig. 1c), the MME under SSP5-8.5 shows a slight increasing trend ($+0.03 \text{ mm day}^{-1} \text{decade}^{-1}$), whereas precipitation remains nearly unchanged under the SAI scenario ($-0.003 \text{ mm day}^{-1} \text{decade}^{-1}$). This muted hydrological response is consistent with previous studies showing that solar climate intervention effectively offsets greenhouse-gas-induced warming but only partially compensates for changes in the global hydrological cycle because reductions in incoming shortwave radiation suppress surface evaporation and latent heat fluxes (Bala et al., 2008; Tilmes et al., 2020). However,

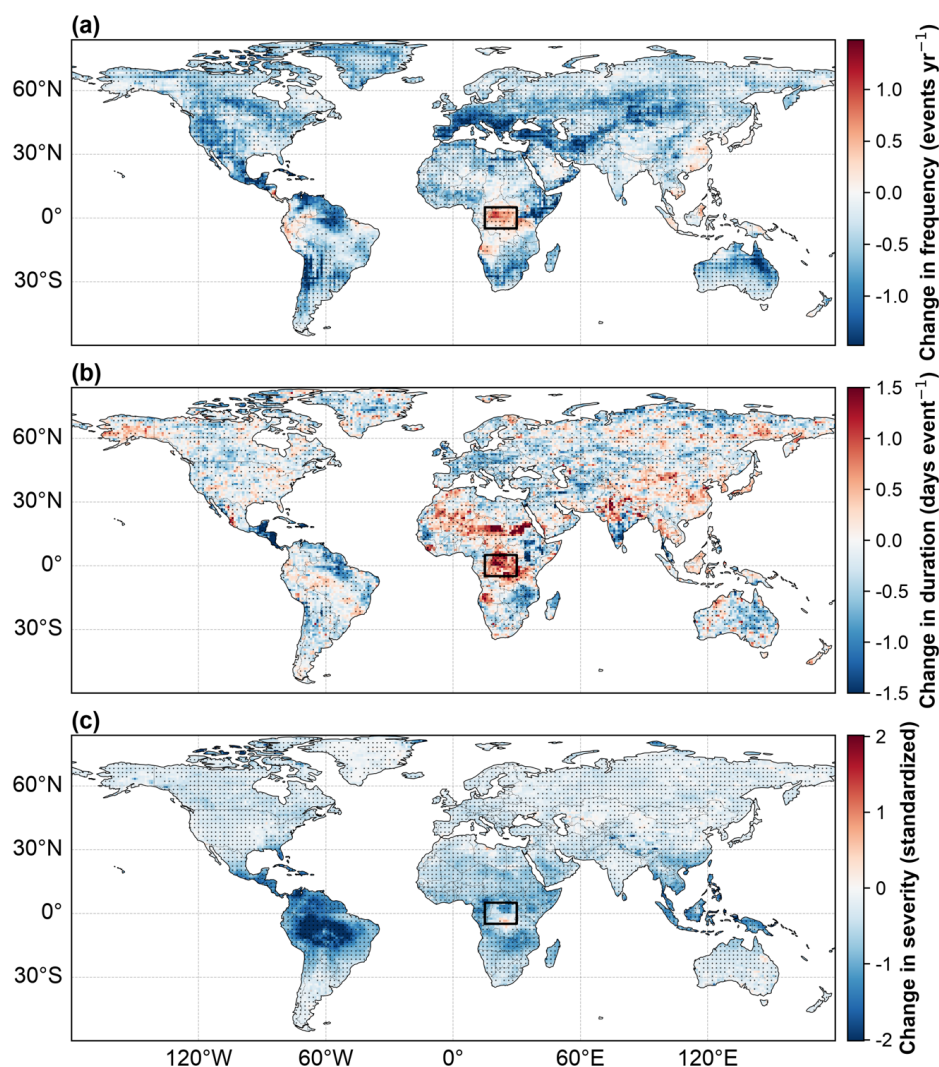


310 the global land mean is not well suited for diagnosing the ITCZ displacement discussed in Sect. 1,
311 because a meridional shift primarily redistributes tropical rainfall rather than substantially changing
312 its total over land. As a result, the displacement signal is largely masked in the global land mean.
313 The regional expression of that redistribution over Central Africa is examined in Sect. 3.3 (Fig. 8i).
314 By construction, precipitation during CDHWs (Fig. 1d) remains low under both scenarios
315 throughout the twenty-first century. The MME exhibits an approximately stationary trend under
316 SSP5-8.5 ($-0.002 \text{ mm day}^{-1} \text{ decade}^{-1}$), whereas a slight increasing trend is found under the SAI
317 scenario ($+0.01 \text{ mm day}^{-1} \text{ decade}^{-1}$). However, the magnitude of this increase is small relative to the
318 spread among individual models. This indicates that the MME response is sensitive to inter-model
319 differences. This sensitivity is expected because only three models are available, with each
320 contributing a single ensemble member. Therefore, structural model differences and internal
321 variability cannot be cleanly separated.

322 Pronounced inter-model variability is evident in both the climatological mean state and during
323 CDHWs. In particular, the IPSL-CM6A-LR model consistently simulates substantially higher
324 precipitation than the CNRM-ESM2-1 and MPI-ESM1-2-LR models under both scenarios, leading
325 to the strongest positive precipitation trend during CDHWs ($+0.04 \text{ mm day}^{-1} \text{ decade}^{-1}$ under SAI).
326 By contrast, the CNRM-ESM2-1 and MPI-ESM1-2-LR models exhibit nearly stationary or slightly
327 decreasing trends, indicating limited hydrological changes under SAI. Consequently, the weak
328 positive trend in the MME primarily reflects the relatively large precipitation simulated by IPSL-
329 CM6A-LR. Similar inter-model differences are also apparent in temperature (Fig. 1b), where the
330 CNRM-ESM2-1 model consistently simulates higher absolute Tmax during CDHWs than the IPSL-
331 CM6A-LR and MPI-ESM1-2-LR models. These inter-model offsets do not affect the event statistics
332 that follow, because all thresholds are defined as percentiles of each model's own reference-period
333 climatology at each grid cell (Sect. 2.2). Consequently, bias adjustment against observations would
334 alter Fig. 1 but would not affect Figs. 2–8. These results demonstrate that although the temperature
335 response to SAI is robust across models, projections of precipitation remain considerably more
336 model dependent, emphasizing the importance of accounting for inter-model uncertainty when
337 assessing the impacts of SAI on CDHWs (Kravitz et al., 2013; Visoni et al., 2021).



338 The spatial distribution of the MME differences in the frequency, duration, and severity of
339 CDHWs under the G6sulfur scenario relative to SSP5-8.5 is presented in Fig. 2. The analysis here
340 focuses on end-of-century climatological differences, complementing the transient changes shown
341 in Fig. 1. Unless otherwise stated, the spatial maps (Figs. 2, 4, 5, and 8) show differences between
342 the 2071–2100 means under G6sulfur and SSP5-8.5, whereas the time series (Figs. 1, 3, and 6) show
343 trends over the full 2020–2100 period. The spatial maps therefore characterize the magnitude and
344 geographical distribution of the end-of-century response, while the time series describe its temporal
345 evolution. Overall, SAI alleviates CDHWs across most global land areas, although the magnitude
346 and spatial coherence of the response differ among the three metrics. The three panels are described
347 in turn below, beginning with severity (Fig. 2c), which shows the most pronounced and spatially
348 coherent response, before turning to frequency and duration (Figs. 2a and 2b). Severity decreases,
349 with widespread negative anomalies over tropical and subtropical land regions. The largest
350 reductions occur across the Amazon Basin, where severity locally decreases by more than 2
351 standardized units, while moderate decreases are also evident over Central Africa, Southeast Asia,
352 and northern Australia. Because all thresholds are defined relative to the 1985–2014 climatology of
353 each individual grid cell (Sect. 2.2), CDHWs over perennially humid regions such as the Amazon
354 denote anomalously hot conditions that coincide with locally dry spells. These dry conditions are
355 defined relative to the local wet climatology rather than in an absolute sense. Consequently, the
356 characteristics of such events differ from those occurring in semi-arid regions. The large reduction
357 in severity is primarily associated with the reduction in the thermal component (Fig. 5a), as SAI
358 substantially offsets the strong warming projected for the region. These widespread reductions are
359 consistent with previous studies showing that SAI effectively offsets greenhouse-gas-induced
360 warming and consequently weakens temperature-related climate extremes (Kravitz et al., 2013;
361 Tilmes et al., 2020; Vioni et al., 2021).



362

363 **Figure 2.** Spatial distribution of changes in compound dry-heat wave characteristics under G6sulfur
364 relative to SSP5-8.5 during 2071–2100. (a–c) Changes in CDHW frequency, duration, and severity,
365 respectively. Values represent the MME differences between G6sulfur and SSP5-8.5 during the local
366 summer season. Frequency is expressed as events yr⁻¹, duration as days event⁻¹, and severity as a
367 dimensionless standardized index. The Central African region is outlined with black boxes in each
368 panel. Stippling indicates grid cells where all models with a defined value agree on the sign of the
369 change. Statistical significance was assessed separately, by applying a two-sided Welch t-test to the
370 30 interannual values of 2071–2100 at each grid cell for each model (a Mann-Whitney U test gives
371 almost identical results). The difference is significant at the 5% level in at least two of the three
372 models over 4.8%, 0.3% and 18.4% of land for (a), (b) and (c) respectively, compared with sign
373 agreement over 15.6%, 37.8% and 66.4%. With a single ensemble member per model, interannual



variability is large relative to the scenario difference for the event-count and event-length metrics in particular, so the patterns in (a) and (b) should be read as indicative rather than as changes detected at the grid-cell scale. Maps are displayed for 60°S–84°N.

Compared with severity, the responses of event frequency and duration are spatially more heterogeneous (Figs. 2a&2b). Frequency generally decreases across North America, Europe, South America, southern Africa, and Australia, whereas localized increases occur over parts of Central Africa and South Asia. Changes in event duration are comparatively weaker and exhibit greater regional variability, although negative anomalies dominate most tropical and subtropical land regions. A notable exception is Central Africa, where frequency and duration both increase, indicating that SAI may locally enhance the occurrence and persistence of CDHWs despite its widespread mitigation effects elsewhere.

To evaluate the robustness of these MME patterns, the responses simulated by the three individual climate models are presented in Fig. S1. The large-scale reduction in CDHW severity is consistently reproduced by all three climate models, particularly over the Amazon Basin and parts of southern Africa, supporting the robustness of the MME results. In contrast, larger inter-model differences are found for event frequency and duration, especially over Central Africa and portions of Asia. Although the three models differ in the magnitude and regional extent of the response, the positive anomalies over Central Africa are mainly associated with the MPI model and, to a lesser extent, the IPSL model, whereas the CNRM model exhibits relatively weak changes in this region. In regional-mean terms the changes are +0.89, +0.23 and +0.07 events yr⁻¹ (frequency) and +0.95, +1.01 and +0.24 days event⁻¹ (duration) for MPI-ESM1-2-LR, IPSL-CM6A-LR and CNRM-ESM2-1, respectively. Although the MME change is positive at more than 90% of the Central African grid cells, all models agree on its sign at only 39% (frequency) and 51% (duration) of them (stippling in Fig. 2). These results suggest that the projected regional response of CDHWs remains more uncertain than the large-scale reduction in event severity, highlighting the importance of considering inter-model spread when assessing regional impacts of SAI (Kravitz et al., 2013; Vioni et al., 2021). In addition, Central Africa (15°–30°E, 5°S–5°N) stands out as the major hotspot exhibiting coherent increases in the frequency and duration of CDHWs under SAI (the sensitivity of this regional signal to the dry-day criterion is assessed in Sect. 4 and Fig. S5). This distinct regional response motivates



403 the subsequent analyses aimed at understanding its temporal evolution and underlying physical
404 mechanisms.

405 The temporal evolution of CDHW frequency, duration, and severity over global land and
406 Central Africa is compared in Fig. 3. For the global mean (Figs. 3a–c), all three CDHW
407 characteristics exhibit persistent upward trends under SSP5-8.5, indicating a progressive
408 intensification of compound dry-heat extremes throughout the twenty-first century. In contrast, these
409 increasing trends are substantially weakened under G6sulfur. The MME frequency, duration, and
410 severity remain consistently lower than those under SSP5-8.5, and the differences between the two
411 scenarios become increasingly pronounced toward the end of the century. Linear trend estimates
412 further illustrate the mitigation effect of SAI. Relative to SSP5-8.5, the rates of increase in CDHW
413 frequency, duration, and severity are reduced from 0.178 to 0.097 events $\text{yr}^{-1} \text{decade}^{-1}$, 0.086 to
414 0.044 days $\text{event}^{-1} \text{decade}^{-1}$, and 0.230 to 0.105 severity units decade^{-1} , respectively (Figs. 3a–c).
415 These results indicate that SAI substantially suppresses the long-term intensification of compound
416 dry-heat extremes in this global land aggregate. This is consistent with previous studies showing
417 that SAI offsets a large fraction of greenhouse-gas-induced warming and reduces temperature-
418 related climate extremes (Kravitz et al., 2013; Irvine et al., 2019; Vioni et al., 2021).
419

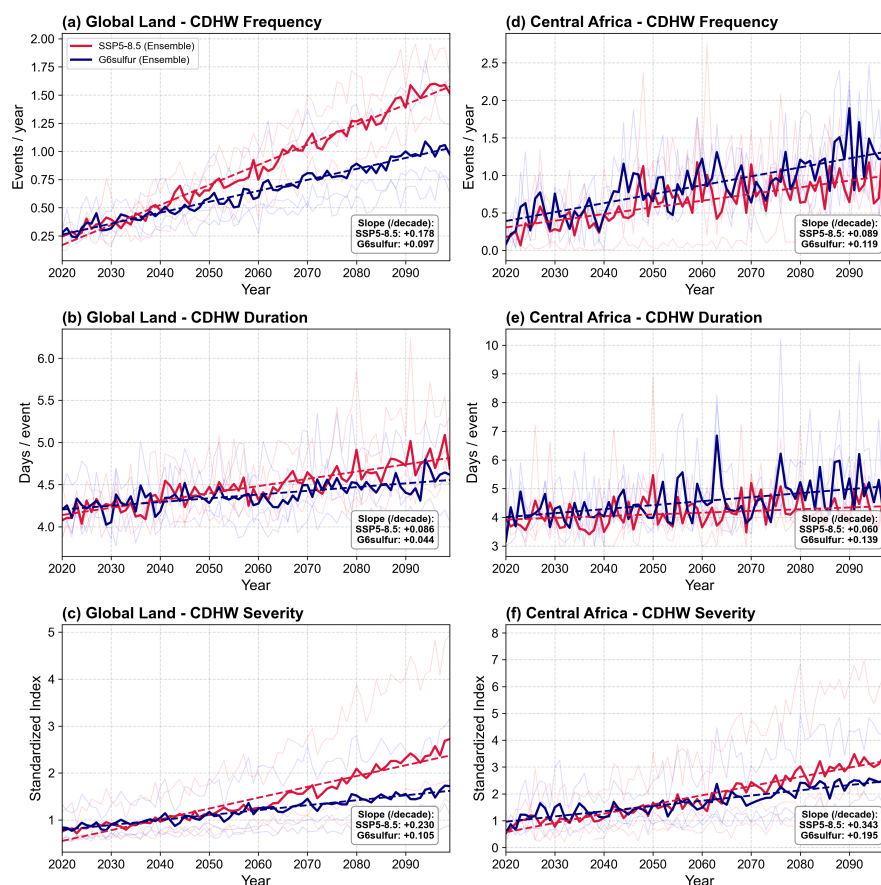


Figure 3. Temporal evolution of CDHW frequency, duration and severity (as distinct from the driving variables of Fig. 1) under G6sulfur and SSP5-8.5 during 2020–2100. (a–c, left column) Global land-mean CDHW frequency, duration, and severity, respectively; (d–f, right column) corresponding changes over Central Africa (15°–30°E, 5°S–5°N). Solid lines show the MME, and linear regression lines indicate the long term trends during 2020–2100. Frequency is expressed as events yr⁻¹, duration as days event⁻¹, and severity as dimensionless standardized units. Thin lines show the individual models.

In contrast to the global response, Central Africa exhibits a distinctly different temporal evolution (Figs. 3d–f). Under G6sulfur, the increasing trends of CDHW frequency and duration exceed those under SSP5-8.5, whereas the severity trend is weaker. Although the two scenarios remain comparable during the early decades, their trajectories gradually diverge after the mid-century, with G6sulfur exhibiting more rapid increases thereafter. The linear trends of CDHW frequency and duration are both larger under G6sulfur than under SSP5-8.5, increasing from 0.089



434 to 0.119 events $\text{yr}^{-1} \text{decade}^{-1}$ and from 0.060 to 0.139 days event $^{-1} \text{decade}^{-1}$, respectively (Figs. 3d
435 and 3e). In contrast, CDHW severity increases considerably more slowly under G6sulfur than under
436 SSP5-8.5 (0.195 versus 0.343 severity units decade^{-1} ; Fig. 3f), so the regional enhancement is
437 confined to the occurrence and persistence of events rather than to their intensity.

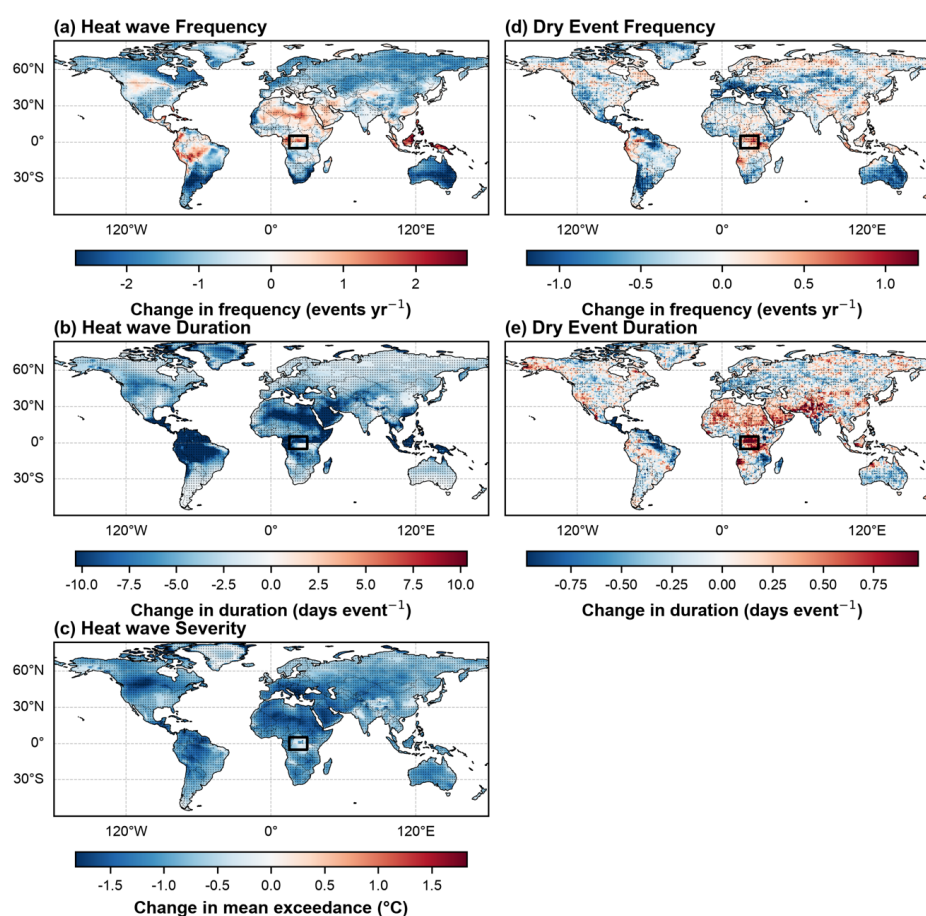
438 The opposite responses observed at the global and regional scales reveal a pronounced spatial
439 heterogeneity in the impacts of SAI on compound climate extremes. While SAI substantially
440 mitigates the long-term intensification of CDHWs over global land, this benefit is not spatially
441 uniform. Instead, Central Africa emerges as the region where CDHW risk is most clearly enhanced
442 rather than reduced under SAI. The Amazon is not the only land area showing a positive anomaly.
443 Smaller increases in CDHW frequency or duration also appear over parts of South Asia and the
444 Maritime Continent (Figs. 2a and 2b). However, the Amazon contiguous the largest area where
445 frequency and duration increase together, with a coherent temporal evolution (Figs. 3d–f) that is
446 further supported by the event decomposition and process-based analyses presented below. We
447 define the region as 15°–30°E, 5°S–5°N, the box that encloses the coherent positive anomaly in
448 Figs. 2a and 2b. The regional means quoted below are not sensitive to modest changes in these limits.
449 This motivates a detailed investigation of the physical processes responsible for the regional
450 response in the following sections.

451 **3.2 Contrasting contributions of heat and dry conditions over Central Africa**

452 To identify the relative contributions of heat and dry conditions to the projected changes in
453 CDHWs, we examine the HW and dry event responses separately (Fig. 4). Taking the panels in
454 order, the response of HW frequency (Fig. 4a) is the most heterogeneous of the three. Negative
455 anomalies dominate the mid and high latitudes of the Northern Hemisphere, whereas positive
456 anomalies emerge across several tropical and subtropical regions, notably the Amazon Basin,
457 Central Africa, and parts of the Maritime Continent. In several of these regions, HW frequency
458 increases by more than 2 events yr^{-1} , indicating that SAI does not uniformly reduce the number of
459 heat wave events. HW duration (Fig. 4b) and severity (Fig. 4c), by contrast, decrease across most
460 global land areas, with widespread negative anomalies over North America, South America, Europe,
461 northern Asia, southern Africa, and Australia. These reductions indicate that SAI suppresses the



462 persistence and intensity of heat waves by offsetting greenhouse-gas-induced warming, consistent
463 with previous studies (Kravitz et al., 2013; Tilmes et al., 2020). As shown in Fig. 6, the increase in
464 event count and the decrease in event length are not independent. When exceedances remain nearly
465 continuous under SSP5-8.5, even a modest cooling can interrupt the continuity and split a single
466 long event into several shorter events.



467 **Figure 4.** Spatial distribution of changes in individual heat wave and dry event characteristics under
468 G6sulfur relative to SSP5-8.5 during 2071–2100. (a–c, left column) Changes in heat wave frequency,
469 duration, and severity, respectively; (d–e, right column) changes in dry event frequency and duration,
470 respectively. All values represent MME differences between G6sulfur and SSP5-8.5 during the local
471 summer season. Heat wave frequency and dry event frequency are expressed as events yr⁻¹, duration
472 as days event⁻¹, and heat wave severity as the mean exceedance of T_{max} above the local 95th
473 percentile threshold (°C). The Central African region is outlined with black boxes in each panel.
474 Stippling indicates grid cells where all models with a defined value agree on the sign of the change.
475 As in Fig. 2, significance was assessed separately using a two-sided Welch's t-test based on the 30
476



interannual values from 2071–2100. The differences are significant at the 5% level in at least two of the three models across 60.9%, 67.5%, and 70.4% of the global land area for the heatwave panels (a–c), respectively. In contrast, significant differences occur over only 1.2% and 0.5% of the land area for the dry event panels (d) and (e), respectively, indicating substantially greater uncertainty in these changes. The colour scale of (b) is truncated at ± 10 days event⁻¹ (the 90th percentile of the absolute MME change over land). Larger reductions, which reach about –40 days event⁻¹ locally, are shown by the end colour of the scale. Maps are displayed for 60°S–84°N.

Dry events exhibit a different spatial response. SAI generally reduces dry event frequency across much of North America, Europe, southern Africa, Australia, and Asia (Fig. 4d). Nevertheless, positive frequency anomalies remain evident over parts of Central Africa and Maritime Southeast Asia, indicating that SAI does not uniformly suppress dry event occurrence. An even stronger regional signal is found in dry event duration (Fig. 4e), which exhibits substantial positive anomalies over Central Africa exceeding approximately one day per event, along with localized increases over Maritime Southeast Asia. Conversely, prominent negative anomalies are observed over the Amazon Basin. These results suggest that although SAI generally reduces the occurrence of dry events globally, it may simultaneously prolong individual dry spells in specific tropical regions like Central Africa. Overall, the responses of HWs and dry events reveal pronounced regional heterogeneity. While SAI substantially weakens heat wave duration and severity and reduces dry event occurrence across most land areas, certain tropical regions, particularly Central Africa, exhibit an increase in HW frequency accompanied by a prolongation of dry event duration. This suggests that the regional response of compound extremes differs fundamentally from the global mean.

The MME responses are generally supported by the individual climate models (Figs. S2&S3). All three models consistently reproduce the widespread reductions in HW duration and severity, whereas larger inter-model differences are evident for HW frequency and dry event characteristics, particularly over tropical regions. Nevertheless, positive anomalies in HW frequency and dry event duration over Central Africa are reproduced by multiple models, indicating that the regional enhancement identified in the MME is not solely driven by a single model, although uncertainties remain regarding its exact magnitude and spatial extent.

To test whether the projected enhancement of CDHWs over Central Africa is accompanied by any increase in the intensity of the individual events, we decompose the projected changes in CDHW severity into thermal and hydrological components (Fig. 5). The purpose of this figure is to establish



508 a null result: if the Central African enhancement were an intensity signal, it would appear here as a
509 positive anomaly in one of the two components over the region, and it does not. The thermal
510 component (Fig. 5a) exhibits widespread negative anomalies across most global land areas,
511 reflecting the overall reduction in heat-related stress resulting from SAI-induced surface cooling,
512 consistent with previous GeoMIP studies (Kravitz et al., 2013; Irvine et al., 2019). Central Africa is
513 not an exception in this respect. Its thermal component also decreased by 0.48 standardized units,
514 which is consistent with the reduction in absolute Tmax under SAI (Fig. 8j). By comparison, the
515 hydrological component (Fig. 5b) displays much weaker spatial variability. Over Central Africa it
516 is essentially zero (+0.01 standardized units). The thermal component therefore accounts for almost
517 the whole of the local change in severity (-0.48 of -0.47 standardized units). These results indicate
518 that the enhancement of CDHWs over Central Africa is not primarily an intensity signal. Instead, it
519 arises from more frequent and more persistent exceedances of the joint threshold. We caution,
520 however, that because the standardized severity metric is dominated by its thermal component (Sect.
521 2.3), the reduction in event severity largely restates the SAI-induced cooling itself and should not
522 be read as an independent measure of overall event impacts.

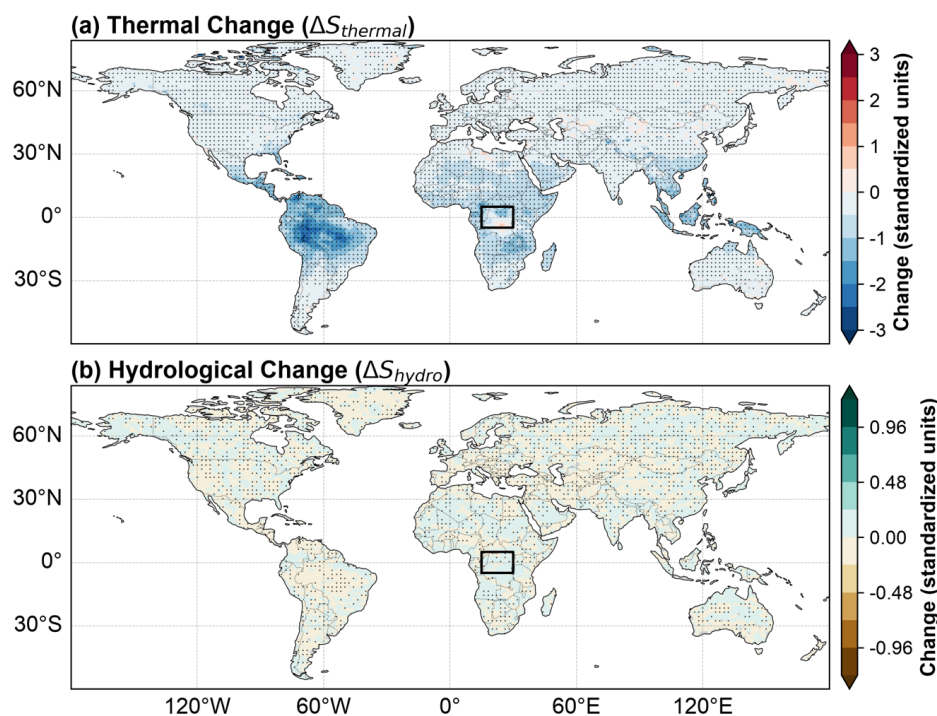


Figure 5. Spatial distribution of the thermal and hydrological contributions to changes in CDHW severity under G6sulfur relative to SSP5-8.5 during 2071–2100. (a) Change in the thermal component of CDHW severity and (b) change in the hydrological component. The thermal component represents the contribution associated with exceedance of T_{max} above the local 95th percentile threshold, whereas the hydrological component represents the contribution associated with precipitation deficits below the local 10th percentile threshold. Values are shown as multi-model mean differences between G6sulfur and SSP5-8.5 during the local summer season. The Central African region is outlined with black boxes in each panel. Stippling indicates grid cells where all models agree on the sign of the change. Maps are displayed for 60°S–84°N.

To identify which event characteristics drive the enhanced CDHW response over Central Africa, Fig. 6 presents the temporal evolution of individual HW and dry event characteristics. Heat wave frequency increases under both scenarios throughout the twenty-first century (Fig. 6a). Although the long-term trend is slightly weaker under G6sulfur than under SSP5-8.5 (0.340 versus 0.395 events decade⁻¹), the MME remains generally higher during much of the simulation period after the 2040s, indicating more frequent heat wave events under SAI. This raises the question of whether the higher event count is partly an artefact of how exceedances are grouped into events. Under SSP5-8.5, the region remains persistently above its historical 95th percentile, causing heat waves to



merge into a few very long episodes. A modest cooling can then split one such episode into several shorter ones without adding any hot days. Counting total exceedance days per summer over Central Africa confirms that this is what happens for heat waves. Between SSP5-8.5 and G6sulfur, heat wave days fall from 52.6 to 36.5 per summer in CNRM-ESM2-1 and from 85.8 to 76.0 in IPSL-CM6A-LR while the event count rises in IPSL-CM6A-LR (2.3 to 3.4 events yr^{-1}) and mean event length collapses (56.9 to 33.7 days event^{-1}). Only MPI-ESM1-2-LR shows more heat wave days (18.0 to 25.6). The increase in heat wave frequency over Central Africa is therefore largely a fragmentation effect and should not be read as an increase in heat wave exposure. The compound metric shows a different response. Total CDHW days per summer increase in all three models, with increases of 0.7, 2.1 and 5.7 days in CNRM-ESM2-1, IPSL-CM6A-LR and MPI-ESM1-2-LR, respectively. The increase is statistically significant at the 5% level in MPI-ESM1-2-LR and marginally significant in IPSL-CM6A-LR. Total dry days also increase by 1.1, 2.3 and 4.3 days in three models, respectively. The enhancement of compound dry heat exposure over Central Africa is thus robust to the event-counting convention, and it is carried by the dry component rather than by the thermal one. In contrast, heat wave duration (Fig. 6b) increases substantially more slowly under G6sulfur (1.934 versus 3.153 days $\text{event}^{-1} \text{decade}^{-1}$), suggesting that SAI partly suppresses the persistence of individual heat wave events. Likewise, heat wave intensity (Fig. 6c) increases under both scenarios but at a reduced rate under G6sulfur (0.139 versus 0.279 $^{\circ}\text{C} \text{decade}^{-1}$), consistent with the overall cooling effect of SAI (Irvine et al., 2019).

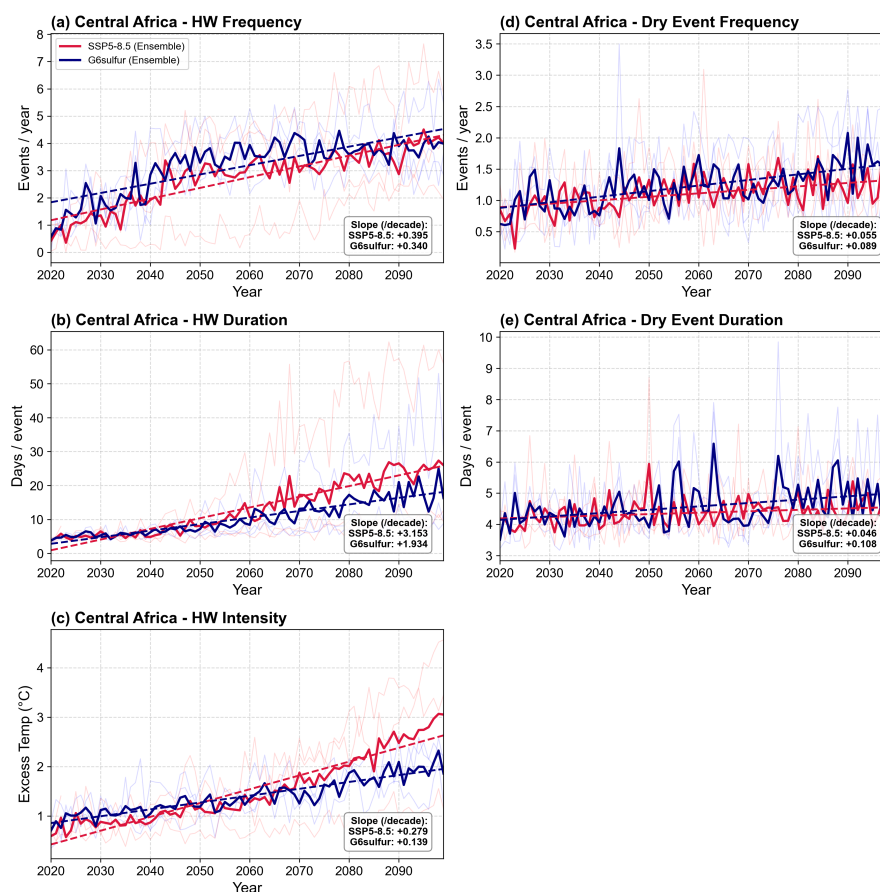


Figure 6. Temporal evolution of heat wave and dry event characteristics over Central Africa during 2020–2100. (a–c, left column) Heat wave frequency, duration, and severity, respectively; (d–e, right column) dry event frequency and duration, respectively. Solid lines represent the MME under G6sulfur and SSP5-8.5, and linear regression lines indicate the corresponding long-term trends. Heat wave and dry event frequency are expressed as events yr⁻¹, duration as days event⁻¹, and heat wave severity as the mean Tmax exceedance above the local 95th percentile threshold (°C). The Central African region is defined as 15°–30°E and 5°S–5°N.

In contrast, dry events exhibit a different temporal evolution. Although their absolute frequencies remain broadly consistent across both scenarios (Fig. 6d), G6sulfur drives a slightly accelerated increasing trend (0.089 versus 0.055 events decade⁻¹). Similarly, dry event duration increases more rapidly under G6sulfur than under SSP5-8.5 (0.108 versus 0.046 days event⁻¹ decade⁻¹), indicating a tendency toward more persistent dry conditions under SAI (Fig. 6e). Taken together with the exceedance day totals given above, these hydrological changes are the primary



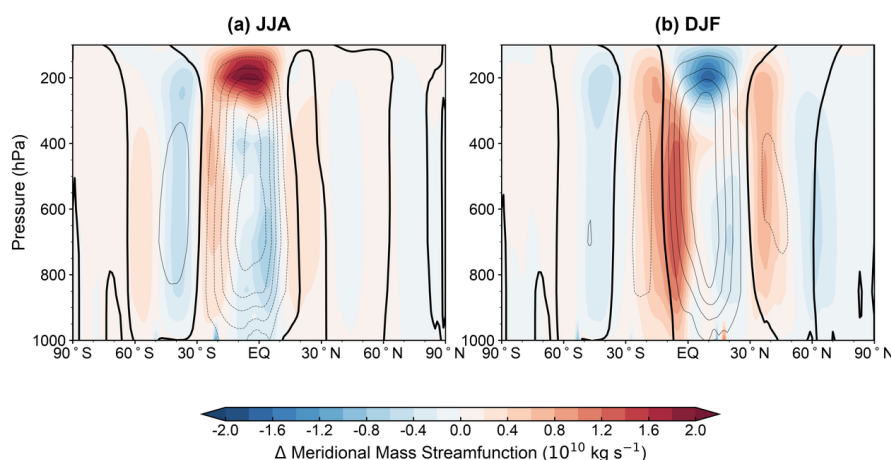
575 driver of the compound signal rather than a secondary contribution. Dry days and dry event duration
576 increase over Central Africa in all three models, whereas total heat wave days decrease in two of
577 the three, so the enhanced CDHWs under G6sulfur cannot be attributed to an increase in heat
578 exposure. The thermal component contributes primarily to a change in the temporal structure of
579 exceedances rather than their total number. Heat waves become shorter and more numerous, raising
580 the number of separate opportunities for a hot spell to coincide with a dry spell, while the dry spells
581 themselves lengthen. The compound enhancement emerges from that coincidence. This regional
582 behavior contrasts sharply with the global mean response, highlighting the importance of regional
583 climate processes in modulating the impacts of stratospheric aerosol injection on compound climate
584 extremes.

585 **3.3 Circulation and land–atmosphere processes underlying CDHW enhancement**

586 The MME differences in the zonal-mean meridional mass streamfunction between G6sulfur
587 and SSP5-8.5 during JJA and DJF are shown in Fig. 7. Overall, SAI produces a modest but
588 seasonally structured adjustment of the Hadley circulation. Quantified as in Sect. 2.3, the dominant
589 cell in each season changes in strength by about 1% and its poleward edge shifts by less than 0.4°
590 of latitude, while the anomaly pattern is distinctly seasonal and hemispherically asymmetric. During
591 JJA, positive streamfunction anomalies are concentrated over the deep tropics, with the largest
592 values located in the upper troposphere and lower stratosphere, while negative anomalies are
593 distributed along the flanks of the tropical overturning circulation. In contrast, the circulation
594 response reverses during DJF, with positive anomalies dominating the tropical Southern
595 Hemisphere and negative anomalies extending across the tropics north of the equator. These
596 contrasting anomaly patterns indicate a pronounced seasonal dependence in the large-scale
597 circulation response to SAI. The circulation anomalies are primarily confined to the upper
598 troposphere and lower stratosphere, suggesting that the response extends through a deep vertical
599 layer of the atmosphere rather than being limited to the mid-troposphere. Such a vertically coherent
600 and seasonally varying response is consistent with previous studies showing that stratospheric
601 aerosol forcing can modify the meridional temperature gradient and alter the large scale tropical
602 overturning circulation (Ferraro et al., 2014; Staten et al., 2018). Moreover, hemispherically



603 asymmetric radiative forcing has been shown to influence the position of the ITCZ and the
604 associated tropical precipitation distribution (Haywood et al., 2013; Robock et al., 2008). Those
605 single hemisphere experiments represent extreme cases, and feedback-controlled strategies that
606 balance injections across latitudes (e.g., GLENS and ARISE-SAI) show that such asymmetries can
607 be largely avoided by design (Kravitz et al., 2017; Richter et al., 2022). However, a zonal-mean
608 adjustment of this magnitude fails to explain the localized response confined to 15°–30°E. Since a
609 zonal-mean anomaly is by definition identical at all longitudes, it applies equally to the Amazon
610 Basin, which lies within the same latitude band but exhibits the opposite response (Figs. 2, 4e and
611 8f). Therefore, following the description of the regional fields, Sect. 3.4 explores the zonally
612 asymmetric tropical circulation response. It further assesses how much of the Central African signal
613 stems from the aerosol layer as opposed to the reduced greenhouse-gas forcing. The individual
614 model results (Fig. S4) broadly reproduce the seasonal anomaly patterns shown by the MME,
615 despite differences in anomaly magnitude and vertical extent among three models, indicating that
616 the simulated adjustment of the Hadley circulation, although modest in amplitude, represents a
617 robust response to SAI as implemented here (sustained tropical injection) rather than being
618 dominated by a single model. The amplitude of the tropical response depends on where the aerosol
619 is placed, and single-point injection experiments indicate that equatorial injection produces a weaker
620 tropical circulation and ITCZ response than injection at off-equatorial latitudes, so the modest
621 adjustment found here is plausibly a lower bound on what other injection strategies would produce
622 (Kravitz et al., 2019; MacMartin et al., 2022).



623

624 **Figure 7.** Seasonal response of the Hadley circulation to G6sulfur relative to SSP5-8.5. Multi-model
625 ensemble mean differences in the zonal-mean meridional mass streamfunction during (a) JJA and
626 (b) DJF. The meridional mass streamfunction is expressed in $10^{10} \text{ kg s}^{-1}$, with positive and negative
627 values denoting opposite directions of the zonal-mean meridional overturning circulation. The
628 vertical coordinate is pressure (hPa), and the anomalies are calculated as G6sulfur minus SSP5-8.5
629 over 2071–2100. Black contours show the SSP5-8.5 multi-model mean climatological
630 streamfunction (contour interval $4 \times 10^{10} \text{ kg s}^{-1}$; solid for positive and dashed for negative values),
631 and the thick contour marks the zero line separating the two Hadley cells.

632 The atmospheric and land-surface responses that characterize Central Africa are examined in
633 Fig. 8. The regional response is characterized by a coherent coupling between atmospheric stability,
634 moisture conditions, precipitation, and land-surface energy partitioning. At 500 hPa, positive omega
635 anomalies are evident over Central Africa, indicating enhanced subsidence, while the wind at 850
636 hPa anomalies are weak within the box itself (below 0.05 m s^{-1} in both components) and the coherent
637 low-level signal lies to its west, where an easterly anomaly of about 0.4 m s^{-1} over the Gulf of
638 Guinea and the western Congo Basin (0° – 15°E) opposes the climatological low-level westerly
639 inflow from the tropical Atlantic (MME 1.0 m s^{-1} under SSP5-8.5, although CNRM-ESM2-1
640 simulates weak easterlies there) (Fig. 8a). The moisture flux convergence at 850 hPa anomaly over
641 Central Africa is relatively weak and locally positive, suggesting that the reduction in precipitation
642 cannot be attributed simply to a large decrease in low level moisture convergence (Fig. 8b). Instead,
643 the enhanced mid-tropospheric subsidence is accompanied by a decrease in relative humidity at 500
644 hPa (Fig. 8c), indicating a drier and potentially less favorable environment for deep convection.
645 Consistent with these changes, cloud cover decreases over Central Africa (Fig. 8d), while



precipitation exhibits a corresponding negative anomaly (Fig. 8i). These anomalies are part of a broader tropical pattern rather than a feature confined to the box: relative humidity, precipitation and cloud cover decrease over roughly two-thirds of tropical land. What singles out Central Africa is the amplitude and the co-location of the individual terms. Its subsidence anomaly lies in the 95th percentile of tropical land, its relative-humidity at 500 hPa anomaly in the driest 7% and its precipitation anomaly in the driest 11%, and these coincide with the region where the CDHW metrics of Sect. 3.1 increase. Such a linkage between large-scale circulation changes and regional precipitation responses is consistent with previous studies showing that stratospheric aerosol forcing can modify tropical circulation and precipitation through changes in atmospheric heating and stability (Ferraro et al., 2014; Robock et al., 2008).

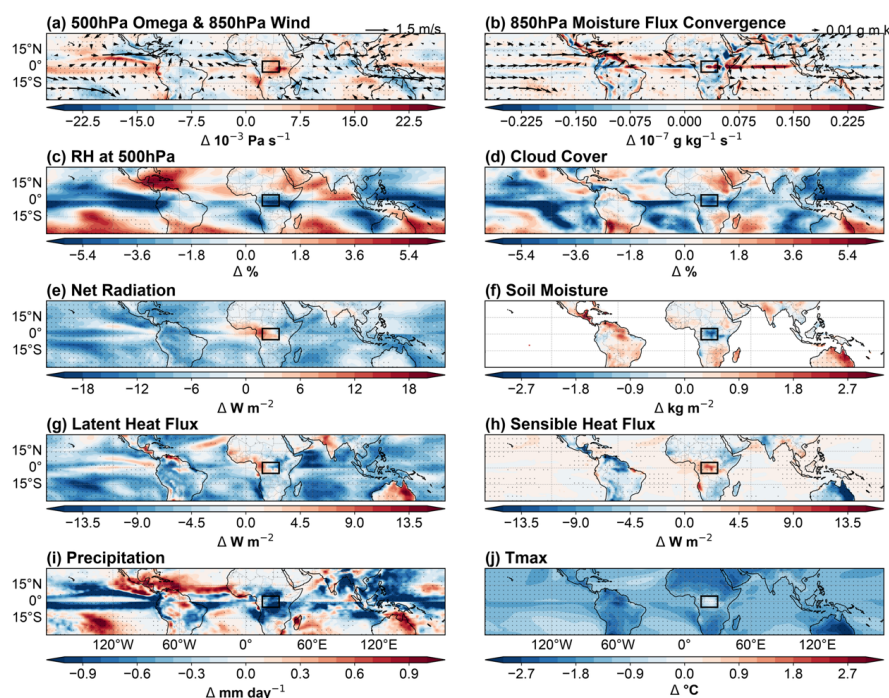


Figure 8. Atmospheric and land surface responses over Central Africa to G6sulfur relative to SSP5-8.5 during 2071–2100. (a) 500 hPa vertical velocity (ω) with 850 hPa horizontal wind anomalies, (b) 850 hPa moisture flux convergence, (c) 500 hPa relative humidity, (d) total cloud cover, (e) net surface radiation (net downward shortwave plus longwave radiative flux), (f) soil moisture, (g) latent heat flux, (h) sensible heat flux, (i) precipitation, and (j) daily maximum near-surface air temperature (Tmax). All fields show multi-model mean differences between G6sulfur and SSP5-8.5 during the local summer season. Stippling indicates grid cells where all three models agree on the



664 sign of the change. The Central African region is outlined with black boxes in each panel. Maps are
665 displayed for 30°S–30°N so that the regional structure over Central Africa is legible. The full
666 longitude range is retained so that the extent of the surrounding tropical anomaly pattern remains
667 visible.

668 The atmospheric drying is accompanied by a coupled response of the land surface. Soil
669 moisture decreases over Central Africa. Although the absolute anomaly is modest compared with
670 the more pronounced drying signals observed in some other tropical regions, the reduction relative
671 to the region's own climatological soil moisture is more relevant to the surface energy partitioning
672 discussed below than its absolute magnitude (Fig. 8f). At the same time, the MME latent heat flux
673 decreases over the region whereas the sensible heat flux increases (Figs. 8g–h), although the models
674 agree on the sign of these two flux changes over less than half of the region. This indicates a shift
675 in surface energy partitioning away from evaporative cooling and toward sensible heating.
676 Consistent with this shift, the regional-mean warm season Bowen ratio over Central Africa increases
677 in all three models under G6sulfur relative to SSP5-8.5 (from 0.74 to 0.81 in CNRM-ESM2-1, from
678 0.31 to 0.34 in IPSL-CM6A-LR, and from 0.17 to 0.26 in MPI-ESM1-2-LR), even though the
679 models disagree on the sign of the individual flux changes over much of the region. Such a response
680 is consistent with the established soil moisture energy feedback, whereby reduced soil moisture
681 limits evapotranspiration and latent heat loss, allowing a larger fraction of the available surface
682 energy to be expressed as sensible heat (Seneviratne et al., 2010; Lorenz et al., 2010). Here, net
683 surface radiation is defined as the net downward shortwave plus longwave radiative flux,
684 representing the radiative energy available at the surface for partitioning into the turbulent fluxes.
685 Its response is small over Central Africa and not robust across models, which is consistent with the
686 strong dependence of surface radiative and turbulent fluxes on model parameterizations of clouds
687 and land-surface processes. The regional mean change is $+3.8 \text{ W m}^{-2}$, with CNRM-ESM2-1 and
688 IPSL-CM6A-LR simulating decreases and MPI-ESM1-2-LR simulating an increase (Fig. 8e). In
689 the two models that simulate a reduced latent heat flux, the enhanced sensible heating therefore
690 reflects a redistribution of the available surface energy rather than an increase in the radiative energy
691 input at the surface, whereas in MPI-ESM1-2-LR both turbulent fluxes increase together with net
692 radiation (Figs. 8e, g–h).

693 Importantly, the surface temperature response does not show local warming. In the MME and



694 in two of the three individual models, Tmax decreases over Central Africa under G6sulfur relative
695 to SSP5-8.5, with a regional mean decrease of 0.79 °C. MPI-ESM1-2-LR is the exception, showing
696 a slight increase in Tmax (Fig. 8j). These results are consistent with the overall cooling effect of
697 SAI. Nevertheless, the concurrent reductions in precipitation, soil moisture, and latent heat flux
698 indicate that the regional hydroclimatic response is not simply characterized by uniform cooling.
699 Instead, SAI produces a cooler but drier land surface with weakened evaporative cooling and
700 enhanced sensible heat partitioning, providing a physical environment that can favor the persistence
701 or occurrence of heat extremes despite the reduction in absolute temperature. Soil moisture
702 temperature coupling and associated land-atmosphere feedbacks are known to be particularly
703 important for regulating temperature extremes under moisture limited conditions (Seneviratne et al.,
704 2010; Miralles et al., 2014). These results suggest that the enhanced CDHW response over Central
705 Africa arises from regional hydroclimatic drying and land-surface energy redistribution, rather than
706 a general increase in mean temperature. Although Central Africa and the Amazon Basin are located
707 at similar latitudes, they exhibit contrasting hydroclimatic responses. The subsequent section
708 explores the physical mechanisms behind this spatial contrast and isolates whether the localized
709 drying in Central Africa is directly induced by stratospheric aerosols or solely driven by the reduced
710 greenhouse-gas forcing.

711 **3.4 Removal of greenhouse gas forcing versus aerosol-specific effects**

712 Based on the SSP5-8.5 emission pathway, G6sulfur implements SAI to offset the global-mean
713 radiative forcing down to SSP2-4.5 levels. Consequently, the anomalies of G6sulfur relative to
714 SSP5-8.5 discussed thus far reflect a combined response to both the equivalent reduction in
715 greenhouse-gas forcing and the unique physical effects of the aerosol layer. These two contributions
716 can be separated by including the SSP2-4.5 simulations of the same ensemble members (Sect. 2.1).
717 Specifically, the difference between SSP2-4.5 and SSP5-8.5 isolates the climatic impact of the
718 reduced greenhouse gas. Conversely, the difference between G6sulfur and SSP2-4.5 represents the
719 residual response under matched global-mean forcing, which comprises the aerosol-specific effects
720 together with any imperfection in the forcing match. Fig. 9 presents these differences for vertical
721 velocity at 500 hPa, precipitation, surface soil moisture, and CDHW days per summer over the



Amazon and Africa sectors. The corresponding global tropical fields and regional means are provided in Figs. S6–S7 and Table S1, respectively. By construction, the G6sulfur minus SSP5-8.5 difference equals the sum of the other two.

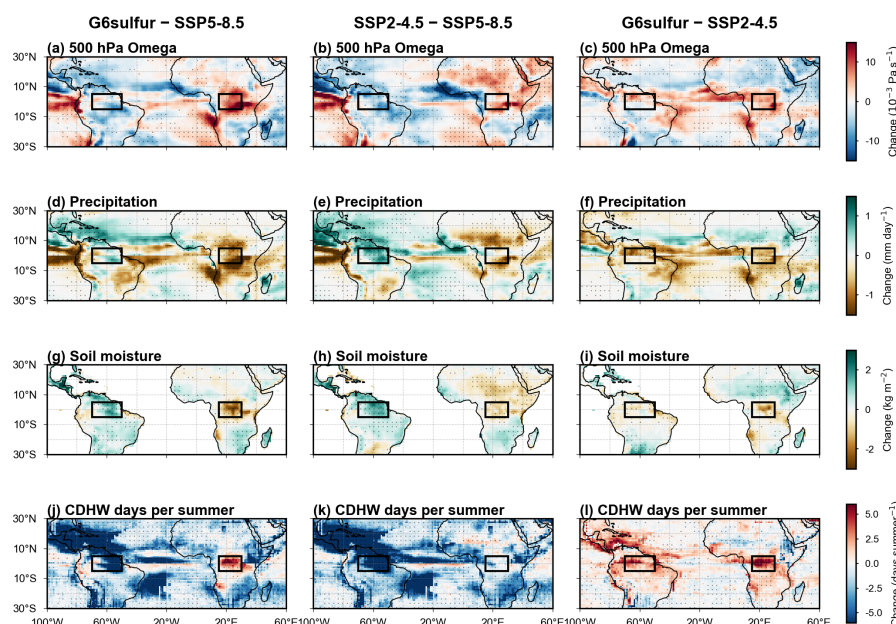


Figure 9. Separation of the G6sulfur response over the Amazon and Africa sector into greenhouse gas and aerosol-specific components during local summer in 2071–2100. The left column shows the difference between G6sulfur and SSP5-8.5, the quantity shown in Figs. 2–8. The middle column shows the difference between SSP2-4.5 and SSP5-8.5, the effect of removing the greenhouse gas forcing that G6sulfur is designed to offset; The right column shows the difference between G6sulfur and SSP2-4.5, the residual at matched global mean forcing. The rows show (a–c) vertical velocity at 500 hPa (positive downward), (d–f) precipitation, (g–i) surface soil moisture, (j–l) CDHW days per summer. MME differences; stippling marks grid cells where all models with a defined value agree on the sign. Boxes mark Central Africa (15°–30°E, 5°S–5°N) and an Amazon box at the same latitudes (70°–50°W, 5°S–5°N). The left column equals the sum of the other two. The global tropics are shown in Figs. S6–S7, the equatorial longitude-pressure sections of vertical velocity in Fig. S8, and the regional means in Table S1.

Over tropical land areas, the equivalent reduction in greenhouse gases primarily drives the spatial anomaly pattern of G6sulfur relative to SSP5-8.5. Across land areas between 30°S and 30°N, the MME changes in summer CDHW days under G6sulfur is spatially correlated with that under SSP2-4.5 ($r = 0.79$, regression slope 0.8), and comparable values are obtained for precipitation and for heat wave days. The reduction forcing effectively alleviates compound dry-hot conditions across



743 most tropical landmasses, particularly in regions where the SSP5-8.5 scenario projects severe
744 surface drying. The Amazon Basin serve as a prime example. Compared to SSP5-8.5, the SSP2-4.5
745 scenario increases Amazon precipitation by 0.6 mm day^{-1} , increases soil moisture, and reduces
746 summer CDHW occurrences by 7.8 days (Figs. 9e, h and k). This response consistent with the
747 projected drying of the basin under strong forcing (Cook et al., 2020). Conversely, Central Africa
748 constitutes the notable exception. Consistent with most CMIP6 models (Almazroui et al., 2020), our
749 three selected models simulate a wetting trend over the Congo Basin under SSP5-8.5. Therefore,
750 the reduction in forcing decreases precipitation by $-0.35 \text{ mm day}^{-1}$ and leaves the summer CDHW
751 frequency essentially unchanged (-1.0 days, with no individual model changing by more than 1.8
752 days; Fig. 9k). Consequently, Central Africa stands out as a unique tropical land region where
753 reduced greenhouse gas forcing fail to alleviate compound dry-hot extremes.

754 The residual response between G6sulfur and SSP2-4.5 represents a general drying of tropical
755 landmass rather than a localized anomaly confined to Central Africa (Figs. 9c, f, i and l; Figs. S6–
756 S7). Under matched global mean forcing, the stratospheric aerosol layer diminishes net surface
757 radiation by -8 W m^{-2} over Central Africa and -7 W m^{-2} over the Amazon. This energy deficit
758 consistently reduces the latent heat flux by $10\text{--}14 \text{ W m}^{-2}$ accross all three models for both regions.
759 Consequently, precipitation decreases by 0.6 mm day^{-1} over Central Africa and by 0.5 mm day^{-1}
760 over the Amazon, and the number of dry days per summer increases by 4.0 and 2.6 days, respectively.
761 These responses correspond to the well-documented hydrological signature of SRM, whereby the
762 reduction in shortwave radiation suppresses evaporation more strongly than the associated cooling
763 reduces the atmospheric moisture demand (Bala et al., 2008; Tilmes et al., 2013), expressed here in
764 the event-based metrics. The residual circulation anomaly exhibits a distinct zonally structure rather
765 than a simple latitudinal shift of the mean Hadley cells (Fig. S8). Relative to SSP2-4.5, G6sulfur
766 produces anomalous subsidence in the mid-tropospheric over Atlantic and African sectors and the
767 Maritime Continent, alongside anomalous ascent over the eastern Pacific. Conversely, the anomalies
768 between G6sulfur and SSP5-8.5 are dominated by ascent over the Amazon and the subsidence over
769 the eastern Pacific, which are attributable to the reduced greenhouse-gas forcing (Figs. S8a & S8b).
770 Over Central Africa, the residual subsidence reaches MME of $+5.4 \times 10^{-3} \text{ Pa s}^{-1}$, primarily driven



771 by the IPSL-CM6A-LR and MPI-ESM1-2-LR models. This sinking motion coincides with a 4.5%
772 decrease in cloud cover across all three models and reduced soil moisture in two of them.
773 Consequently, the aerosol induced drying is slightly stronger over Central Africa than over the
774 Amazon, although this difference is modest compared with the difference between the two regions
775 in the greenhouse gas component.

776 The enhanced CDHW response over Central African reflects the combination of a small
777 greenhouse gas component and a widespread aerosol induced drying across tropical land.
778 Specifically, combining a 1.0 day reduction from greenhouse gases with a 3.9 days increase from
779 the aerosol effect yields a net addition of 2.9 summer CDHW days over Central Africa. Conversely,
780 the Amazon experiences a net decrease of 5.3 days, as the substantial 7.8 days reduction driven by
781 greenhouse gases easily overpowers the 2.4 days aerosol penalty (Table S1). Therefore, Central
782 Africa emerges as a hotspot not due to a unique local reaction to SAI, but because the reduced
783 greenhouse gas forcing fails to provide sufficient compensating benefit to offset the regional aerosol
784 drying. Two caveats should be considered when interpreting these results. First, G6sulfur is designed
785 to achieve a global-mean radiative forcing comparable to that of SSP2-4.5, rather than to reproduce
786 its regional temperature response. Consequently, differences between G6sulfur and the reference
787 pathway can persist at regional scales. For example, summer mean Tmax is higher under G6sulfur
788 by approximately 0.8 °C over Central Africa and 0.5 °C over Amazon. This regional warming is a
789 recognized feature of SRM when compared to equivalent emission reductions (Visioni et al., 2021).
790 Additionally, relying on a single member per model inevitably introduces internal climate variability
791 between the analyzed time periods. Nevertheless, the agreement of all three models on the increases
792 in dry days, CDHW days, and CDHW duration over Central Africa strongly confirms that this
793 residual component is a genuine forced response.

794 **4 Summary and Discussion**

795 This study shows that SAI substantially reduces the global intensification of CDHWs relative
796 to SSP5-8.5, with reductions in event frequency, duration and severity over most land, of which the
797 severity signal is the most spatially coherent and the most robust to the dry-day criterion. However,
798 this mitigation is spatially heterogeneous, and Central Africa emerges as a distinct hotspot where



799 CDHW duration increases under G6sulfur, together with the total number of hot-and-dry days per
800 summer, in all three models. The accompanying increase in event frequency is less robust. It
801 weakens substantially when alternative dry day criteria are applied and is not consistently
802 reproduced across all three models. Therefore, the regional result is best stated as one about
803 persistence and total exposure rather than about event counts. Importantly, this regional increase
804 occurs despite a lower summer mean Tmax, indicating that global cooling does not necessarily
805 translate into uniform reductions in compound climate risk. The Central African response provides
806 an important perspective on how SAI may redistribute rather than uniformly reduce climate risks.
807 The increase in CDHWs is accompanied by longer and more frequent dry events and by more
808 numerous but shorter heat waves. Total heat wave days over the region in fact decrease in two of
809 the three models, so the compound signal is carried by the dry component. A comparison with SSP2-
810 4.5 further shows that this regional signal does not represent a reversal of the greenhouse gas
811 response. Reducing greenhouse gas forcing alone leaves hot-and-dry exposure over Central Africa
812 essentially unchanged, whereas the increase arises entirely from the response to SAI under matched
813 global mean forcing. This suggests that the regional response is not driven by a simple warming of
814 the mean climate, but by changes in the frequency and persistence of threshold-exceeding hot and
815 dry conditions. In particular, precipitation trends can exert a dominant control on the future
816 occurrence of compound dry-heat events (Bevacqua et al., 2022). Thus, evaluating SAI using
817 temperature-based indicators alone may overlook important regional changes in compound hazards.

818 A key question is why Central Africa exhibits a response opposite to that over Amazon Basin,
819 despite the two regions being located at same latitude band. The zonal-mean adjustment of the
820 Hadley circulation identified in Sect. 3.3 cannot account for this contrast, because a zonal-mean
821 anomaly is identical at all longitudes. The regional decomposition in Sect. 3.4 suggests that the
822 contrast arises from differences in the greenhouse gas response between the two regions. Under
823 G6sulfur, relative to SSP2-4.5 and at comparable global mean forcing, tropical land surfaces
824 generally become drier. The reduction in surface shortwave flux and evaporation associated with
825 the aerosol forcing is not fully offset by the accompanying reduction in atmospheric moisture
826 demand. As a result, both Central Africa and the Amazon experience increase in the number of dry



827 days and CDHW days during summer. However, the two regions differ primarily in the greenhouse
828 gas component. Under SSP5-8.5, drying occurs over the Amazon Basin, whereas the Congo Basin
829 becomes wetter. Consequently, the removal of the greenhouse gas induced response under G6sulfur
830 relative to SSP5-8.5 substantially reduces hot-and-dry days over the Amazon, but has a much
831 weaker compensating effect over Central Africa. The Central African hotspot therefore reflects the
832 aerosol-induced drying that is not compensated by the reduced forcing. This regional response is
833 characterized by enhanced mid-tropospheric subsidence, reduced cloud cover and precipitation,
834 lower soil moisture and latent heat flux, and a shift in the surface energy partitioning toward sensible
835 heating (Fig. 8). Together, these changes favor stronger land-atmosphere coupling under moisture
836 limited conditions, whereby reduced soil moisture suppresses evaporative cooling and enhances
837 sensible heating, thereby amplifying temperature extremes (Seneviratne et al., 2010; Miralles et al.,
838 2014). Similar dynamical controls on African precipitation under stratospheric aerosol forcing have
839 been reported previously (Haywood et al., 2013; Da-Allada et al., 2020), and the zonally asymmetric
840 structure of the residual circulation response (Fig. S8) is consistent with the sensitivity of tropical
841 precipitation to stratospheric heating and atmospheric stability (Ferraro et al., 2014; Cheng et al.,
842 2022).

843 These results highlight an important limitation of evaluating the effectiveness of SAI primarily
844 in terms of its ability to reduce global or regional mean temperature. Although SAI can effectively
845 offset greenhouse-gas-induced warming, its effects on precipitation, atmospheric circulation, and
846 land-atmosphere coupling may lead to regionally divergent responses in compound climate
847 extremes (Irvine et al., 2019; Visioni et al., 2021), and the character of these responses will depend
848 on the injection strategy employed (MacMartin et al., 2022). This issue is particularly relevant for
849 Africa, where compound dry-heat wave extremes can pose substantial risks to ecosystems,
850 agriculture, and exposed populations (Weber et al., 2020). These findings therefore suggest that
851 assessments of SAI should explicitly consider compound climate extremes and regional
852 hydroclimatic trade-offs, rather than relying solely on reductions in global-mean temperature as an
853 indicator of climate-risk mitigation.

854 Several limitations should be acknowledged. Event-based metrics are sensitive to how



continuous exceedances are grouped into discrete events. Under strong warming the tropics spend most of the summer above a threshold fixed on the 1985–2014 climatology, so exceedances merge into a few very long events and the event count falls. A cooling that fragments one long episode into several shorter ones then registers as an increase in frequency without any increase in exposure. This is exactly what happens to heat waves over Central Africa (Sect. 3.2), which is why we report total exceedance days alongside the event counts. The compound metric remains robust to this ambiguity, with CDHW days per summer increase in all three models. However, event frequency alone should not be interpreted as a measure of exposure anywhere in this analysis. Our analysis is based on three models that simulated G6sulfur, and although the principal circulation response is broadly consistent across models, differences in magnitude remain. The percentile based dry-event criterion is further sensitive to how each model represents light precipitation. In IPSL-CM6A-LR, precipitation is exactly zero on 24% of local summer days, resulting in a zero 10th percentile threshold over 63% of land area and preventing the identification of dry events in these regions. Those grid cells are excluded from the MME, and Central Africa is unaffected. More generally, the local-summer P_{10} lies below 0.1 mm day⁻¹ at 81–90% of land grid cells, so the percentile criterion discriminates among very light precipitation amounts over much of the land surface. Repeating the CDHW analysis with a fixed 1 mm day⁻¹ threshold and with a wet-day-based P_{10} (Fig. S5) yields a similar global pattern, with SAI reducing CDHW frequency, duration and severity over most land and producing the largest decrease in severity over the Amazon. However, because nearly every summer day is classified as dry over arid land, CDHWs in these areas effectively become heat waves. Consequently, the global land mean reductions in frequency and duration become larger, reaching about -1.0 events yr⁻¹ and -1.5 to -1.6 days event⁻¹, respectively, compared with -0.4 events yr⁻¹ and -0.1 days event⁻¹ under the P_{10} criterion. Over Central Africa, the MME increase in CDHW duration is insensitive to the criterion (+0.66 to +0.72 days event⁻¹ under all three definitions) and the decrease in severity is likewise preserved (-0.50 to -0.57), whereas the MME increase in frequency weakens from +0.39 to +0.06–0.09 events yr⁻¹ and is no longer shared by all three models (CNRM-ESM2-1 shows a decrease under both alternative criteria). The Central African enhancement of CDHW persistence is therefore robust to the dry-day definition, while the frequency



883 component of the regional signal should be regarded as criterion-dependent. Moreover, G6sulfur
884 represents a tropical, idealized, and sustained SAI strategy, and its regional effects should not be
885 generalized to all possible injection strategies. Relatedly, the SSP2-4.5 comparison in Sect. 3.4
886 separates the effect of the reduced greenhouse-gas forcing from the response to SAI at matched
887 forcing, but the latter still combines the spatial pattern of the cooling with aerosol-specific effects
888 such as stratospheric heating and an increased diffuse radiation fraction. The present design
889 therefore cannot determine whether the residual drying would arise under any form of solar radiation
890 modification or is specific to the aerosol layer. A parallel analysis of the G6solar experiment, in
891 which the same forcing target is met by reducing solar irradiance instead, would be required to
892 separate these contributions. Finally, the circulation-drying-land-surface pathway identified here
893 represents a physically consistent interpretation rather than a formal causal attribution. Larger multi-
894 model ensembles and targeted sensitivity experiments will be required to quantify the relative
895 contributions of circulation changes and land-atmosphere feedbacks.



896 **Data available**

897 All model data used in this work are available from the Earth System Grid Federation (WCRP,
898 2026; <https://metagrid.esgf-west.org/>, last access: 10 Sep 2026).

899 **Author contributions**

900 C. Qi performed the analyses and wrote the initial draft. Y. Gao conceived and supervised the
901 study. All authors reviewed and edited the initial draft, discussed the results, and contributed to the
902 final manuscript.

903 **Competing interests**

904 The authors declare that they have no competing interest.

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912 **References**

- 913 Almazroui, M., Saeed, F., Saeed, S., Nazrul Islam, M., Ismail, M., Klutse, N. A. B., & Siddiqui, M.
914 H. (2020). Projected change in temperature and precipitation over Africa from CMIP6. *Earth*
915 *Systems and Environment*, 4, 455–475. <https://doi.org/10.1007/s41748-020-00161-x>
- 916 Bala, G., Duffy, P. B., & Taylor, K. E. (2008). Impact of geoengineering schemes on the global
917 hydrological cycle. *Proceedings of the National Academy of Sciences*, 105(22), 7664–7669.
918 <https://doi.org/10.1073/pnas.0711648105>
- 919 Barichivich, J., Osborn, T., Harris, I., van der Schrier, G., & Jones, P. (2019). Drought: Monitoring
920 global drought using the self-calibrating Palmer Drought Severity Index. *Bulletin of the*
921 *American Meteorological Society*, 100(9), S39–S40.
922 <https://doi.org/10.1175/2019BAMSStateoftheClimate.1>
- 923 Barriopedro, D., Fischer, E. M., Luterbacher, J., Trigo, R. M., & García-Herrera, R. (2011). The hot
924 summer of 2010: redrawing the temperature record map of Europe. *Science*, 332(6026), 220–
925 224. <https://doi.org/10.1126/science.1201224>
- 926 Bevacqua, E., Zappa, G., Lehner, F. & Zscheischler, J. (2022). Precipitation trends determine future
927 occurrences of compound hot–dry events. *Nat. Clim. Chang.* 12, 350–355.
928 <https://doi.org/10.1038/s41558-022-01309-5>
- 929 Brás, T. A., Seixas, J., Carvalhais, N., & Jägermeyr, J. (2021). Severity of drought and heatwave
930 crop losses tripled over the last five decades in Europe. *Environmental Research Letters*, 16(6),
931 065012. <https://doi.org/10.1088/1748-9326/abf004>
- 932 Cheng, W., MacMartin, D. G., Kravitz, B., Vioni, D., Bednarz, E. M., Xu, Y., ... & Deng, X. (2022).
933 Changes in Hadley circulation and intertropical convergence zone under strategic stratospheric
934 aerosol geoengineering. *npj Climate and Atmospheric Science*, 5(1), 32.
935 <https://doi.org/10.1038/s41612-022-00254-6>.
- 936 Clarke, B., Otto, F., Stuart-Smith, R., & Harrington, L. (2022). Extreme weather impacts of climate
937 change: an attribution perspective. *Environmental Research: Climate*, 1(1), 012001.
938 <https://doi.org/10.1088/2752-5295/ac6e7d>
- 939 Cook, B. I., Mankin, J. S., Marvel, K., Williams, A. P., Smerdon, J. E., & Anchukaitis, K. J. (2020).



- 940 Twenty-first century drought projections in the CMIP6 forcing scenarios. *Earth's Future*, 8(6),
941 e2019EF001461. <https://doi.org/10.1029/2019EF001461>
- 942 Crutzen, P. J. (2006). Albedo enhancement by stratospheric sulfur injections: a contribution to
943 resolve a policy dilemma?. *Climatic change*, 77(3-4), 211. [https://doi.org/10.1007/s10584-](https://doi.org/10.1007/s10584-006-9101-y)
944 006-9101-y
- 945 Curry, C. L., Sillmann, J., Bronaugh, D., Alterskjaer, K., Cole, J. N. S., Ji, D., Kravitz, B.,
946 Kristjánsson, J. E., Moore, J. C., Muri, H., Niemeier, U., Robock, A., Tilmes, S., & Yang, S.
947 (2014). A multimodel examination of climate extremes in an idealized geoengineering
948 experiment. *Journal of Geophysical Research: Atmospheres*, 119, 3900–3923.
949 <https://doi.org/10.1002/2013JD020648>
- 950 Da-Allada, C. Y., Baloitcha, E., Alamou, E. A., Awo, F. M., Bonou, F., Pomalegni, Y., ... & Irvine,
951 P. J. (2020). Changes in west African summer monsoon precipitation under stratospheric
952 aerosol geoengineering. *Earth's Future*, 8, e2020EF001595.
953 <https://doi.org/10.1029/2020EF001595>.
- 954 Dai, A. (2013). Increasing drought under global warming in observations and models. *Nature*
955 *climate change*, 3(1), 52-58. <https://doi.org/10.1038/nclimate1633>
- 956 Feng, S., Hao, Z., Zhang, X., & Hao, F. (2019). Probabilistic evaluation of the impact of compound
957 dry-hot events on global maize yields. *Science of the total environment*, 689, 1228-1234.
958 <https://doi.org/10.1016/j.scitotenv.2019.06.373>
- 959 Ferraro, A. J., Highwood, E. J., & Charlton-Perez, A. J. (2014). Weakened tropical circulation and
960 reduced precipitation in response to geoengineering. *Environmental Research Letters*, 9(1),
961 014001. <https://doi.org/10.1088/1748-9326/9/1/014001>
- 962 Fischer, E. M., & Knutti, R. (2015). Anthropogenic contribution to global occurrence of heavy-
963 precipitation and high-temperature extremes. *Nature climate change*, 5(6), 560-564.
964 <https://doi.org/10.1038/nclimate2617>
- 965 Fischer, E. M., Sippel, S., & Knutti, R. (2021). Increasing probability of record-shattering climate
966 extremes. *Nature Climate Change*, 11(8), 689-695. [https://doi.org/10.1038/s41558-021-01092-](https://doi.org/10.1038/s41558-021-01092-9)
967 9



- 968 Hao, Y., Hao, Z., Feng, S., Zhang, X., & Hao, F. (2020). Response of vegetation to El Niño-Southern
969 Oscillation (ENSO) via compound dry and hot events in southern Africa. *Global and Planetary*
970 *Change*, 195, 103358. <https://doi.org/10.1016/j.gloplacha.2020.103358>
- 971 Hao, Z., Hao, F., Singh, V. P., & Zhang, X. (2018). Changes in the severity of compound drought
972 and hot extremes over global land areas. *Environmental Research Letters*, 13(12), 124022.
973 <https://doi.org/10.1088/1748-9326/aace96>
- 974 Hao, Z., Hao, F., Xia, Y., Singh, V. P., & Zhang, X. (2019). A monitoring and prediction system for
975 compound dry and hot events. *Environmental Research Letters*, 14(11), 114034.
976 <https://doi.org/10.1088/1748-9326/ab4df5>
- 977 Haywood, J. M., Jones, A., Bellouin, N., & Stephenson, D. (2013). Asymmetric forcing from
978 stratospheric aerosols impacts Sahelian rainfall. *Nature climate change*, 3(7), 660-665.
979 <https://doi.org/10.1038/nclimate1857>
- 980 He, G., Lin, Y., Hu, J., Chen, Y., Guo, Y., Yu, M., ... & Ma, W. (2024). The trends of non-accidental
981 mortality burden attributed to compound hot-dry events in China and its provinces in a global
982 warming world. *Environment International*, 191, 108977.
983 <https://doi.org/10.1016/j.envint.2024.108977>
- 984 Irvine, P., Emanuel, K., He, J., Horowitz, L. W., Vecchi, G., & Keith, D. (2019). Halving warming
985 with idealized solar geoengineering moderates key climate hazards. *Nature climate change*,
986 9(4), 295-299. <https://doi.org/10.1038/s41558-019-0398-8>
- 987 Jones, A., Haywood, J. M., Alterskjær, K., Boucher, O., Cole, J. N., Curry, C. L., ... & Yoon, J. H.
988 (2013). The impact of abrupt suspension of solar radiation management (termination effect) in
989 experiment G2 of the Geoengineering Model Intercomparison Project (GeoMIP). *Journal of*
990 *Geophysical Research: Atmospheres*, 118(17), 9743-9752. <https://doi.org/10.1002/jgrd.50762>
- 991 Keith, D. W., Weisenstein, D. K., Dykema, J. A., & Keutsch, F. N. (2016). Stratospheric solar
992 geoengineering without ozone loss. *Proceedings of the National Academy of Sciences*, 113(52),
993 14910-14914. <https://doi.org/10.1073/pnas.1615572113>
- 994 Kravitz, B., Caldeira, K., Boucher, O., Robock, A., Rasch, P. J., Alterskjaer, K., ... & Yoon, J. H.
995 (2013). Climate model response from the geoengineering model intercomparison project



- 996 (GeoMIP). *Journal of Geophysical Research: Atmospheres*, 118(15), 8320-8332.
997 <https://doi.org/10.1002/jgrd.50646>
- 998 Kravitz, B., MacMartin, D. G., Tilmes, S., Richter, J. H., Mills, M. J., Cheng, W., ... & Vitt, F. (2019).
999 Comparing surface and stratospheric impacts of geoengineering with different SO₂ injection
1000 strategies. *Journal of Geophysical Research: Atmospheres*, 124(14), 7900-7918.
1001 <https://doi.org/10.1029/2019JD030329>
- 1002 Kravitz, B., Robock, A., Boucher, O., Schmidt, H., Taylor, K. E., Stenchikov, G., & Schulz, M.
1003 (2011). The Geoengineering Model Intercomparison Project (GeoMIP). *Atmospheric Science*
1004 *Letters*, 12, 162–167. <https://doi.org/10.1002/asl.316>
- 1005 Kravitz, B., Robock, A., Tilmes, S., Boucher, O., English, J. M., Irvine, P. J., ... & Watanabe, S.
1006 (2015). The Geoengineering Model Intercomparison Project Phase 6 (GeoMIP6): simulation
1007 design and preliminary results, *Geosci. Model Dev.*, 8, 3379–3392.
1008 <https://doi.org/10.5194/gmd-8-3379-2015>
- 1009 Kravitz, B., MacMartin, D. G., Mills, M. J., Richter, J. H., Tilmes, S., Lamarque, J.-F., Tribbia, J. J.,
1010 & Vitt, F. (2017). First simulations of designing stratospheric sulfate aerosol geoengineering
1011 to meet multiple simultaneous climate objectives. *Journal of Geophysical Research:*
1012 *Atmospheres*, 122(23), 12616–12634. <https://doi.org/10.1002/2017JD026874>
- 1013 Lionello, P., D'Agostino, R., Ferreira, D., Nguyen, H., & Singh, M. S. (2024). The Hadley
1014 circulation in a changing climate. *Annals of the New York Academy of Sciences*, 1534, 69–93.
1015 <https://doi.org/10.1111/nyas.15114>
- 1016 Lorenz, R., Jaeger, E. B., & Seneviratne, S. I. (2010). Persistence of heat waves and its link to soil
1017 moisture memory. *Geophysical Research Letters*, 37(9).
1018 <https://doi.org/10.1029/2010GL042764>
- 1019 Lu, J., Vecchi, G. A., & Reichler, T. (2007). Expansion of the Hadley cell under global warming.
1020 *Geophysical Research Letters*, 34(6). <https://doi.org/10.1029/2006GL028443>
- 1021 MacMartin, D. G., Ricke, K. L., Keith, D. W., Mitchell, D., Allen, M. R., Hall, J. W., ... & Quéré,
1022 C. L. (2018). Solar geoengineering as part of an overall strategy for meeting the 1.5 C Paris
1023 target. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and*



- 1024 Engineering Sciences, 376(2119). <https://doi.org/10.1098/rsta.2016.0454>
- 1025 MacMartin, D. G., Visioni, D., Kravitz, B., Richter, J. H., Felgenhauer, T., Lee, W. R., Morrow, D.
1026 R., Parson, E. A., & Sugiyama, M. (2022). Scenarios for modeling solar radiation modification.
1027 Proceedings of the National Academy of Sciences, 119(33), e2202230119.
1028 <https://doi.org/10.1073/pnas.2202230119>
- 1029 Miralles, D., Teuling, A., van Heerwaarden, C. & Vilà-Guerau de Arellano, J. (2014). Mega-
1030 heatwave temperatures due to combined soil desiccation and atmospheric heat accumulation.
1031 Nature Geoscience, 7, 345–349. <https://doi.org/10.1038/ngeo2141>.
- 1032 Niemeier, U., Richter, J. H., & Tilmes, S. (2020). Differing responses of the quasi-biennial
1033 oscillation to artificial SO₂ injections in two global models. Atmospheric Chemistry and
1034 Physics, 20(14), 8975–8987. <https://doi.org/10.5194/acp-20-8975-2020>
- 1035 Niemeier, U., & Schmidt, H. (2017). Changing transport processes in the stratosphere by radiative
1036 heating of sulfate aerosols. Atmospheric Chemistry and Physics, 17(24), 14871–14886.
1037 <https://doi.org/10.5194/acp-17-14871-2017>
- 1038 O'Neill, B. C., Tebaldi, C., Van Vuuren, D. P., Eyring, V., Friedlingstein, P., Hurtt, G., ... & Sanderson,
1039 B. M. (2016). The scenario model intercomparison project (ScenarioMIP) for CMIP6.
1040 Geoscientific Model Development, 9(9), 3461–3482. [https://doi.org/10.5194/gmd-9-3461-](https://doi.org/10.5194/gmd-9-3461-2016)
1041 2016
- 1042 Oort, A. H., & Rasmusson, E. M. (1970). On the annual variation of the monthly mean meridional
1043 circulation. Monthly Weather Review, 98(6), 423–442. [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0493(1970)098<0423:OTAVOT>2.3.CO;2)
1044 0493(1970)098<0423:OTAVOT>2.3.CO;2
- 1045 Oort, A. H., & Yienger, J. J. (1996). Observed interannual variability in the Hadley circulation and
1046 its connection to ENSO. Journal of Climate, 9(11), 2751–2767. [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0442(1996)009<2751:OIVITH>2.0.CO;2)
1047 0442(1996)009<2751:OIVITH>2.0.CO;2
- 1048 Perkins-Kirkpatrick, S. E., & Lewis, S. C. (2020). Increasing trends in regional heatwaves. Nature
1049 communications, 11(1), 3357. <https://doi.org/10.1038/s41467-020-16970-7>
- 1050 Perkins, S. E. (2015). A review on the scientific understanding of heatwaves—Their measurement,
1051 driving mechanisms, and changes at the global scale. Atmospheric Research, 164, 242–267.



- 1052 <https://doi.org/10.1016/j.atmosres.2015.05.014>
- 1053 Perkins, S. E., and L. V. Alexander. (2013). On the Measurement of Heat Waves. *Journal of Climate*,
- 1054 26, 4500–4517, <https://doi.org/10.1175/JCLI-D-12-00383.1>
- 1055 Philip, S. Y., Kew, S. F., van Oldenborgh, G. J., Anslow, F. S., Seneviratne, S. I., Vautard, R., et al.
- 1056 (2022). Rapid attribution analysis of the extraordinary heat wave on the Pacific coast of the US
- 1057 and Canada in June 2021. *Earth System Dynamics*, 13(4), 1689-1713.
- 1058 <https://doi.org/10.5194/esd-13-1689-2022>
- 1059 Pitari, G., Aquila, V., Kravitz, B., Robock, A., Watanabe, S., Cionni, I., ... & Tilmes, S. (2014).
- 1060 Stratospheric ozone response to sulfate geoengineering: Results from the Geoengineering
- 1061 Model Intercomparison Project (GeoMIP). *Journal of Geophysical Research: Atmospheres*,
- 1062 119(5), 2629-2653. <https://doi.org/10.1002/2013JD020566>
- 1063 Rahman, M. R., & Lateh, H. (2016). Meteorological drought in Bangladesh: assessing, analysing
- 1064 and hazard mapping using SPI, GIS and monthly rainfall data. *Environmental Earth Sciences*,
- 1065 75(12), 1026. <https://doi.org/10.1007/s12665-016-5829-5>
- 1066 Ribeiro, A. F. S., Russo, A., Gouveia, C. M., Páscoa, P., & Zscheischler, J. (2020). Risk of crop
- 1067 failure due to compound dry and hot extremes estimated with nested copulas. *Biogeosciences*
- 1068 Discussions, 2020, 1-21. <https://doi.org/10.5194/bg-17-4815-2020>
- 1069 Richter, J. H., Visioni, D., MacMartin, D. G., Bailey, D. A., Rosenbloom, N., Dobbins, B., Lee, W.
- 1070 R., Tye, M., & Lamarque, J.-F. (2022). Assessing Responses and Impacts of Solar climate
- 1071 intervention on the Earth system with stratospheric aerosol injection (ARISE-SAI): protocol
- 1072 and initial results from the first simulations. *Geoscientific Model Development*, 15(22), 8221–
- 1073 8243. <https://doi.org/10.5194/gmd-15-8221-2022>
- 1074 Robock, A. (2000). Volcanic eruptions and climate. *Rev. Geophys.* 38, 191–219.
- 1075 <https://doi.org/10.1029/1998RG000054>
- 1076 Robock, A., Oman, L., & Stenchikov, G. L. (2008). Regional climate responses to geoengineering
- 1077 with tropical and Arctic SO₂ injections. *Journal of Geophysical Research: Atmospheres*,
- 1078 113(D16). <https://doi.org/10.1029/2008JD010050>
- 1079 Seneviratne, S. I., Corti, T., Davin, E. L., Hirschi, M., Jaeger, E. B., Lehner, I., ... & Teuling, A. J.



- 1080 (2010). Investigating soil moisture–climate interactions in a changing climate: A review. Earth-
1081 Science Reviews, 99(3-4), 125-161. <https://doi.org/10.1016/j.earscirev.2010.02.004>
- 1082 Seneviratne, S. I., Zhang, X., Adnan, M., Badi, W., Dereczynski, C., Luca, A. D., ... & Allan, R.
1083 (2021). Weather and climate extreme events in a changing climate. Climate change 2021: The
1084 physical science basis: Working group I contribution to the sixth assessment report of the
1085 intergovernmental panel on climate change, 1513-1766.
1086 <https://doi.org/10.1017/9781009157896.013>
- 1087 Simpson, I. R., Tilmes, S., Richter, J. H., Kravitz, B., MacMartin, D. G., Mills, M. J., ... &
1088 Pendergrass, A. G. (2019). The regional hydroclimate response to stratospheric sulfate
1089 geoengineering and the role of stratospheric heating. Journal of Geophysical Research:
1090 Atmospheres, 124(23), 12587-12616. <https://doi.org/10.1029/2019JD031093>
- 1091 Staten, P. W., Lu, J., Grise, K. M., Davis, S. M., & Birner, T. (2018). Re-examining tropical
1092 expansion. Nature Climate Change, 8(9), 768-775. [https://doi.org/10.1038/s41558-018-0246-](https://doi.org/10.1038/s41558-018-0246-2)
1093 2
- 1094 Stott, P. A., Stone, D. A., & Allen, M. R. (2004). Human contribution to the European heatwave of
1095 2003. Nature, 432(7017), 610-614. <https://doi.org/10.1038/nature03089>
- 1096 Tebaldi, C., Debeire, K., Eyring, V., Fischer, E., Fyfe, J., Friedlingstein, P., ... & Ziehn, T. (2021).
1097 Climate model projections from the scenario model intercomparison project (ScenarioMIP) of
1098 CMIP6. Earth System Dynamics, 12(1), 253-293. <https://doi.org/10.5194/esd-12-253-2021>
- 1099 Tilmes, S., Fasullo, J., Lamarque, J. F., Marsh, D. R., Mills, M., Alterskjaer, K., ... & Watanabe, S.
1100 (2013). The hydrological impact of geoengineering in the Geoengineering Model
1101 Intercomparison Project (GeoMIP). Journal of Geophysical Research: Atmospheres, 118(19),
1102 11-036. <https://doi.org/10.1002/jgrd.50868>
- 1103 Tilmes, S., MacMartin, D. G., Lenaerts, J. T. M., van Kampenhout, L., Muntjewerf, L., ... & Robock,
1104 A. (2020). Reaching 1.5 and 2.0 °C global surface temperature targets using stratospheric
1105 aerosol geoengineering, Earth System Dynamics, 11, 579–601, [https://doi.org/10.5194/esd-11-](https://doi.org/10.5194/esd-11-579-2020)
1106 579-2020.
- 1107 Tilmes, S., Mills, M. J., Niemeier, U., Schmidt, H., Robock, A., Kravitz, B., ... & English, J. M.



- 1108 (2015). A new Geoengineering Model Intercomparison Project (GeoMIP) experiment designed
1109 for climate and chemistry models. *Geoscientific Model Development*, 8(1), 43-49.
1110 <https://doi.org/10.5194/gmd-8-43-2015>
- 1111 Tilmes, S., Richter, J. H., Kravitz, B., MacMartin, D. G., Mills, M. J., Simpson, I. R., ... & Ghosh,
1112 S. (2018). CESM1 (WACCM) stratospheric aerosol geoengineering large ensemble project.
1113 *Bulletin of the American Meteorological Society*, 99(11), 2361-2371.
1114 <https://doi.org/10.1175/BAMS-D-17-0267.1>
- 1115 Visioni, D., Kravitz, B., Robock, A., Tilmes, S., Haywood, J. M., Boucher, O., ... & Muri, H. (2022).
1116 Opinion: The scientific and community-building roles of the Geoengineering Model
1117 Intercomparison Project (GeoMIP)-past, present, and future. *Atmospheric chemistry and*
1118 *physics Discussions*, 2022, 1-44. <https://doi.org/10.5194/acp-23-5149-2023>
- 1119 Visioni, D., MacMartin, D. G., Kravitz, B., Boucher, O., Jones, A., Lurton, T., ... & Tilmes, S. (2021).
1120 Identifying the sources of uncertainty in climate model simulations of solar radiation
1121 modification with the G6sulfur and G6solar Geoengineering Model Intercomparison Project
1122 (GeoMIP) simulations. *Atmospheric Chemistry and Physics*, 21(13), 10039-10063.
1123 <https://doi.org/10.5194/acp-21-10039-2021>
- 1124 Von Buttler, J., Zscheischler, J., Rammig, A., Sippel, S., Reichstein, M., Knohl, A., ... & Mahecha,
1125 M. D. (2018). Impacts of droughts and extreme-temperature events on gross primary
1126 production and ecosystem respiration: a systematic assessment across ecosystems and climate
1127 zones. *Biogeosciences*, 15(5), 1293-1318. <https://doi.org/10.5194/bg-15-1293-2018>
- 1128 Wang, P., Tang, J., Sun, X., Wang, S., Wu, J., Dong, X., & Fang, J. (2017). Heat waves in China:
1129 Definitions, leading patterns and connections to large-scale atmospheric circulation and SSTs.
1130 *Journal of Geophysical Research: Atmospheres*, 122, 10,679–10,699.
1131 <https://doi.org/10.1002/2017JD027180>.
- 1132 Weber, T., Bowyer, P., Rechid, D., Pfeifer, S., Raffaele, F., Remedio, A. R., ... & Jacob, D. (2020).
1133 Analysis of compound climate extremes and exposed population in Africa under two different
1134 emission scenarios. *Earth's Future*, 8(9), e2019EF001473.
1135 <https://doi.org/10.1029/2019EF001473>



- 1136 Wells, A. F., Henry, M., Bednarz, E. M., MacMartin, D. G., Jones, A., Dalvi, M., & Haywood, J. M.
1137 (2024). Identifying climate impacts from different stratospheric aerosol injection strategies in
1138 UKESM1. *Earth's Future*, 12(3), e2023EF004358. <https://doi.org/10.1029/2023EF004358>
- 1139 Wu, X., Hao, Z., Zhang, X., Li, C., & Hao, F. (2020). Evaluation of severity changes of compound
1140 dry and hot events in China based on a multivariate multi-index approach. *Journal of*
1141 *Hydrology*, 583, 124580. <https://doi.org/10.1016/j.jhydrol.2020.124580>
- 1142 Zscheischler, J., Martius, O., Westra, S., Bevacqua, E., Raymond, C., Horton, R. M., ... & Vignotto,
1143 E. (2020). A typology of compound weather and climate events. *Nature reviews earth &*
1144 *environment*, 1(7), 333-347. <https://doi.org/10.1038/s43017-020-0060-z>
- 1145 Zscheischler, Sonia I. Seneviratne. (2017). Dependence of drivers affects risks associated with
1146 compound events. *Science Advances*. 3, e1700263. <https://doi.org/10.1126/sciadv.1700263>.
- 1147 Zscheischler, J., Westra, S., Van Den Hurk, B. J., Seneviratne, S. I., Ward, P. J., Pitman, A., ... &
1148 Zhang, X. (2018). Future climate risk from compound events. *Nature climate change*, 8(6),
1149 469-477. <https://doi.org/10.1038/s41558-018-0156-3>