



Impacts and limitations of avoiding temperature overshoot with stratospheric aerosol injections

Jerry F. Tjiputra¹, Dirk Olivie², Jörg Schwinger¹, Nadine Goris¹, Norman J. Steinert³, Rosie Fisher³, and Huiji Lee⁴

¹NORCE Research AS, Bjerknes Centre for Climate Research, Bergen, Norway

²Norwegian Meteorological Institute, Oslo, Norway

³CICERO Center for International Climate Research, Oslo, Norway

⁴School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea

Correspondence: Jerry F. Tjiputra (jetj@norce-research.no)

Abstract. Recent trends in CO₂ emissions and pledged emission reductions suggest that achieving the Paris Agreement without an ‘overshoot’ represents a great challenge. In such a pathway, substantial net-negative emissions via Carbon Dioxide Removal (CDR) are required, to reduce atmospheric CO₂ and bring global mean temperatures back to the target. While the feasibility of globally net-negative emissions is under intense discussion, Stratospheric Aerosol Injection (SAI) has been proposed as a

5 feasible approach to temporarily offset anthropogenically-induced global warming, often framed as a temporary measure to both avoid exceedance of climatic tipping points and to reduce climate impacts on humanity. The complexities of the coupled response of the Earth system to CDR and SAI remain under-explored. Here, we apply a fully-interactive Earth system model to perform idealized temperature overshoot simulations, with prescribed emissions pathways and SAI deployment in three different regions: the tropics, Northern Hemisphere mid-latitudes, and Southern Hemisphere mid-latitudes. All SAI deployments

10 limit mean global warming to below two degrees, though the tropical SAI deployment yields the strongest global cooling effect due to longer aerosol lifetime and enhanced radiative effectiveness. The climatic consequences of SAI depend strongly on deployment location, primarily via the evolution of the Atlantic Meridional Overturning Circulation (AMOC). While all experiments exhibit an initial CO₂-driven AMOC weakening, Northern Hemisphere SAI most effectively mitigates this decline by reducing northern high-latitude warming and preserving North Atlantic deep-water formation. Southern Hemisphere

15 SAI induces the least influence on Northern Hemisphere high-latitude warming and therefore does not substantially mitigate the AMOC decline. These contrasting AMOC responses propagate throughout the Earth system, influencing Northern Hemisphere surface temperature, sea-ice evolution, tropical precipitation patterns, ocean ventilation, carbon uptake, and regional ocean acidification. Although many aspects of the surface climate converge after SAI termination, substantial legacy effects persist in slow components of the climate system for centuries, such as sea-level rise and permafrost. Our results demonstrate

20 that SAI deployment strategies should not be evaluated solely by their ability to regulate global temperature; furthermore, the geographical distribution of SAI strongly influences regional climate trajectories and the long-term legacy of this intervention.



1 Introduction

More than a decade after the Paris Agreement was ratified, CO₂ emissions from human activities continue to rise (Friedlingstein et al., 2025), despite progress on renewable energy development, improved energy efficiency, and some limited deployments of conventional carbon dioxide removal (CDR) mostly through afforestation/reforestation and forest management (Nemet et al., 2026). The latest Emissions Gap Report from the United Nations Environment Programme indicates that we will likely pass 1.5° warming within the next decade, and are on track for 2.8° warming by the end of this century when considering current pledges and policies (United Nations Environment Programme, 2025). Climate change has already had profound impacts on society, e.g. through increased intensity and frequency of extreme events, rising sea-level, rapid loss of Arctic sea-ice, and ocean acidification (IPCC, 2022). Exceeding the Paris Agreement temperature thresholds would further increase the negative climate impacts and the risk of triggering cascading effects that are irreversible over long time scales (McKay et al., 2022; Richardson et al., 2023; Ritchie et al., 2026). Adapting to and managing such disruptive changes would likely be very challenging for humanity (e.g. Matthews et al., 2025), which could increase the willingness to invest into large scale CDR methods to try to push the Earth system towards a lower and safer global temperature level. This temporary exceedance of a certain global temperature level followed by a decline is often referred to as temperature overshoot and this logic is pervasive through the higher ambition scenarios to be used in CMIP7 (Coupled Model Intercomparison Project phase 7 Van Vuuren et al., 2026).

Standard overshoot frameworks and climate-economic models assume that the least-cost solution to meeting temperature targets requires massive investments in large-scale CDR (Riahi et al., 2017). Whether such large-scale CDR is financially, physically or technically plausible is a question under considerable debate (Smith et al., 2016; Bednar et al., 2019; Anderegg et al., 2020; Smith et al., 2023; Deprez et al., 2024; Kazlou et al., 2024). Schleussner et al. (2024) argue that there is problematic "over-confidence in overshoot", and that we cannot be confident that temperature declines after overshoot are achievable within the timescales expected today.

Nonetheless, in recent years, a growing number of studies have investigated the impacts of temperature overshoots in model simulations that reverse (or partly reverse) the anthropogenic climate forcing (e.g. Boucher et al., 2012; Tokarska and Zickfeld, 2015; Jeltsch-Thömmes et al., 2020; Schwinger et al., 2022c; Meyer and Trisos, 2023). In general, these studies indicate that components of the climate system with low inertia follow the reversal of the radiative forcing with relatively little delay, but that an overshoot induces long-term changes in key components of the climate system, for example, the Atlantic Meridional Overturning Circulation (AMOC), sea level rise, permafrost thawing, or ice sheet collapse. This, in turn, may trigger cascading irreversible negative consequences on ecosystems and biodiversity (Santana-Falcón et al., 2023). The extend of these impacts will largely depend on two factors: the magnitude (the amount of warming above a certain temperature threshold) and the duration (the length of the period with temperatures above this threshold) of overshoot. Evidence suggests that limiting these factors to a minimum would be most beneficial (e.g. Drouet et al., 2021; Schwinger et al., 2022c; Steinert et al., 2026; Ritchie et al., 2026; Dickau et al., 2026).

Avoiding temperature overshoot requires rapid reductions in greenhouse gas emissions together with the deployment of CDR technologies to reduce atmospheric CO₂ concentrations. However, given the continued rise of emissions, additional ap-



proaches have been proposed to temporarily limit warming during an overshoot period. One approach that has been researched intensively is solar radiation modification (SRM), which aims at modifying the Earth's radiative balance to mitigate greenhouse gas-induced warming (Crutzen, 2006; Lawrence et al., 2018). One proposed method is stratospheric aerosol injection (SAI) - drawing an analog from volcanic eruptions - which injects aerosols in the lower stratosphere to reflect more incoming shortwave radiation back to space, resulting in a net cooling (Robock et al., 2009).

Modeling studies have indicated that, in the case of insufficient efforts to reduce CO₂ emissions, SAI could be applied to offset parts of the excess anthropogenic warming and mitigate some of the related dangerous climate impacts (Niemeier and Tilmes, 2017; Muri et al., 2018). However, there are also many side effects of SAI on various aspect of the Earth's climate and ecosystem, many of which create regional disparity (e.g., Tjiputra et al., 2016; Park et al., 2019; Fan et al., 2021; Müller et al., 2024), and our understanding of the impacts of SAI remains incomplete (Roberts et al., 2026).

We note that SAI deployment is a deeply contested subject. Critics argue that the prospect of SAI deployment risks deterring investment in emissions reduction (Tsipiras and Grant, 2022; Shayegh, 2025), but empirical evidence for this moral hazard effect is nuanced (Cherry et al., 2021; Andrews et al., 2022). Biermann et al. (2022) argue that SRM is not governable in a globally inclusive and just manner, and that it should be subject to non-use agreements, and indeed argue that SRM research itself risks normalizing SRM as a viable substitute for mitigation. Conversely, others highlight that SRM could help avoid the exceedance of critical climate tipping points (Smith et al., 2024; Futerman et al., 2025; Kim et al., 2025; Zhao et al., 2025; Parry et al., 2026), and dampen economic inequality and climate impacts, especially in the global South (Irvine et al., 2019; Harding et al., 2020). Futerman et al. (2025) emphasize that literature quantifying these effects remains extremely sparse.

To improve the knowledge base available for humanity to assess decision-making around SAI deployment, we investigate the use of SAI as a temporary measure to limit temperature overshoot in an idealized overshoot pathway that combines CDR with SAI in an emissions-driven Earth system model framework. The objective of our study is to (i) determine the feasibility of applying SAI to "shave-off" a temperature overshoot, (ii) provide a non-exhaustive overview of associated Earth system responses during the SAI implementation, and (iii) investigate the long-term responses and legacies of SAI. Acknowledging previous studies on the importance of SAI deployment regions (MacMartin et al., 2017; Kravitz et al., 2019), we assess the sensitivity of the climatic responses to three SAI deployment locations: the tropics and northern and southern hemispheres' mid latitudes.

2 Methods

2.1 Model description

We use the Norwegian Earth System Model version 2 (NorESM2-LM, Seland et al., 2020a; Tjiputra et al., 2020), which participated in the Coupled Model Intercomparison Project phase 6 (CMIP6, Eyring et al., 2016). It is based on the Community Earth System Model (CESM2.1, Danabasoglu et al., 2020), but uses a different ocean component, the Bergen Layered Ocean Model (BLOM) coupled with the isopycnic coordinate Hamburg Ocean Carbon Cycle (iHAMOCC) model and a different interactive atmospheric aerosol scheme (OsloAero6, Kirkevåg et al., 2018). Additional modifications and parameterization



choices are documented in Seland et al. (2020a). In this study, we use the fully-interactive carbon cycle (emission-driven) configuration of the NorESM2-LM, which prognostically simulates atmospheric CO₂ concentrations taking into account CO₂ emissions into the atmosphere and carbon fluxes from the ocean and land components.

To be able to simulate the effect of SAI, we have modified the aerosol scheme of NorESM2-LM for this study as outlined further below after the description of the standard aerosol scheme. NorESM2-LM contains an interactive aerosol scheme aimed to describe the fate/evolution of aerosols in the atmosphere, and their interactions with radiation and clouds (Kirkevåg et al., 2018). The aerosol scheme models black carbon (BC), primary organic matter (POM), secondary organic aerosol (SOA), sulphate, mineral dust and seasalt, and is able to simulate both primary and secondary aerosol (SOA and sulphate), by explicitly modelling the emission, nucleation, coagulation, activation, dry deposition and below-cloud scavenging. For the formation of secondary aerosols, oxidants such as OH, HO₂, NO₃ radical and ozone (O₃) are prescribed as three-dimensional climatologies obtained from a model with more explicit atmospheric chemistry. For sulfur emissions, we assume that 97.5 % is emitted as SO₂ (gas phase) and the remaining 2.5 % as SO₄ (aerosol). This reflects the fact that high concentrations of SO₂ just after the combustion of (fossil) fuels lead to formation of aerosol – a process happening on temporal- and spatial scales not represented by global atmospheric models

In the stratosphere, the most important aerosol is sulphate. It is mainly formed from SO₂ that is emitted by strong volcanic eruptions reaching into the stratosphere, and from the oxidation of carbonyl sulfide (OCS). OCS has its main sources in the ocean. In the standard version of NorESM2-LM, the stratospheric aerosol is not calculated interactively, but prescribed. The model takes into account the effect of a stratospheric aerosol layer by prescribing the extinction, single scattering albedo and asymmetry of the stratospheric aerosol. The distribution and values of these optical properties are often based on model simulations where the stratospheric aerosol is calculated interactively. NorESM2-LM has 16 longwave and 14 shortwave radiation bands, and the values of these optical parameters are wave-length dependent.

In the first studies of SAI performed with NorESM, SAI was implemented by, similarly to volcanic aerosol, prescribing the optical properties or mass concentration of a stratospheric aerosol layer, which is often generated from other models (Tjiputra et al., 2016; Muri et al., 2018). This limits the SAI scenarios or sensitivity simulations that can be investigated. Here, in order to investigate various SAI scenarios (different strength, geographical location, and height), we describe the formation and evolution of stratospheric aerosol from SAI more explicitly. This is done by emitting SO₂ explicitly in the stratosphere, where it can react with oxidants to form sulfuric acid (H₂SO₄). H₂SO₄ has a low vapour pressure and, once formed, easily nucleates into new particles or condensates onto existing aerosol. Simultaneously, we have replaced the prescribed optical properties representing the effect of the presence of an aerosol layer in the stratosphere. Deep volcanic eruptions are now represented by injecting SO₂ into the stratosphere, and OCS is now also explicitly modelled in the model. OCS has mainly natural sources in the troposphere, of which emissions from the ocean are the most important. It is relatively inert in the troposphere but is converted in the stratosphere into SO₂. To describe the chemical evolution, we have added the following two reactions in NorESM2-LM:





and



125 where OH is taken from the prescribed climatology. The reaction of $\text{OCS} + \text{O}$ is not taken into account. As the sources of OCS are not well known and as it has a reasonably homogeneous tropospheric concentration due to its relatively long tropospheric lifetime, we nudge the surface concentration of OCS towards the best estimate values of Meinshausen et al. (2017). We impose a surface concentration of around 0.5 ppb for OCS in our simulations. The optical depth obtained from explicitly injecting SO_2 from volcanoes and modelling OCS resulted in reasonable aerosol optical depth when 90% of the volcanic sulfur is emitted as
 130 SO_2 and 10 % as SO_4 as done in Seland Graff et al. (2025). When injecting sulfur for SAI, we have assumed the same ratio.

2.2 Experimental setup

To investigate the impacts of applying SAI to "shave-off" temperature overshoots, we first perform an idealized baseline simulation, producing a temperature overshoot by prescribing a positive followed by a negative CO_2 emission phase (each of 100 years length), and finally an extended period (200 years) with zero emissions. Other climatic forcings are kept at
 135 preindustrial level. Figure 1 illustrates the CO_2 emission trajectory that follows a bell-shaped curve, leading to cumulative emissions of 2000 and 1500 Pg C after 100 and 200 years, respectively. This reference simulation (hereafter referred to as CDR) leads to a temporary global mean temperature overshoot above 2°C , as shown in Fig. 1 (gray lines) and Schwinger et al. (2022c). To estimate the required SAI forcing for the mitigation of the temperature overshoot, we perform a second baseline simulation without negative emissions but with cumulative emissions reaching 1500 Pg C by year 100 (purple lines
 140 in Fig. 1). This 'no overshoot' simulation leads to a mean global warming below 2°C . The difference in the atmospheric CO_2 concentration between the two simulations is then used to estimate the sulfur injection required to offset the additional warming in the CDR relative to the no overshoot simulation (brown shading in Fig. 1a; see also Appendix A for more details).

Next we design a set of CDR+SAI experiments where the additional negative radiative forcing due to the sulfur injection is intended to keep the global warming below 2°C . Furthermore, to investigate the sensitivity of Earth system responses to
 145 different SAI deployment locations (e.g., Cao and Fang, 2026), we deploy a spatially uniform SAI forcing across the (i) tropics, (ii) northern and (iii) southern hemisphere mid-latitudes. The deployment altitude is set to 22.5-23.5 km, which is approximately the upper limit achievable by conventional aircraft and sufficiently above the tropopause to allow for a longer aerosol lifetime (Smith, 2020). Unlike earlier studies that typically apply temporally constant SAI forcing (e.g. Bednarz et al., 2025; Vioni et al., 2023), our SAI implementation represents a more realistic scenario in which injection strengths gradually
 150 ramp up and down before a complete termination (Fig. 1a). The three experiments combining CDR and SAI, together with the baseline/reference (CDR) experiment are summarized in Table 1. To ensure the robustness of our simulated signals, each experiment consists of three ensemble members, each covering 400 model years. During the final 200 years of the simulations neither CDR nor SAI are applied (and CO_2 emissions are zero) to allow us assessing the potential long-term impacts of SAI on various components of the Earth system.

Table 1. List of idealized sensitivity experiments with different configurations of CDR (carbon dioxide removal) and SAI (stratospheric aerosol injection). For each experiment, three ensemble members were performed using initial conditions from different preindustrial periods.

Experiment ID	Description	SAI forcing
CDR	Positive and negative CO ₂ emissions	None
CDR+SAI(TR)	Same as CDR plus SAI in the tropics	30°S-30°N
CDR+SAI(NH)	Same as CDR plus SAI in the NH mid-latitude	15°N-60°N
CDR+SAI(SH)	Same as CDR plus SAI in the SH mid-latitude	15°S-60°S

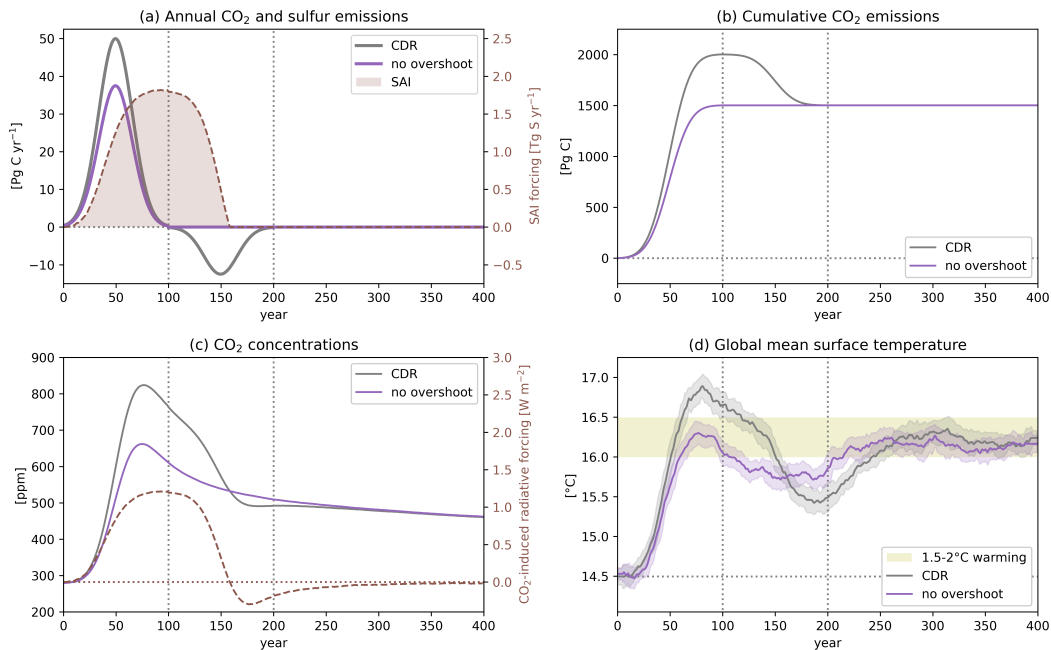


Figure 1. Annual global CO₂ and sulfur emissions. Time-series of (a) annual CO₂ emissions to produce the overshoot (CDR) and no overshoot idealized scenarios, (b) cumulative CO₂ emissions, and the projected (c) atmospheric CO₂ concentrations and (d) global mean surface temperature. Panel (a) also depicts the calculated annual global sulfur injection needed to offset the (c) radiative surplus associated with the atmospheric CO₂ difference between CDR and overshoot experiments.

155 **3 Results**

3.1 Surface temperature

Here, we focus on the transient and spatial temperature responses to SAI deployment. The long-term SAI impacts beyond this period are presented in Section 3.9. In the reference (CDR) experiment without any SAI deployment, the global mean temper-

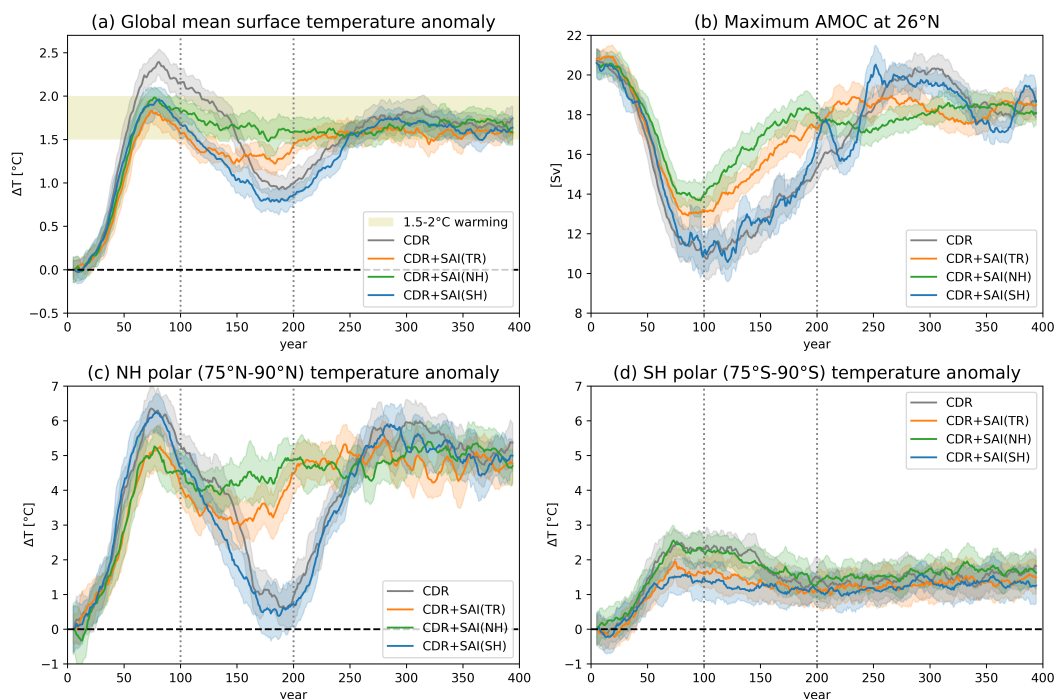


Figure 2. Simulated time-series of annual mean (a) global surface temperature, (b) Atlantic Meridional Overturning Circulation (AMOC) strength, (c) northern and (d) southern polar (75°-90° latitudes) surface temperature from experiments (grey) CDR, (red) CDR+SAI(TR), (green) CDR+SAI(NH), and (blue) CDR+SAI(SH). Shadings depict interannual variability range averaged over three ensemble members from each experiment. Time-series have been smoothed with an 11-year moving average.

ature reaches its maximum warming of nearly 2.5°C towards the end of the first 100 years, before positive emissions cease.
 160 This is followed by a gradual cooling throughout the second century (Fig. 2a, gray line), closely resembling the atmospheric CO₂ concentration trend (Fig. 1c). In all three SAI experiments, the global temperature peaks occur at approximately the same time as in the CDR experiment, but at or slightly below 2.0°C warming, effectively removing the temperature overshoot as intended. Experiment CDR+SAI(TR) simulates the lowest warming peak because sulfur injection in the tropic produces the largest global radiative cooling due to the longer aerosol lifetime and the availability of incoming shortwave radiation for
 165 reflection, making the cooling less seasonal-dependent (Duffey et al., 2026).

All three SAI experiments experience gradual cooling trends in the second century, though with different strengths. The evolution of the northern (NH) and southern (SH) hemispheres' polar temperatures (Fig. 2c,d) reveals that the strongest cooling during this period originates from the polar NH. The strong warming-cooling-warming cycles seen here are associated with the simulated transient decline and recovery of the Atlantic Meridional Overturning Circulation (AMOC, Fig. 2b), which are
 170 described in more detail in Section 3.2 (see also Schwinger et al., 2022a). While most of the simulated surface warming in the first 100 years is dominated by the increasing atmospheric CO₂ concentrations, the different cooling trends are linked to

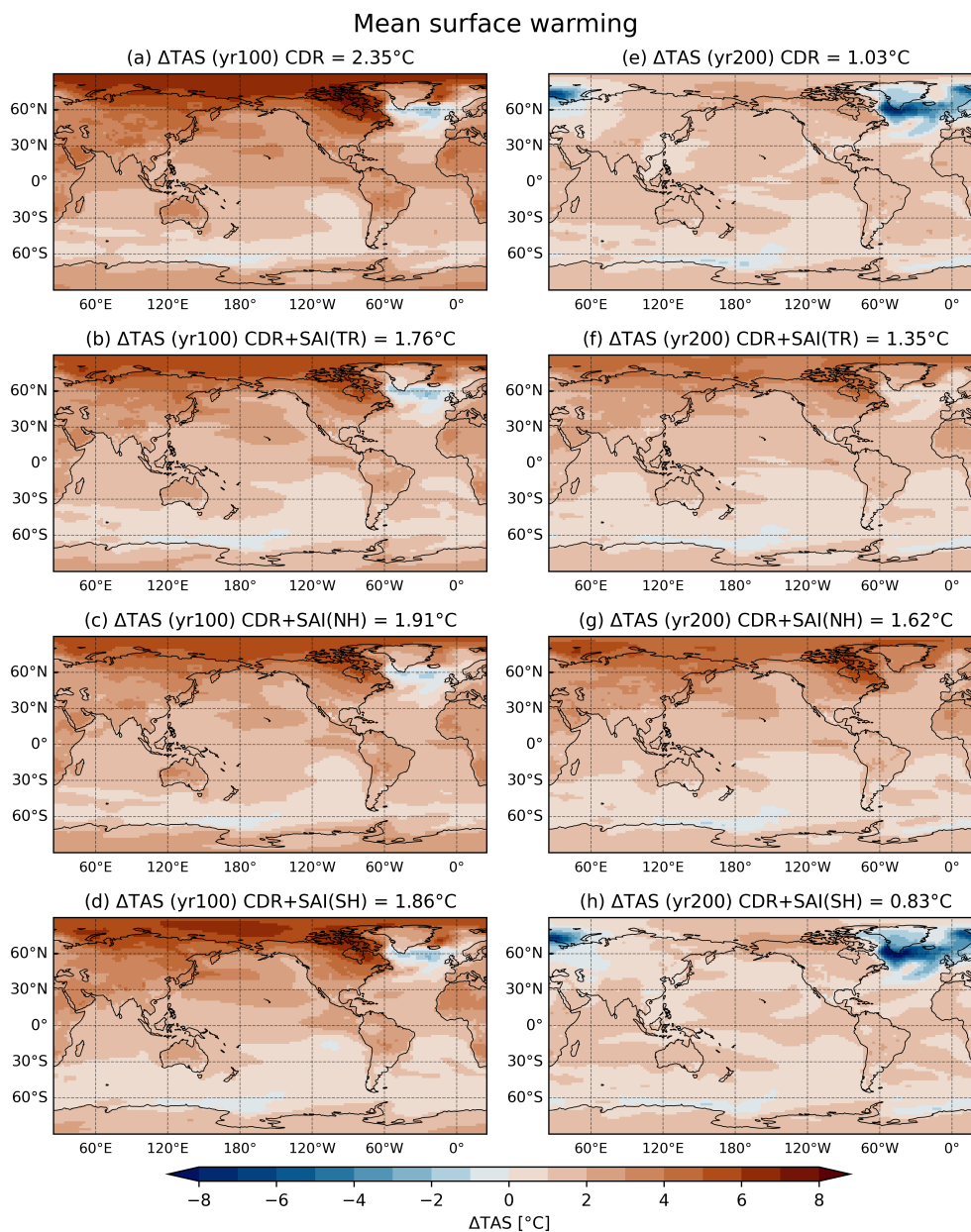


Figure 3. Spatial patterns of surface temperature anomalies after (first-column) 100 and (second-column) 200 years of simulations in (a,e) CDR, (b,f) CDR+SAI(TR), (c,g) CDR+SAI(NH), and (d,h) CDR+SAI(SH). Values are averaged over 20 years (81-100 and 181-200) from each experiment minus 20 years of the preindustrial control.

different rates of decreasing AMOC strength, reducing the poleward heat transport to and in the NH. In CDR+SAI(NH), where AMOC declines the least, the respective temperature cooling over the second century is weakest, generating the most stable



global temperature evolution following the peak warming. In contrast, CDR+SAI(SH) simulates the strongest AMOC decline and the longest duration of a low AMOC state, both of which translate into the strongest NH cooling rate (with NH polar temperatures even briefly returning to their preindustrial state), and consequently the strongest cooling in the global temperature. The AMOC evolution in the reference CDR experiment is similar to that in CDR+SAI(SH). The global temperature follows a similar trend but remains higher because of the absence of SAI-induced cooling (Fig. 2a) during the first 200 years. It is worth noting that the impact of SAI on the simulated polar SH temperatures is mainly during the SAI deployment period, i.e. this region appear to be less affected by the large AMOC changes. In the polar SH, SAI deployments in either the tropics or the mid-latitude SH produce very similar cooling effects and their transient spatially averaged temperature evolutions closely resemble one another (Fig. 2d). Here, SAI deployment in the NH shows a very similar projection to experiment CDR.

There is a clear planetary scale surface warming at the end of year 100 (Fig 3), with the exception of the North Atlantic Subpolar Gyre (associated with the AMOC decline) and parts of the polar Southern Ocean. Warming is stronger over land and in the northern hemisphere. Without SAI (experiment CDR), the global warming reaches 2.35°C, which is reduced to 1.76, 1.91, and 1.86°C with SAI deployments in the tropics, NH and SH, respectively. At the end of the second century, the overall surface temperature drops, with the strongest cooling signals in northeastern North America, northwestern Europe, as well as the North Atlantic subpolar gyre and the Arctic. This strong cooling, as described above, is associated with the lower AMOC, which is still on its recovery pathway. The cooling pattern in the polar SH is much more subtle.

3.2 AMOC

During the first century of all experiments, there is a persistent AMOC slow-down driven by CO₂-induced warming. This is followed by an AMOC recovery in the second century, which occurs over a longer timescale than the preceding slowdown, consistent with an ongoing cooling induced by negative CO₂-emissions (see also Schwinger et al., 2022c) and the SAI deployment. The different locations of the SAI deployment affect the simulated AMOC minimum and therefore the length of the recovery periods (Fig. 2b). In CDR+SAI(NH), the AMOC slows down the least and recovers the earliest. SAI injection in the tropics (CDR+SAI(TR)) reduces the AMOC strength the second least (from 21 to 13 Sv) before it gradually rebounds to roughly 18 Sv by year 200. The CDR and CDR+SAI(SH) experiments simulate the strongest warming in the high latitude NH during the first 100 years, leading to the largest AMOC reductions and minimum AMOC values of 11 Sv and a recovery that starts later in time.

AMOC variations are driven by changes in surface ocean temperature and salinity, which determine seawater density. Surface warming and freshening reduce density in the North Atlantic, increasing the surface-to-deep density gradient and leading to an AMOC slow-down, whereas cooling and increased salinity enhance surface density and favor AMOC recovery (Rahmstorf, 1995; Stouffer et al., 2006; Weijer et al., 2019). To elucidate the drivers of the projected AMOC variations, we analyze the temporal evolution of vertical density, temperature, and salinity profiles in the North Atlantic subpolar gyre (here: 53°N-66°N, 60°W-15°W) across all four experiments (Fig. 4). In both CDR and CDR+SAI(SH), the strong AMOC decline is accompanied by an extended period of strong stratification and shallow winter mixed layer depth (Fig. 4a,d). The simulated lower density across the vertical space coincides with upper ocean changes in temperature (colder) and salinity (fresher), with



Anomalies of density vertical profile in the subpolar North Atlantic

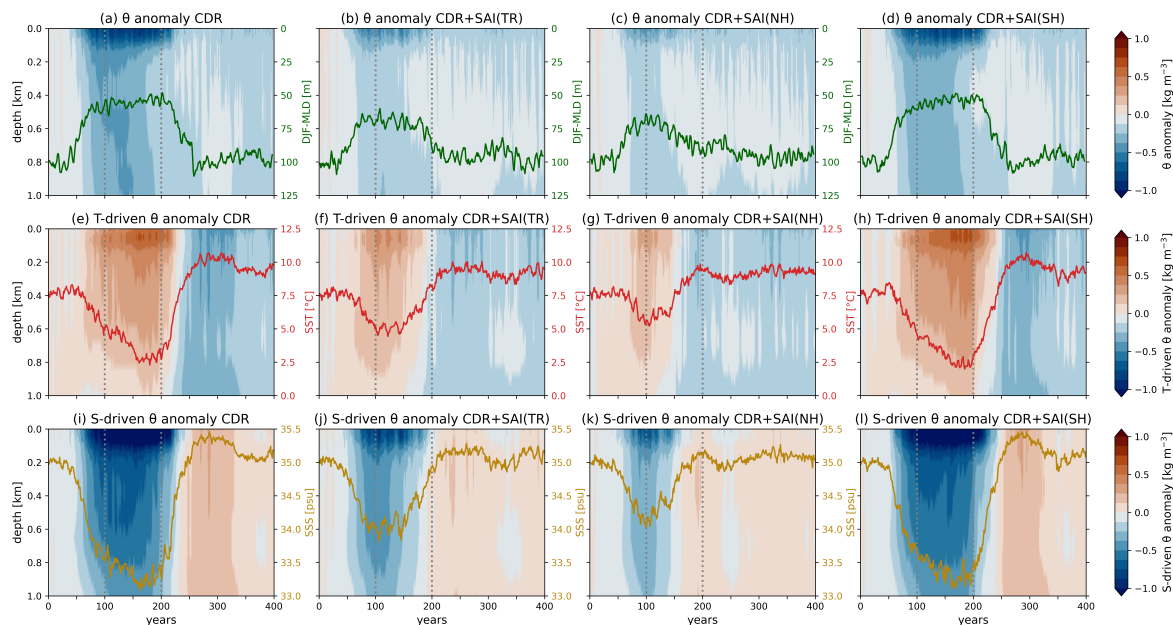


Figure 4. Temporal anomalies of (a-d) potential density, (e-h) temperature-driven potential density, and (i-l) salinity-driven potential density vertical profiles (surface to 1000 m depth) in the subpolar North Atlantic (53°N – 66°N , 60°W – 15°W). Anomalies are calculated relative to the preindustrial control. The overlay lines depict (a-d) annually averaged (a-d) winter (DJF) mixed layer depth, (e-h) sea surface temperature, and (i-l) sea surface salinity in the same horizontal domain. Shown are values from all four experiments.

the largest signals occurring near the surface layers. To a first order, these temperature and salinity anomalies are related to the AMOC strength, which transports heat and salt poleward from the tropics (see Fig. 4e-l for vertical density changes driven by temperature and salinity). The stronger stratification is primarily driven by salinity decline due to the weaker upper ocean salt transport leading to a net reduction in surface density. While the magnitude of AMOC weakening is initially driven by the strength of surface warming (Fig. 2b,c), the surface salinity feedback could enhance this weakening by further freshening and stratifying the water column. Depending on how low the AMOC becomes, our simulations suggest, this surface salinity-induced stratification can consequently hinder or delay AMOC recovery when surface air temperature cools in the following second century.

3.3 Precipitation

Global warming is projected to enhanced the Earth's hydrological cycle (Trenberth, 2011). Hence, an increase in global precipitation as the atmosphere warms during the first century is expected in all experiments. In the reference CDR experiment, precipitation increases by around 0.06 – 0.08 mm d^{-1} between years 80 and 160, followed by a temporary minimum of 0.04 mm d^{-1}

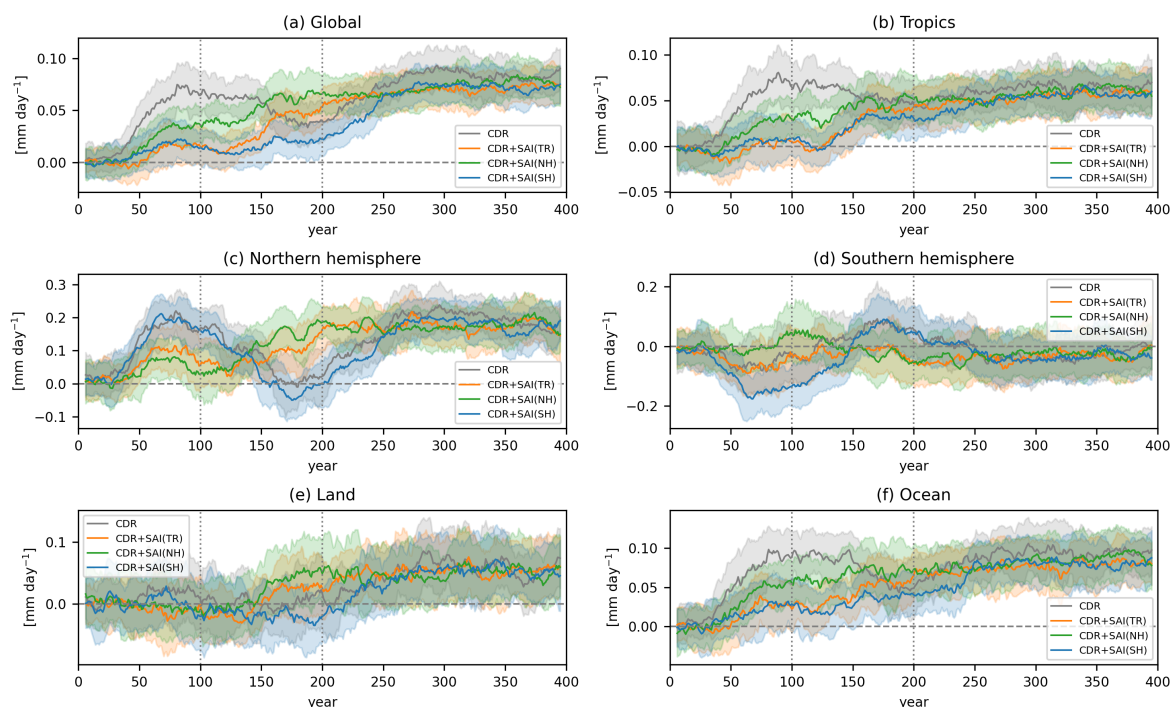


Figure 5. Time-series of precipitation anomalies relative to the preindustrial for the different scenarios in various regions: (a) global, (b) tropical (28°S – 28°N), (c) NH, (d) SH, (e) over land, and (f) over the oceans from experiments (gray) CDR, (red) CDR+SAI(TR), (green) CDR+SAI(NH), and (blue) CDR+SAI(SH). Shadings depict interannual variability range averaged over three ensemble members from each experiment. Time-series have been smoothed with an 11-year moving average.

around year 180. Global precipitation changes are initially weaker in all SAI experiments, with increases at year 80 of around 0.04 mm d^{-1} in CDR+SAI(NH), and only 0.02 mm d^{-1} in CDR+SAI(TR) and CDR+SAI(SH). At around year 200, however, CDR+SAI(NH) and CDR+SAI(TR) show a stronger response (0.05 – 0.06 mm d^{-1}) than CDR, but CDR+SAI(SH) still shows a smaller response. The global mean precipitation response is dominated by changes in the tropics (Fig. 5b) and over the oceans (Fig. 5f). The precipitation response in the NH (Fig. 5c) to some extent resembles the NH high-latitude temperature evolution (Fig. 2c). The precipitation response in the SH is overall weaker (Fig. 5d) with less precipitation during the SAI-deployment in the SH due to the related cooling. Finally, the precipitation response averaged over all land areas is relatively weak in all experiments for the first 150 years (Fig. 5e). In CDR+SAI(TR) and CDR+SAI(NH), precipitation starts to increase after year 150, but for CDR+SAI(SH) and CDR it takes more than 200 years before a significant increase becomes visible. At the end of the simulations, the global-average precipitation increases in all four scenarios and settles at a similar increase of around 0.07 – 0.08 mm d^{-1} (Fig. 5a).

Figure 6 depicts regional changes in precipitation in all experiments, where it increases most strongly around the Equator and the inter-tropical convergence zone (ITCZ) over the ocean. However, drying patterns are simulated over the Amazon and the

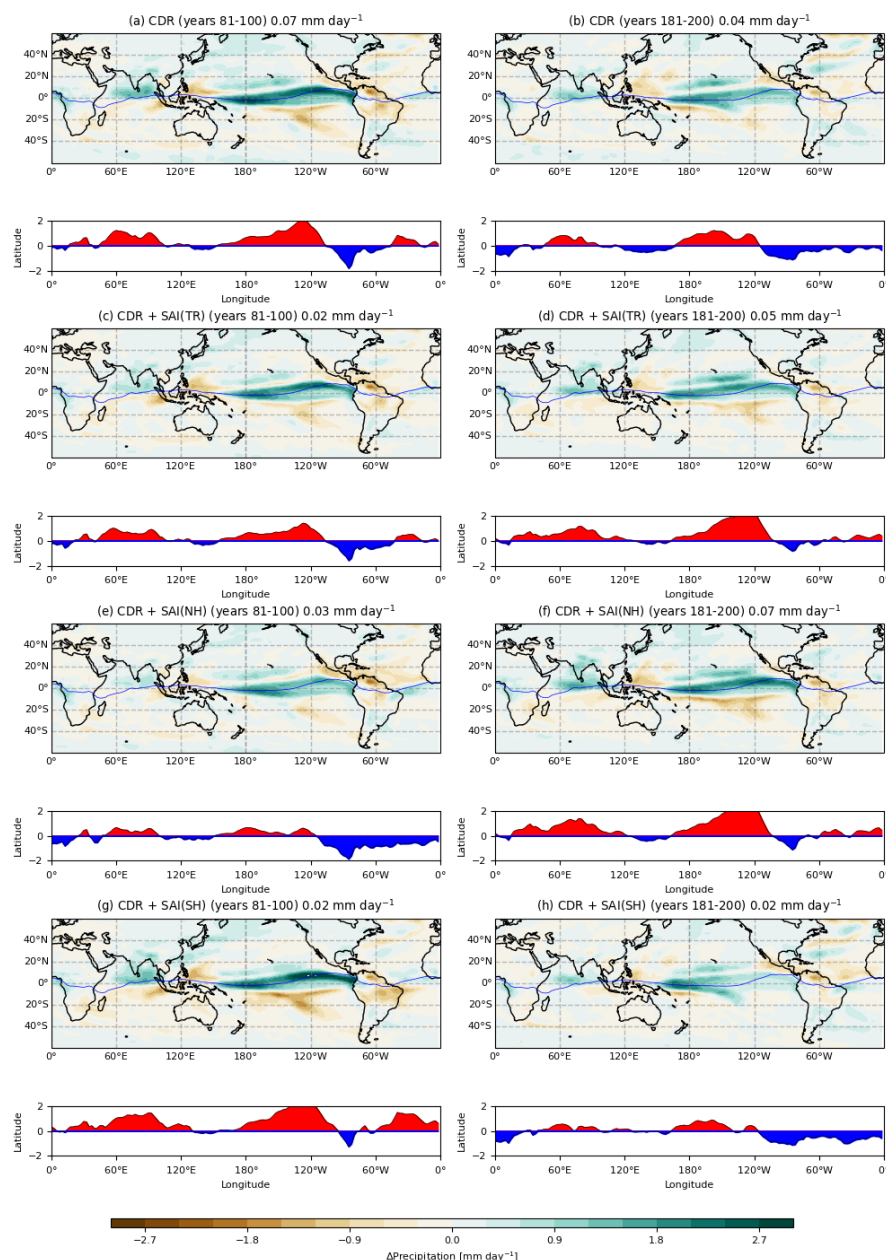


Figure 6. Maps of simulated changes in precipitation (relative to the preindustrial) for the four experiments for the periods 81–100 (left column) and 181–200 (right column). The blue lines on the maps indicate the mean position of the ITCZ in the preindustrial. Shown below each map is the position of the ITCZ compared to the preindustrial with northward (red) and southward (blue) shifts indicated. Numbers on each panel title depict the global mean change in precipitation.



maritime continent (Indonesia). NH land areas and NH Indian and Pacific oceans show an increase in precipitation, whereas SH ocean shows a general decrease in precipitation. The ITCZ shifts in all experiments mainly to the north, in accordance to the principle that the ITCZ moves to the hemisphere which warms the most (often the NH due to its abundance in land mass). In NorESM2, a northward-shift is further favored by the fact that the Southern Ocean warms only slowly in response to anthropogenic forcing (Gjermundsen et al., 2021). We see a northward shift mainly over the Indian Ocean, and over the western and central Pacific Ocean. Strongest southward shift of the ITCZ is seen over the eastern Pacific, but weaker southwards shifts can also be seen over western Africa, Indonesia and South-America. In the CDR and CDR+SAI(SH) experiments, the precipitation and ITCZ response are stronger in the period 81–100 than in the period 181–200, whereas for CDR+SAI(TR) and CDR+SAI(NH), the responses are stronger the period 181–200 than in the period 81–100. This pattern arises from the AMOC-induced temperature evolution in the NH, where both experiments CDR+SAI(TR) and CDR+SAI(NH) result in weaker NH warming than CDR and CDR+SAI(SH) over the first century (Fig. 2c). The corresponding opposite cooling pattern can be observed in the second century.

3.4 Sea-ice

Arctic sea-ice extent follows to first order the anomalies of the northern high latitude surface temperatures (Figs. 2c, 7a-c). Therefore, the magnitude of SAI-induced NH cooling influences the Arctic sea-ice changes in the first 100 years (see also Kim et al., 2025). Due to the strong AMOC decline in experiments CDR and CDR+SAI(SH), and the resulting pronounced warming-cooling-warming cycle in the northern high latitudes, Arctic sea-ice decreases, increases, and finally decreases again approaching a quasi-equilibrium state with only small trends towards the end of the simulations (Fig. 7a-c). The quasi-equilibrium state of the last 100 years is similar for all simulations, but experiments with SAI consistently show a somewhat larger Arctic summer sea-ice area. The loss of Arctic summer (September) sea-ice relative to the beginning of the simulation is between 4 and 4.2 million km² for CDR+SAI (ensemble mean average over the last 11 years) and 4.8 million km² for the CDR experiments, while Arctic winter (March) sea-ice loss is less at about 1.2 million km² for all experiments. During the period where the northern high latitude temperatures cool (Fig. 2c) in experiments CDR and CDR+SAI(SH), the Arctic winter sea-ice extent grows beyond its preindustrial boundaries, until it reaches an extent that is about 2 million km² larger towards the end of the negative emission phase in year 200. However, the September sea-ice extent remains consistently smaller than in the preindustrial period during the intermediate cooling period, with CDR+SAI(SH) reaching the largest extent, about 1 million km² less than the preindustrial sea-ice cover. Since CDR+SAI(TR) and CDR+SAI(NH) reduce the AMOC decline relative to CDR and CDR+SAI(SH) and thereby mitigate the intermediate cooling in NH high latitudes to a large extent, there are much smaller oscillations of Arctic sea-ice extent in these simulations.

The differences with respect to the evolution of the northern high latitude temperatures between CDR, CDR+SAI(SH) on the one hand and CDR+SAI(TR) and CDR+SAI(NH) on the other lead to very different histories of sea-ice cover (Fig. 8). In the CDR and CDR+SAI(SH) simulations, which show a large intermediate cooling, there are vast areas of the Arctic that lose their summer ice cover (we define an area as ice covered if sea-ice concentration is larger than 10%), regain it due to the intermediate cooling, but are again ice-free during summer towards the end of the simulations (red areas in Fig. 8). In CDR and

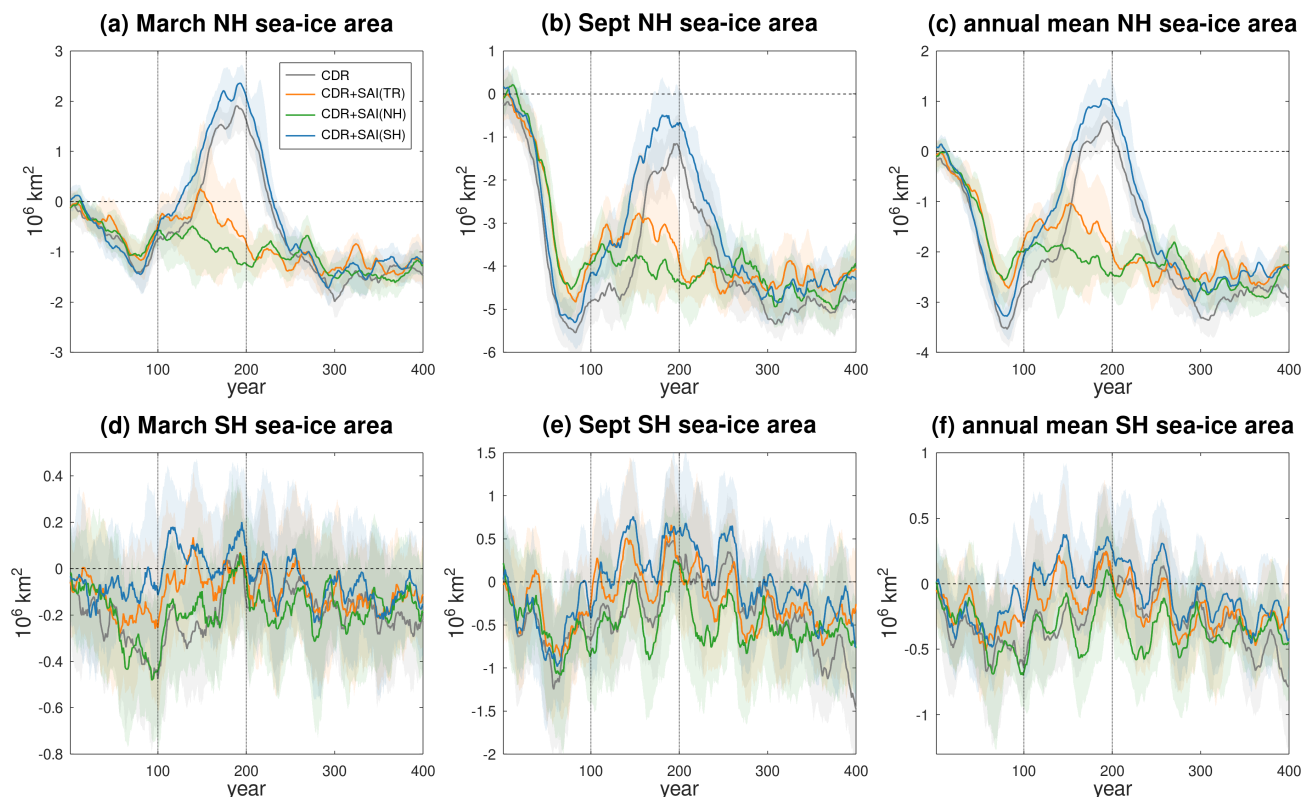


Figure 7. Time-series of (a-c) Arctic and (d-f) Antarctic sea-ice extent anomalies relative to preindustrial for (a,d) March, (b,e) September, and (c,f) the annual mean. The lines depict mean of 3 ensemble members. Time-series have been smoothed with an 11-year moving average and shadings indicate the ensemble range of the smoothed time-series.

CDR+SAI(SH), the only area that shows a stable summer ice cover over the whole period is a narrow strip along the northern coast of Greenland and the Canadian Arctic Archipelago. In contrast, experiment CDR+SAI(NH) essentially only shows two classes of areas: sea-ice covered during summer for the whole simulation period, and areas of permanent summer sea-ice loss.

270 Figure 8 also demonstrates that the larger summer sea-ice extent found in the CDR+SAI simulations towards the end of the simulation is located in the central Arctic with more sea-ice coverage extending into the East Siberian Sea and the Kara Sea (light blue and orange areas).

Changes in Antarctic sea-ice extent are generally much smaller than in the Arctic, but also follow the evolution of southern high latitude surface temperatures (Figs. 2d, 7d-f). In experiments CDR and CDR+SAI(NH), Antarctic summer (March) sea ice is reduced by about 0.5 million km² during the positive emission phase and recovers to preindustrial levels during the negative emission phase.

275 Antarctic summer sea-ice loss is reduced in experiments CDR+SAI(TR) and CDR+SAI(SH), and the sea-ice extent slightly overshoots preindustrial levels in CDR+SAI(SH) during the negative emission phase. During the last

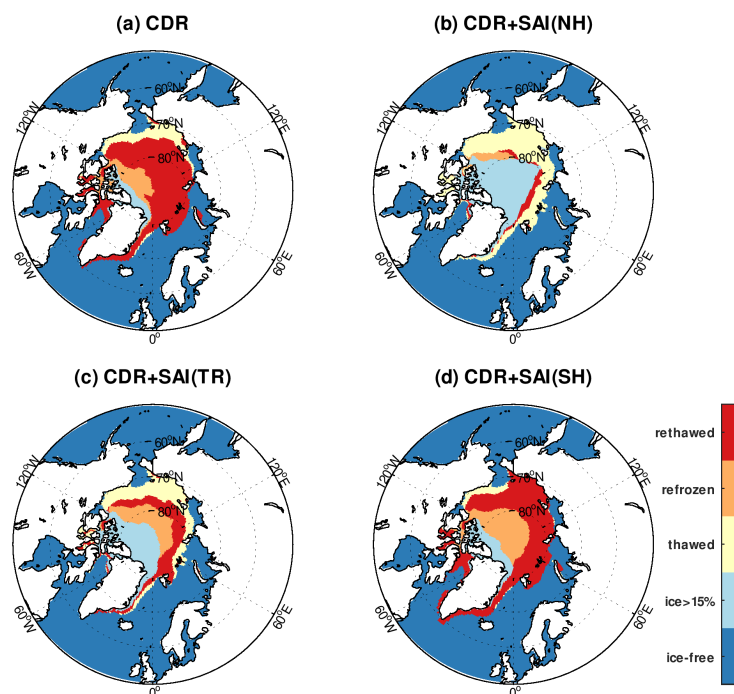


Figure 8. History of Arctic summer (September) sea-ice coverage: ice covered throughout the whole simulation (light blue); ice covered in the preindustrial, but thawed during the simulation (yellow); thawed, but refrozen (i.e., ice covered at the end of the simulation, orange); thawed, ice covered during the temporary cooling period and thawed again at the end of the simulation (red). Ice covered is defined as a sea-ice cover of 15% or more. Shown is the classification for the ensemble means of simulations (a) CDR, (b) CDR+SAI(NH), (c) CDR+SAI(TR), and (d) CDR+SAI(SH).

200 years of zero emission, all experiments show a very similar evolution of Antarctic summer sea-ice with a gradual decline from preindustrial levels to a reduction of 0.1 to 0.2 million km². The evolution of Antarctic winter (September) sea ice is qualitatively very similar to the summer sea-ice. The main difference is that the decrease of sea ice extent is larger than in summer (about 1.2 million km²) during the positive emission phase and that there is no effect of SAI on this decrease.

3.5 Ocean heat content and Sea-level rise

Increasing ocean heat content (OHC) is the driver of steric sea-level rise. OHC increases rapidly in all experiments during the positive emission phase and continues to increase throughout the negative and zero-emission phases, although at a slower rate (Fig. 9a). This continued increase reflects the slow equilibration of the deep ocean. Although atmospheric CO₂ concentrations and surface temperatures begin to decline after the emissions peak, heat continues to be transported from the upper ocean into deeper layers, allowing the ocean to accumulate heat long after the surface climate starts to recover (Held et al., 2010;

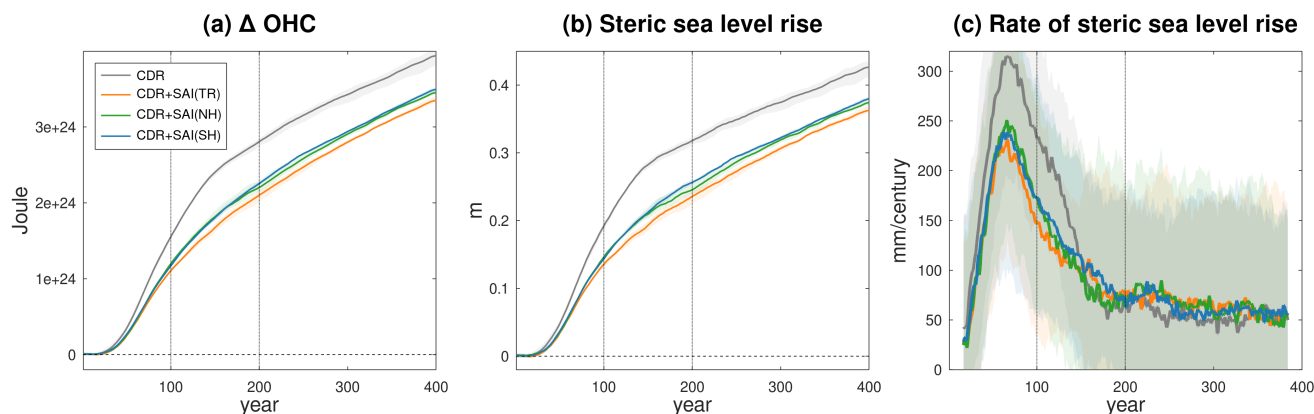


Figure 9. Change in (a) ocean heat content, (b) global mean steric sea-level rise, (c) and rate of steric sea-level rise. Shown is the ensemble mean, smoothed with a 11-year running mean. The rate of steric sea-level rise has been smoothed by applying a 31-year running mean. Shading indicates the ensemble range of the smoothed time-series

Frölicher et al., 2014). Due to the reduced radiative forcing in the experiments with SAI, the increase of OHC and steric sea level is smaller in the SAI experiments than in the reference CDR simulation (Fig. 9a,b).

290 The rate of steric sea level rise peaks at $315 \text{ mm century}^{-1}$ at year 65 in experiment CDR, 15 years after positive emissions peak and around the period of highest CO_2 concentration (Fig. 9c). The peak rate in all SAI simulations has similar timing but remains lower with values between 230 and $250 \text{ mm century}^{-1}$. After the peak, the rate of sea level rise steeply decreases for the remainder of the positive emission phase and during negative emissions (Fig. 9c). While the SAI simulations show a smaller rate of sea level rate during positive emissions and SAI deployment, the CDR simulation shows a slightly lower
 295 rate of steric sea level rise compared to the SAI simulations during the zero emission phase. This is consistent with the fact that the ocean loses some of the excess heat previously taken up during the overshoot back to the atmosphere, leading to slightly warmer surface temperatures (Fig. 2a). All simulations converge towards the same rate of steric sea level rise of around $50 \text{ mm century}^{-1}$ towards the end of the experiment, with only a very slowly decreasing trend. The difference in steric sea level at year 200, that is the sea level rise that has been mitigated by the application of SAI is between -8.2 cm for CDR+SAI(TR)
 300 and -6.1 cm for CDR+SAI(SH).

3.6 Ocean carbon fluxes

In the baseline experiment CDR, ocean CO_2 uptake increases rapidly during the positive emissions phase, reaching a maximum of approximately 9 Pg C yr^{-1} around year 50, before atmospheric CO_2 peak (Melnikova et al., 2021), and followed by a decline that continues into the negative emissions phase (Fig. 10a). The ocean uptake briefly reverses (the ocean is turned into a source
 305 of carbon) around year 150 before recovering to a persistent carbon sink during the remainder of the simulation.

All three SAI experiments closely follow the evolution of experiment CDR, indicating that the global ocean carbon sink is only weakly affected by SAI deployment. Differences in the global uptake rates relative to experiment CDR generally remain

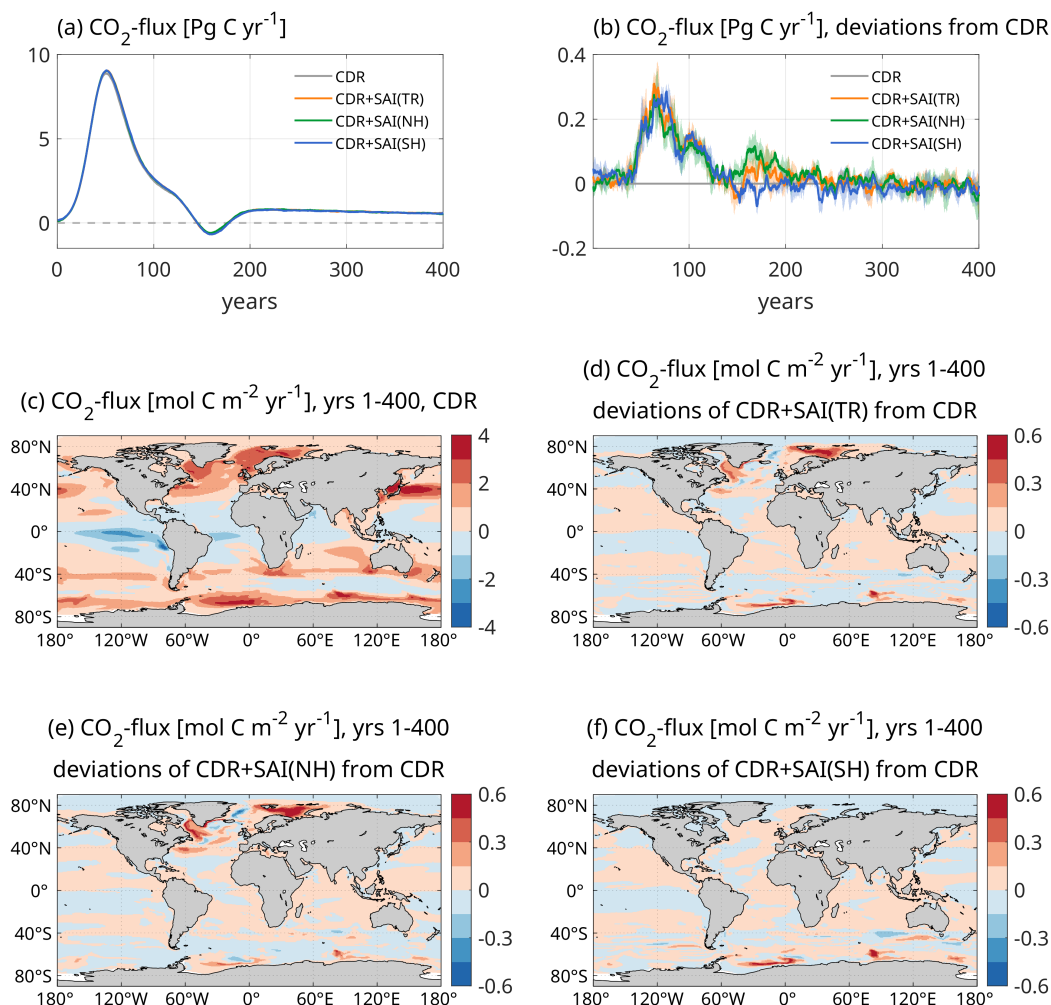


Figure 10. (a) Time-series of global air-sea CO₂ fluxes for all experiments and (b) deviations in global CO₂ fluxes relative to the CDR experiment. Both time-series have been smoothed by applying a (symmetric) 10-year running mean. Shading in (a,b) indicates the ensemble spread. (c) Map of mean air-sea CO₂ flux over years 1–400 in the CDR experiment. Mean air-sea CO₂-flux anomalies relative to the CDR experiment for the (d) CDR+SAI(TR), (e) CDR+SAI(NH), and (f) CDR+SAI(SH) experiments. Positive values depict ocean uptakes

below 0.4 Pg C yr⁻¹ and are largest during the period of active SAI deployment (Fig. 10b). The long-term mean CO₂ flux in the experiment CDR exhibits strongest uptake in the North Atlantic and the Southern Ocean, while regions of weaker uptake or net outgassing occur primarily in the tropics (Fig. 10c). SAI deployment induces modest regional perturbations (Fig. 10d–f), with the largest responses occurring in the Southern Ocean and, for CDR+SAI(TR) and CDR+SAI(NH), in the North Atlantic.



We investigate how SAI modifies the ocean carbon sink under the CDR overshoot pathway in six regions with distinct carbon-cycle characteristics (Fig. 11a). The regional boundaries were selected to capture dominant modes of air–sea CO₂ exchange by combining (i) the long-term CO₂ flux pattern in the experiment CDR (Fig. 10c), (ii) the regions exhibiting the largest CO₂ flux anomalies in experiments with SAI (Fig. 10d–f), and (iii) the dominant seasonal controls on surface ocean pCO₂ (i.e. thermal vs non-thermal processes) during the first decade of experiment CDR (Fig. A1) following Goris et al. (2018, 2023). In CDR, the SH subpolar region exhibits the largest CO₂ uptake during the positive emission phase, followed by the SH ice-covered region and the NH and SH subtropical and North Atlantic subpolar regions (Fig. 11b). During the atmospheric CO₂ decline, the ocean warming does not decline proportionally because of climate-system hysteresis (Fig. A1). In most regions, this persistent warming results in a larger reduction in CO₂ uptake during the atmospheric CO₂-decline phase than the increase observed during the preceding atmospheric CO₂ increase. During the extension period, the ocean gradually returns toward its initial climate state, and CO₂ uptake also slowly approaches its initial level. The ice-covered regions, however, exhibit an additional opposing effect: persistent warming reduces sea-ice cover, increasing the area available for air–sea gas exchange and thereby enhancing CO₂ uptake. Thus, in these regions, the net response reflects the competing effects of persistent warming and sea-ice loss.

The SAI deployments redistribute regional CO₂ uptake while producing only minor changes in the global ocean carbon sink (Fig. 10b). The global anomalies are dominated by only three high-latitude regions. The SH ice-covered region contributes the most to the enhanced ocean carbon uptake in all three SAI experiments, while the North Atlantic subpolar and NH ice-covered regions provide the largest additional contributions in experiments CDR+SAI(TR) and CDR+SAI(NH). Other regions contribute only weakly to the global deviations from the CDR experiment (Fig. 11c–h). The CO₂ uptake of the subtropical regions is primarily controlled by the sea surface temperature (SST) (Fig. A1) and therefore highly sensitive to SAI-induced temperature changes. Tropical and NH SAI produce the strongest cooling in the NH subtropics (Fig. A2), resulting in enhanced CO₂ uptake during the first century (Fig. 11c). However, because mixed layers remain shallow (Fig. A3), only a fraction of the absorbed anthropogenic carbon is transferred into the ocean interior, causing an enhanced outgassing as atmospheric CO₂ declines. In contrast, CDR+SAI(SH) induces only weak cooling in the NH subtropics and consequently only a small CO₂ flux anomaly. All SAI experiments produce only minor changes in the SH subtropics (Fig. 11f), consistent with the comparatively weak SST response in this region (Fig. A2). The SH subpolar region exhibits only weak CO₂ flux anomalies under all SAI experiments (Fig. 11g) since SAI induces only minor changes in SST, mixed-layer depth, and surface DIC (dissolved inorganic carbon concentration) here (Figs. A2– A3, DIC not shown).

The NH ice-covered region responds the most strongly to tropical and NH SAI (Fig. 11e). Both experiments substantially reduce sea-ice cover (Fig. 7a–c), particularly compared to experiment CDR, exposing a larger ocean surface to air–sea gas exchange. Together with deeper mixed layers (Fig. A3), this promotes the transfer of anthropogenic carbon into the ocean interior, resulting in sustained increases in CO₂ uptake. In contrast, CDR+SAI(SH) produces only a weak response. A similar but stronger response occurs in the North Atlantic subpolar region (Fig. 11d). Tropical and NH SAI mitigate the AMOC decline to some extent (Fig. 2b), leading to deeper mixed layers (Fig. A3) and enhanced winter ventilation. This results in a more efficient transfer of anthropogenic carbon into the ocean interior, thereby sustaining enhanced CO₂ uptake. Again,

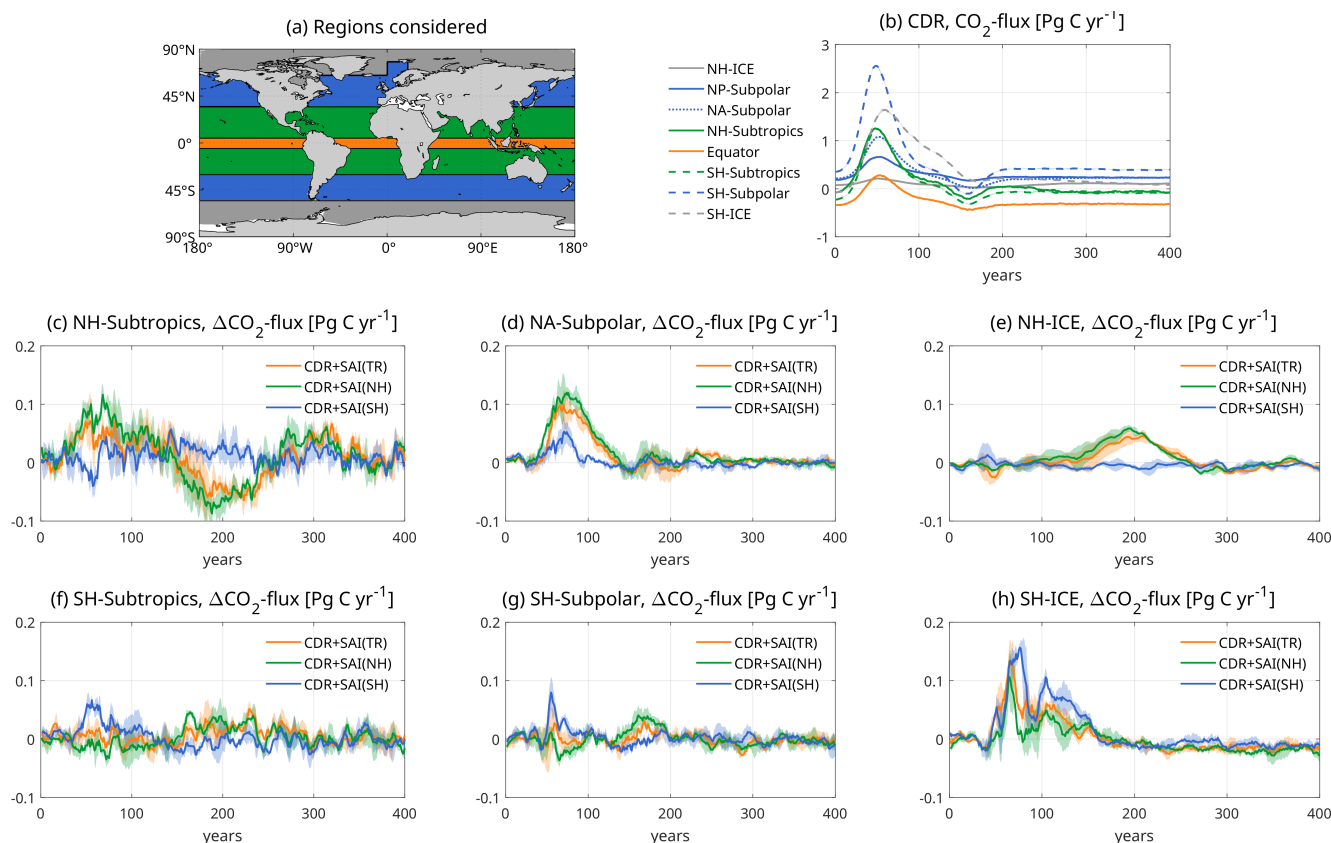


Figure 11. Time-series of regional air-sea CO_2 fluxes for all experiments. (a) Regional mask applied, (b) time-series of regional air-sea CO_2 flux in the CDR experiment and (c-h) time-series of regional CO_2 flux anomalies relative to the CDR experiment. Shading in (b-h) indicates the ensemble spread. All time-series have been smoothed by applying a (symmetric) 10-year running mean

SAI deployment in SH has only a minor influence on the regional circulation and consequently produces only weak CO_2 flux anomalies. The largest regional CO_2 -flux response occurs in the SH ice-covered region (Fig. 11h), where CDR-SAI(SH) produces the strongest enhancement of CO_2 uptake, followed by CDR+SAI(TR) and CDR+SAI(NH). This response is strongest in the Weddell, Ross and Kerguelen Gyres and appears to be mainly driven by local warming and a reduction in Antarctic sea-ice cover in the CDR+SAI(NH) simulation (Fig. 7d-f) as well as by local cooling in the CDR+SAI(SH) and CDR+SAI(TR) simulations, respectively.

3.7 Ocean acidification

Ocean acidification is a byproduct of anthropogenic CO_2 absorption by the ocean (see also Sect. 3.6), and is identified as one of nine planetary boundaries that indicate how much human activities have pushed the Earth system outside its safe operating space (Richardson et al., 2023). Figure 12a-d shows that the vertical and temporal evolution of ocean acidification in our



Vertical profiles of pH anomalies

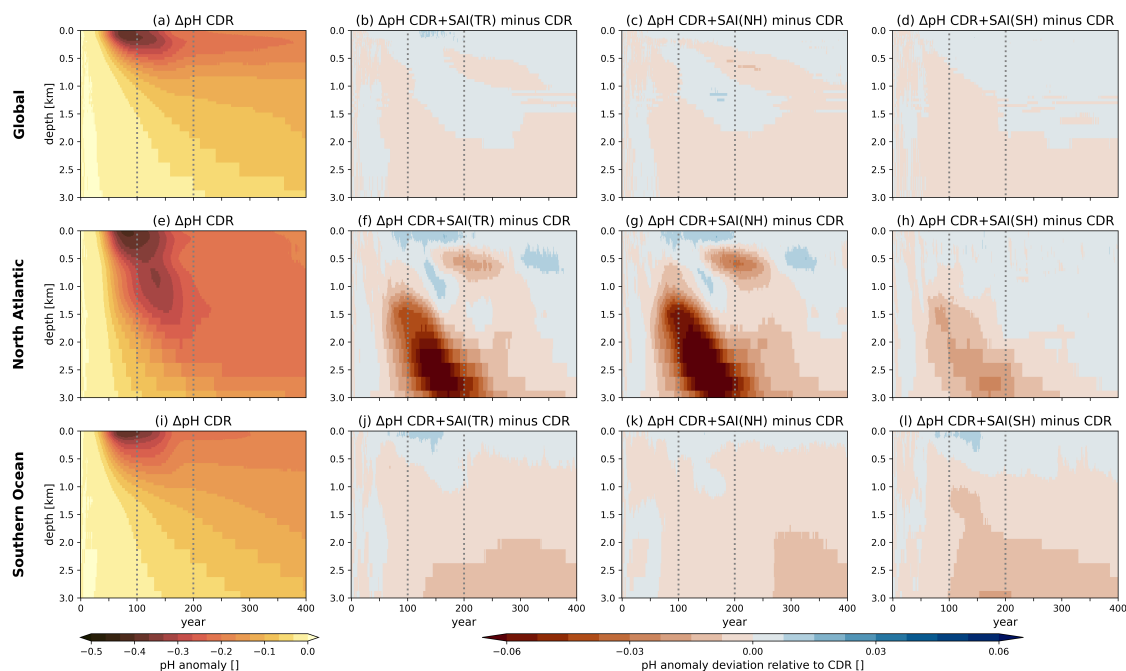


Figure 12. Temporal evolution of vertical (surface to 3 km depth) pH profile anomalies in the ocean, averaged for the (a) global ocean, (e) North Atlantic ($>30^{\circ}\text{N}$), and (i) Southern Ocean ($>45^{\circ}\text{S}$) for experiment CDR. Anomalies are calculated relative to the preindustrial. Deviation relative to CDR are shown for experiments (b,f,j) CDR+SAI(TR), (c,g,k) CDR+SAI(NH), and (d,h,l) CDR+SAI(SH).

experiments generally follows that of the atmospheric CO_2 concentration (as noted by Lauvset et al., 2015) with stronger co-variation in the upper ocean and delayed signals in the interior (see also Tjiputra et al., 2023). In all experiments, the largest surface pH deficits, greater than 0.3 pH unit, occur just before year 100, and then pH gradually recovers but stays below its preindustrial values. Below the surface, both ocean acidifications and their recoveries occur at a considerably slower pace with deeper layers still showing ongoing acidification at the end of the simulations. Consistent with earlier studies (e.g., Tjiputra et al., 2016), SAI generally does not considerably alter the projection of global mean pH values (Fig. 12a-d). However, at regional scale, especially oceanic regions that contain deep water formation sites such as the North Atlantic and the Southern Ocean (Fig. 12e-l), the acidification signals are more pronounced and more strongly affected by the SAI deployments than the global mean signal. In the North Atlantic, CDR+SAI(TR) and CDR+SAI(NH) simulate a stronger pH deficit below 1 km depth, due to their stronger AMOC and hence a more effective transport of low-pH watermasses into the interior. In the CDR+SAI(SH), the simulated cooler Southern Ocean also leads to stronger and deeper acidification.

The impact of acidification on marine ecosystems is measured by the carbonate ion saturation state of the seawater (e.g., Aragonite saturation state or Ω_A), in which values lower than one indicate under-saturation and could have detrimental consequences for calcifying marine organisms (Orr et al., 2005). Figure A4a-d) shows that Ω_A decreases with depth and that



Simulated ΔT , ΔDIC , ΔpH , and Ω_A at 2000m year 200

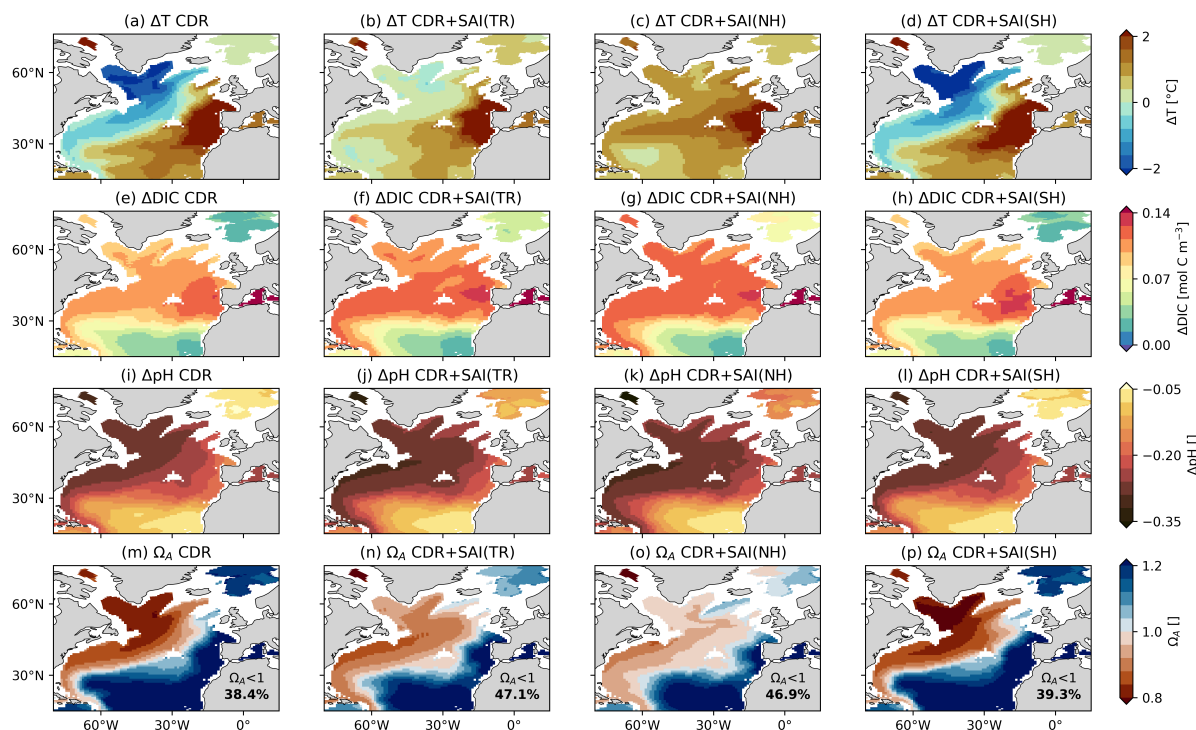


Figure 13. Maps of changes in ocean (a-d) temperature, (e-h) dissolved inorganic carbon concentration, (i-l) pH, and (m-p) aragonite carbonate saturation state distributions in the North Atlantic at 2000 m depth by the end of year 200 (181-200) relative to the preindustrial period as simulated in (a,e,i,m) CDR, (b,f,j,n) CDR+SAI(TR), (c,g,k,o) CDR+SAI(NH), and (d,h,l,p) CDR+SAI(SH) experiments.

initially under-saturation is found below 2 km in the global mean, but substantially deeper (>3 km) in the North Atlantic. When averaged globally, the Ω_A under-saturation of each experiment shoals to approximately 0.3 km by year 100 before it gradually deepens again as atmospheric CO_2 is lowered through CDR. In the deep ocean below 1 km, Ω_A continues to decline until the end of the simulation, consistent with projected pH (Figure 12a-d). While the SAI deployment does not change the globally averaged Ω_A values much in our simulations, Ω_A declines faster in the North Atlantic for CDR+SAI(TR) and CDR+SAI(NH) (Fig. A4e-h) due to the higher transport of anthropogenic carbon into the deep ocean, consistent with these simulations having a more vivid AMOC. In the Southern Ocean (Fig. 12i-l), SAI deployment in the SH alleviates the region's warming and stratification, slightly enhancing the ocean acidification and leading to lower Ω_A rates in the Southern Ocean.

When averaged over the North Atlantic domain, both CDR+SAI(TR) and CDR+SAI(NH) project the strongest decline in Ω_A when compared to the other experiments. Interestingly, when looking at the horizontal spatial patterns at 2 km depth, Fig. 13m-



p reveals although CDR+SAI(TR) and CDR+SAI(NH) project the largest area with under-saturated Ω_A by year 200, CDR and CDR+SAI(SH) project more severe under-saturation state in the subpolar gyre. Although the CO_2 increase in the seawater is largest in CDR-SAI(NH), as also reflected in the wide spread low pH area (see dissolved inorganic carbon concentration and pH in Fig. 13e-l), the Ω_A under-saturation state is surprisingly the least severe compared to the other experiments. This can be attributed to the thermodynamic of seawater carbonate chemistry. In addition to dissolved inorganic carbon concentration, Ω_A also depends on the temperature (warmer seawater leads to higher Ω_A). The relatively strong AMOC in CDR-SAI(NH), not only transports high- CO_2 but also warmer water masses in the interior, where the latter opposes the severity of Ω_A decline. In contrast, despite having lower anthropogenic CO_2 invasion in experiments CDR and CDR+SAI(SH), the considerably weakened AMOC induces an anomalously cold subpolar North Atlantic that extends into depths and toward the southwest, which exacerbates the low Ω_A values.

3.8 Land carbon cycle

The terrestrial carbon cycle response to CDR and CDR+SAI interventions reveals significant differences in how carbon is regionally stored and reallocated. Global vegetation carbon (cVeg) peaks across all simulations around year 130 at an increase of about 500 Pg C, then drops by about 100 Pg C over the next decades, followed by a gradual decline as CO_2 concentrations and temperatures stabilize (Fig. 14a). There is a notable time lag of around 50 years of the cVeg response to atmospheric CO_2 concentration due to enhanced carbon retention under negative emissions (Park and Kug, 2022; Asaadi et al., 2024; Swann et al., 2026). Global cVeg stays elevated until the end in all simulations because of the stabilization of temperatures and higher than initial CO_2 levels. Both experiments CDR+SAI(NH) and CDR+SAI(SH) yield similar projections of global vegetation carbon inventory, and are very close (within 10 Pg C) to the baseline experiment CDR. Global soil carbon (cSoil) shows a similar response (Fig. 14b). It increases by around 100 Pg C in all simulations, with a difference between experiments ranging about 10 Pg C. Here, experiments CDR+SAI(TR) and CDR+SAI(SH) gain slightly more cSoil than CDR and CDR+SAI(NH) experiments. In contrast to the cVeg response, the global cSoil does not lose carbon during the negative emissions phase but rather stabilize. The cVeg and cSoil trends superimpose in the global cumulative net biome production (NBP; Fig. 14c). Although differences are marginal, all CDR+SAI experiment variants peak slightly higher than the CDR reference between year 100–200 but end up slightly below by the end of the simulations, reflective of the lower cVeg inventory. This indicates that SAI deployments - irrespective of location - are slightly less effective at preserving global land carbon sinks in the long term in NorESM2-LM.

In the high-latitude regions, the land carbon response is largely controlled by changes in permafrost, which stores large amounts of frozen organic carbon. As permafrost thaws under warmer conditions, this carbon becomes increasingly susceptible to decomposition, reducing cSoil storage (Schuur et al., 2022). Permafrost extent responds inversely to the northern high-latitude temperature trajectories, and stabilizes at approximately 10 million km^2 irrespective of the experiment (Fig. 15a–b). At the same time, warmer conditions and increasing atmospheric CO_2 concentrations enhance vegetation productivity by extending the growing season and increasing photosynthetic carbon uptake, leading to a gain in cVeg (Swann et al., 2026). Although the changed conditions stimulate an increase in high-latitude cVeg from about 35 Pg C to approximately 60 Pg

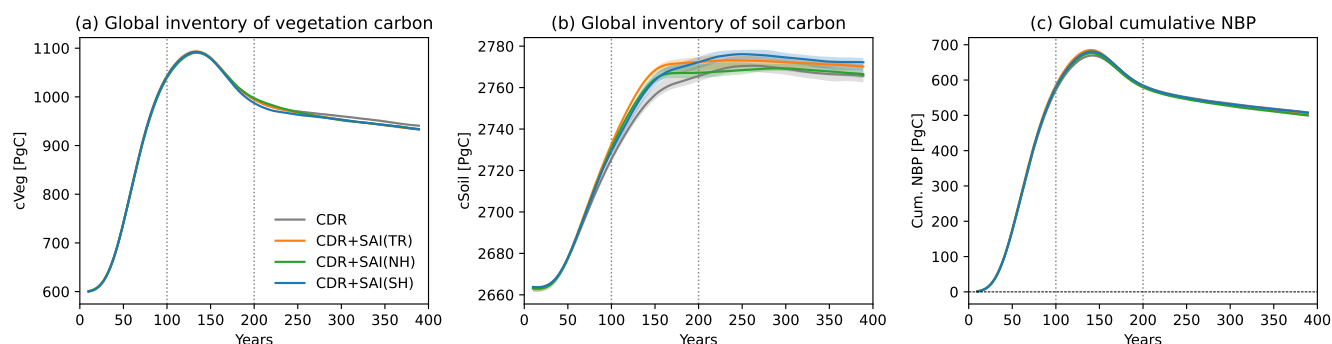


Figure 14. Global mean time-series of (a) vegetation and (b) soil carbon inventory, and (c) cumulative net biome productivity (NBP), as simulated in all experiments. A running mean of 21 years is shown. The lines denote the average of 3 ensemble members, and the shading its spread.

415 C, this gain is ultimately outpaced by the permafrost thaw-driven continuous and substantial loss of cSoil of up to 70 Pg C (Fig. 15c–d). This imbalance causes the high-latitude cumulative NBP to turn negative after about 200–300 years, hence the region transitions from a net carbon sink to a net source (Fig. 15f) (Varney et al., 2026). The strength of this transition is determined by the cumulative warming exposure of the various experiments, so that the AMOC-induced strong cooling in CDR+SAI(SH) produces the lowest high-latitude cSoil loss (Steinert et al., 2026). Soil carbon loss is most severe in the CDR
 420 experiment, where the lack of SAI-induced cooling leads to greater and faster permafrost thaw and carbon release. Notably, the relatively larger initial warming (first 200 years) in the CDR experiment drives the northern high latitudes to become a net carbon source about 100 years earlier than in the SAI experiments, with an offset sustained until the end of the simulation due to differences between the experiments in cumulative warming exposure (Steinert et al., 2026).

The Amazon region exhibits a similar peak-and-decline evolution in carbon storage, with cumulative NBP reaching its max-
 425 imum around year 140, indicative of being predominantly driven by the vegetation carbon response. Notably, both vegetation and soil carbon remain higher throughout all simulations. The effect of SAI deployment on vegetation soil carbon uptake in this region is minimal (Fig. 15j–k). However, attenuated temperature and precipitation fluctuations in the CDR+SAI(TR) and CDR+SAI(NH) experiments (Fig. 15h–i) lead to temporarily lower cVeg accumulation between years 200–300. While the CDR and SAI+CDR(SH) experiments show significant fluctuations in regional temperature and precipitation due to the
 430 AMOC effect on the ITCZ displacement and its subsequent effects on the tropical hydrological cycle (Steinert et al., 2025), these fluctuations also have a minimal effect on the carbon storage variability.

A spatial comparison of the CDR+SAI(TR) and CDR simulations highlights the specific regional effects of the tropical SAI deployment on NBP (Fig. 15e). CDR+SAI(TR) damps carbon storage primarily across tropical and subtropical land areas because of decreased total vegetation carbon accumulation, with most prominent decreases in cumulative NBP in the Amazon
 435 Basin, Central Africa and Southeast Asia. A similar effect can be seen in the boreal forest of Eurasia and North America. Meanwhile, increases in cumulative NBP predominantly appear across the northern high latitudes in the CDR+SAI(TR) relative

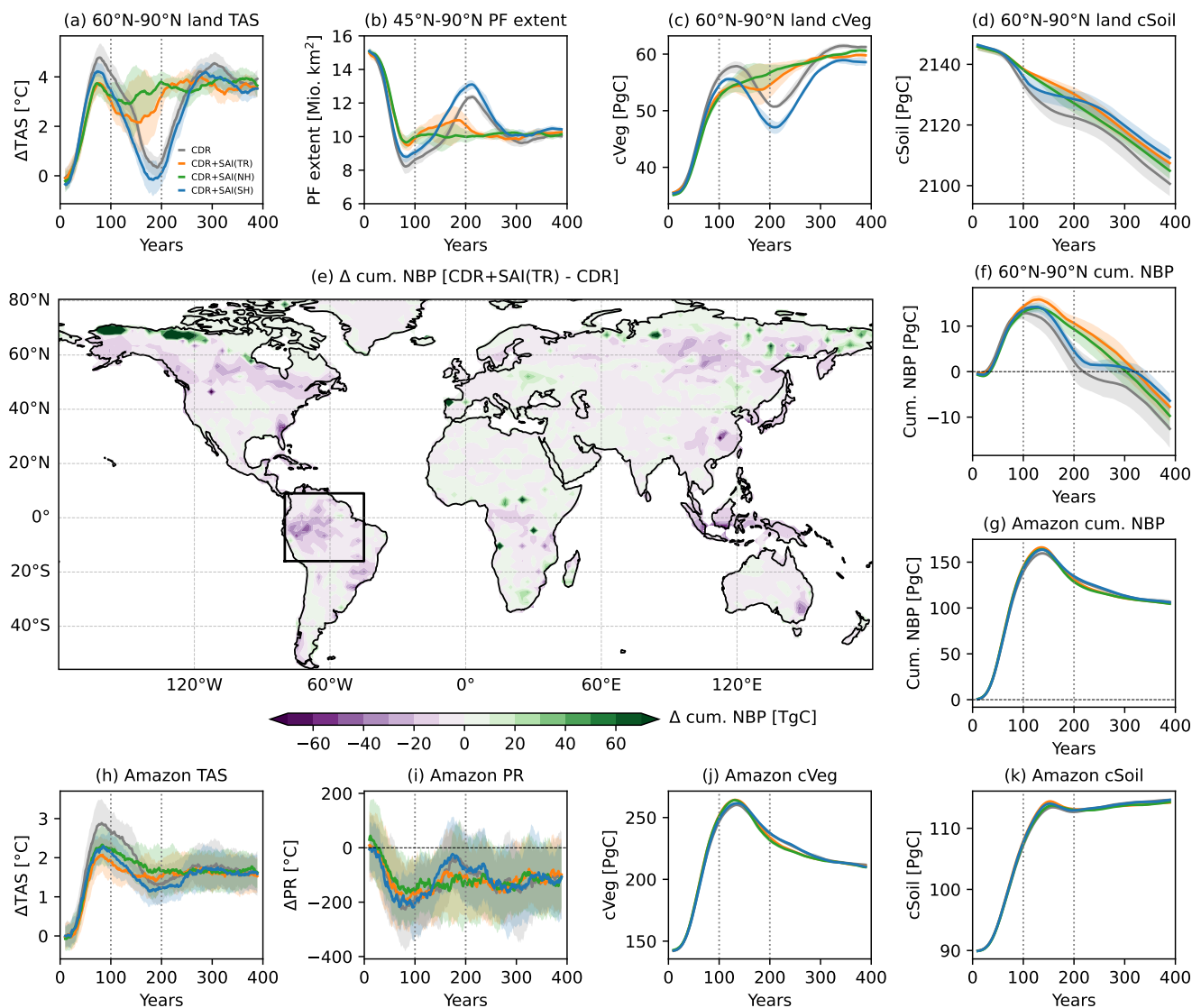


Figure 15. Regional mean time-series of the (a) high latitudes 60°N-90°N land surface temperature (TAS), (b) 45°N-90°N permafrost extent, (c) 60°N-90°N vegetation and (d) soil carbon inventories, and (f) 60°N-90°N cumulative NBP. Spatial difference of 400-year cumulative NBP between CDR+SAI(TR) and CDR experiments ensemble mean is depicted in panel (e). Mean timeseries of the Amazon (80°W-45°W, 16°S-9°N; black box) (h) temperature, (i) precipitation, (j) vegetation and (k) soil carbon inventories, and (g) cumulative NBP. The time series show a 21-year running mean. The solid lines denote the average of 3 ensemble members, and the shading its spread. The reference period for TAS and PR is the average of the first 10 years of the respective simulation.

to CDR, where permafrost degradation is mitigated due to SAI-induced cooling (see also Lee et al., 2019). These results indicate that, despite lowering peak temperatures, SAI could impair the capacity and potentially the stability of tropical land carbon



sinks compared to the higher-temperature overshoot conditions associated with the reference CDR scenario (Lee et al., 2021; Munday et al., 2025; Swann et al., 2026).

3.9 Legacy

Here, we assess the simulated climate change over the last 200 years (third and fourth centuries) where neither CDR nor SAI forcing are implemented and the net anthropogenic CO₂ emissions are kept at zero. The objective is to assess the long-term residual impacts and memory of SAI on different aspects of the climate system.

The AMOC, which is a key long-term climate regulator in all our experiments, is still evolving in the last two centuries. This slow adjustment reflects the long response timescale of the ocean circulation, which is governed by the gradual redistribution of heat and freshwater within the ocean and can persist for centuries after an external forcing is reduced (Rahmstorf, 1995; Jackson, 2013). Additionally, evidence from paleo proxy and modeling studies indicate that natural variability of AMOC, e.g. during the last Interglacial, occurs at multi-centennial time scales (Galaasen et al., 2020; Kessler et al., 2020). Here, our experiments demonstrate that SAI deployments aimed to shave-off temperature overshoot modify the transient AMOC response, and that the recovery of the circulation are slow-paced and hence continue after the cessation of the intervention. Figure 2b shows that AMOC gradually recovers from its minimum level around year 100 onward. In the last 200 simulation years, some pronounced temporal variations are still observed in all experiments. The recovery pathways differ substantially among the experiments depending on the magnitude and duration of the perturbations. CDR+SAI(NH), which experiences the weakest AMOC decline, shows the fastest recovery and becomes relatively stable towards the end of the simulations. In contrast, CDR+SAI(SH), characterized by the strongest and most persistent AMOC reduction, exhibits a slower recovery but eventually and temporarily reaches the highest AMOC strength among the experiments. The AMOC in CDR+SAI(SH) also maintains a strong temporal variability over the last two centuries. These differences highlight the nonlinear and history-dependent nature of AMOC adjustment, where the magnitude and duration of the initial perturbation influence the subsequent recovery trajectory (Stommel, 1961; Rahmstorf, 1995; Thomas et al., 2019).

Experiments CDR and CDR+SAI(SH) show clear long-term trends in global mean surface temperature that persist over nearly the whole third century (Fig. 2a), by up to 0.1° decade⁻¹. Experiments CDR+SAI(NH) and CDR+SAI(TR) maintain a relatively stable temperature evolution in the last 200 years. These pathway-dependent temperature trends are amplified in the polar NH (Fig. 2c), where CDR and CDR+SAI(SH) simulate a third century warming rate that is six times faster than their global mean warming rate. Regionally, the subpolar North Atlantic is warming at >1.0° decade⁻¹ (Fig. 16), while in the low latitudes and SH, the warming rates are more subtle with parts of the subantarctic experiencing slight cooling. The spatial warming patterns in the NH are linked to the evolution of AMOC in experiments CDR and CDR+SAI(SH), which is in both cases still recovering from its previously weak state (Fig. 2b). Particularly, both CDR and CDR+SAI(SH) show an increasing AMOC strength from roughly 16 to 20 Sv in the third century, steadily increasing the heat transport into the North Atlantic.

In the fourth century, global temperature stabilizes between 1.5-1.7° over preindustrial levels in all experiments, which is slightly lower than the peak warming in the first century. Figure 16 shows that the regional trends have also dissipated, with most regions experiencing trends well below 0.01° decade⁻¹. In CDR and CDR+SAI(SH), parts of the Arctic have cooling

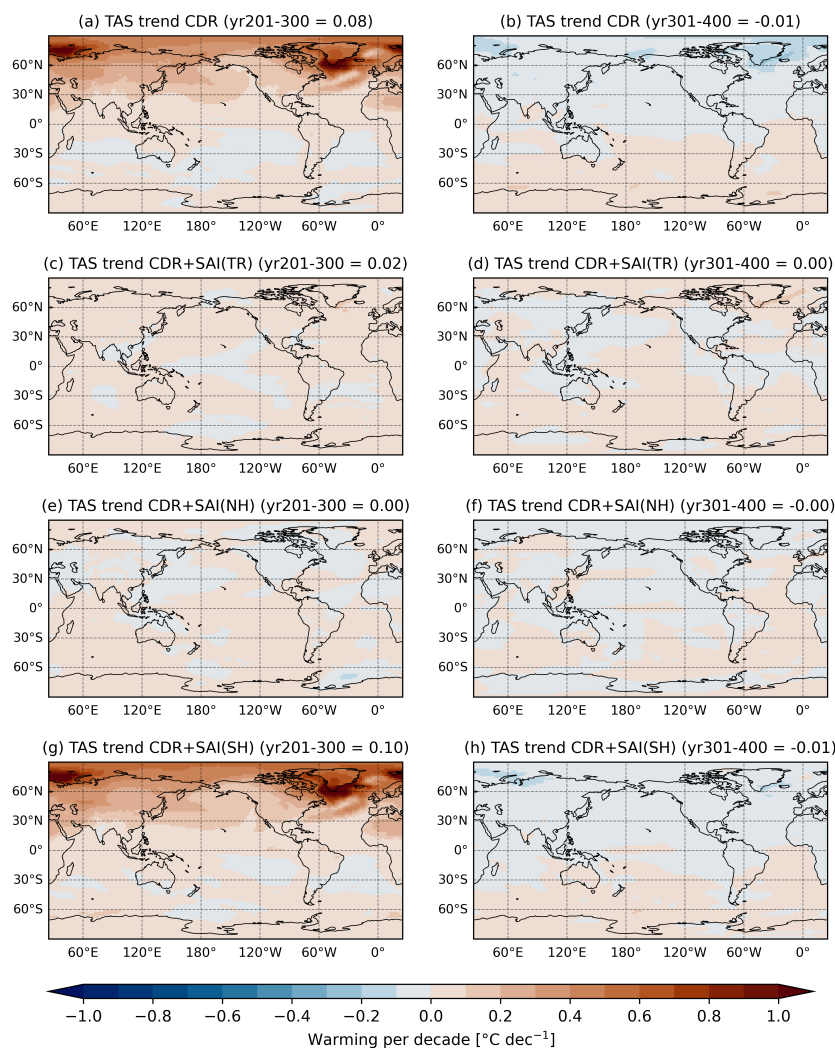


Figure 16. Regional trends in surface temperature over the third (years 201-300) and fourth (years 301-400) centuries as simulated by (a,b) CDR, (c,d) CDR+SAI(TR), (e,f) CDR+SAI(NH) and (g,h) CDR+SAI(SH) experiments. Numbers in the panel titles depict global mean trends.

trends below $-0.1^{\circ} \text{decade}^{-1}$, which are likely associated with the declining AMOC during the last century of the simulation. The similar stabilized temperature levels across all experiments are consistent with the TCRC (Transient Climate Response to cumulative Emissions) relationship (i.e. the cumulative anthropogenic emissions in all experiments are the same at 1500 Pg C)



(see also Schwinger et al., 2022c). Nevertheless, even though all experiments have cumulative emissions of 1500 Pg C (Fig. 1b), temperature oscillations can still occur due to the non-equilibrated AMOC evolution associated with the SAI deployments.

For many of the climate metrics considered in the previous subsections, significant trends and variability can still be observed over the third century that tend to stabilize in the final century. These patterns can often be linked to the non-equilibrated AMOC states, for instance, as seen in the shift in ITCZ-induced precipitation changes. Similarly, in regions that are strongly affected by physical changes that are induced by AMOC, such as the NH-ICE region and the NH-Subtropics, the CO₂-flux to the ocean is still regionally recovering in the third century (Fig. 11), although the atmospheric CO₂ concentration is relatively similar across all experiments (Fig. A6). Delayed AMOC-induced changes in interior ventilation also continue to influence the transport pathways of dissolved inorganic carbon and therefore the spatial acidification rates. In addition, the carbonate saturation state - a measure of ocean acidification impacts on marine ecosystems - is furthermore affected by the SAI-induced perturbation in the watermass temperature transported into the deep ocean, e.g., in the subpolar North Atlantic. Therefore, interior ocean global pH is still declining at the end of simulation period and the signals of SAI-dependent interior carbonate saturation state persist beyond the SAI deployment time scale (e.g., Figs. 12, 13, and A4).

The legacy of SAI can also be observed on the projected ocean heat content, steric sea level rise, Amazon vegetation, and NH permafrost extent, where the initial regional cooling in the first two centuries has a lasting impact into the fourth century. For instance, the rates of NH permafrost loss, ocean heat content and sea level rise are alleviated during the period SAI is applied, while the Amazon vegetation tends to experience faster loss rate. Arctic summer (September) sea-ice extent is slightly larger during the last century of our simulations for the cases where SAI has been deployed (independent of the location) compared to the simulation with CDR only.

495 **4 Discussions and summary**

The impact of using stratospheric aerosol injection (SAI) to partly offset anthropogenic climate change and climate impacts has profound implications for potential decision space in future climate simulations. While arguments for and against the deployment of SAI are outside the scope of this study, the discussion around whether SAI deployment may or may not increase the probability of avoiding societally and climatically critical tipping points is a compelling motivation to enhance the evidence base necessary to inform climate mitigation actions.

Here, in the context of keeping global warming below the Paris Agreement target, we perform a set of idealized model experiments in emissions-driven, fully coupled Earth system model, that consider SAI deployments together with large-scale carbon dioxide removal (CDR) to keep global warming under 2°C, complimenting earlier study by Baur et al. (2025). We simulate SAI deployments in three different regions: the tropics and the northern and southern hemisphere mid-latitudes. In addition to improving our understanding of SAI efficacy to avoid global temperature overshoot, our study also aims at better understanding the long-term adjustment of slow Earth system components after the CDR and SAI interventions have ceased. We therefore investigate not only the transient Earth system response but also extend our simulations for 200 years after termination of the CDR and SAI interventions. This approach complements earlier overshoot peak-shaving experiments

through SAI implementation (Tilmes et al., 2020), while extending them through an explicit comparison of deployment regions
 510 and an assessment of the multi-centennial climate legacy.

In the baseline experiment (CDR without SAI), a temperature overshoot ($>2^{\circ}\text{C}$) that lasts for 57 years is simulated. The
 application of SAI is suitable to "shave off" this overshoot, with SAI deployment in the tropics leading to a stronger global
 cooling effect than injections at mid-latitudes of either hemispheres (i.e. based on same rate of sulfur injections used in all
 experiments). This is related to the longer aerosol lifetime in the tropical stratosphere and less seasonal constraint of reflecting
 515 incoming solar radiation, consistent with Duffey et al. (2026). Beyond the peak warming, the temperature trajectories diverge
 considerably among the different experiments. We show that this is driven by the Atlantic Meridional Overturning Circulation
 (AMOC) evolution, which is sensitive to the SAI deployment location.

The subpolar North Atlantic is a key region controlling AMOC variability (Chafik et al., 2022). In our experiments, SAI
 deployment in the northern hemisphere produces a cooling in this region and ameliorates the projected AMOC decline and
 520 advances the rebound period. On the contrary, the deployment of SAI in the SH has only a small direct effect on the NH
 temperature, such that the AMOC decline and recovery is not mitigated compared to the experiment without any SAI. Despite
 the strongest global average cooling effect, SAI injection in the tropics mitigates less of the AMOC decline compared to
 northern mid-latitude SAI but more than the southern mid-latitude SAI deployment. Throughout the study, we show how these
 diverging AMOC trajectories manifest different responses in various components of the Earth system. While previous studies
 525 have shown that the AMOC response depends strongly on the latitude of SAI deployment (Li et al., 2023; Bednarz et al., 2025),
 our study places these differences in the context of a temporary temperature-overshoot pathway, where SAI is applied only to
 "shave off" the the global mean temperature overshoot. By comparing tropical, Northern Hemisphere and Southern Hemisphere
 deployments, and following the coupled Earth system for two centuries after SAI termination, we demonstrate how deployment
 location shapes not only the transient AMOC response but also the long-term evolution of ocean circulation and its cascading
 530 impacts on multiple Earth system components. These results highlight that similar near-term global temperature trajectories
 can conceal markedly different long-term regional climate responses, depending on the chosen deployment strategy.

Our results demonstrate that SAI deployment strategies that produce similar near-term global mean temperature trajectories
 could nevertheless lead to a substantially different regional climate evolution, where precipitation patterns, ocean circulation,
 cryosphere responses, and carbon cycle dynamics differ substantially. We therefore reiterate the need for caution when gauging
 535 climate change mitigation based on a single simplified metric such as global mean warming. This finding is consistent with
 increasing recognition that regional climate impacts and Earth system responses cannot be inferred from global mean temper-
 ature alone (e.g., Seneviratne et al., 2025). Similarly, changes in low frequency climate variability and their respective delayed
 impacts should also be taken into account when assessing any mitigation plans.

Our study is unique to most earlier studies as it considers simulations that include both SAI and CDR climate intervention,
 540 but uncertainties remain due to the use of a single model. Previous model intercomparison studies have shown substantial
 model spread in projections of AMOC decline to future warming, reflecting structural uncertainties in the representation of
 deep-water formation, ocean mixing and air-sea interactions (Weijer et al., 2020). NorESM2 simulates a comparatively strong
 AMOC response to anthropogenic forcing (Weijer et al., 2020; Menary et al., 2020; Schwinger et al., 2022a). At the same



time, compared to many other CMIP6 models, NorESM2 exhibits a relatively low equilibrium climate sensitivity (Seland et al.,
 545 2020a), implying that, for a given radiative forcing, simulated global warming is relatively moderate. Consequently, while we
 expect that the mechanisms identified in our study are qualitatively robust, the magnitude of the simulated temperature and
 AMOC responses, and therefore the associated Earth system impacts, may differ in models with different climate sensitivity
 or ocean circulation characteristics. For instance, Li et al. (2023) demonstrate diverging AMOC responses to similar SAI
 forcing in two different model systems, which they attribute to variations in the underlying physical processes involved in the
 550 AMOC responses and consequent impacts. Acknowledging the complexity of the AMOC and its model-dependent responses to
 external forcing, future coordinated multi-model studies applying a similar experimental concept and extending those proposed
 in the Geoengineering Model Intercomparison Project (GeoMIP, Vioni et al., 2024) will be valuable for quantifying the
 robustness of the mechanisms identified here and assessing model-dependent uncertainties.

The key findings of our study can be summarized as follows:

- 555 – Temporary deployment of SAI can effectively avoid an overshoot of global mean surface temperature in an idealized
 CDR scenario.
- The latitude of SAI deployment strongly influences the transient and long-term evolution of the AMOC.
- SAI deployment in the Northern Hemisphere most effectively mitigates the magnitude and duration of AMOC weaken-
 ing, thereby reducing long-term temperature variability in the Northern Hemisphere.
- 560 – AMOC-induced NH warming modulation modifies the meridional temperature gradient, which can shift the ITCZ posi-
 tion and precipitation patterns.
- Changes in the AMOC also propagate into multiple slow Earth system components, including Arctic sea-ice, permafrost,
 ocean heat uptake, steric sea-level rise, precipitation and land carbon storage.
- In our simulations, SAI only marginally enhances ocean carbon uptake, primarily through changes in deep-water forma-
 565 tion and ventilation.
- Changes in ocean circulation modify subsurface North Atlantic acidification through both altered carbon transport and
 ocean temperature.
- Regional SAI deployment alters the long-term distribution of land carbon through differences in vegetation growth and
 permafrost soil carbon loss.
- 570 – Long-term Earth system adjustment continues for centuries after SAI termination, demonstrating that deployment strat-
 egy leaves a persistent climate legacy.

Finally, SAI influences a much broader range of Earth system processes than those investigated here. While substantial
 progress has been made in understanding the climatic response to SAI, important uncertainties remain regarding its impacts on
 societally-relevant processes such as marine ecosystem, extreme events, drought, vegetation mortality, agriculture and human



575 systems (Lauvset et al., 2017; Fan et al., 2021; Müller et al., 2024; Zhang et al., 2025; Roberts et al., 2026). We note that
marine ecosystem impacts and extreme events analysis of our experiments are planned in a separate manuscript. Our study
demonstrates that, even when SAI is applied only temporarily to avoid a temperature overshoot, its impacts extend well beyond
global mean temperature. The deployment location shapes the long-term evolution of the AMOC, which in turn affects regional
climate trajectories, sea-ice evolution, permafrost, ocean heat uptake, steric sea-level rise, ocean and land carbon cycling, and
580 subsurface ocean acidification for centuries after SAI termination. These results highlight that assessing SAI solely in terms
of its ability to meet a global temperature target is insufficient, and that regional near-term as well as long-term Earth system
responses must also be considered. Our study does not advocate for the deployment of SAI; efforts to rapidly and strongly
reduce CO₂ emissions should be prioritized. Nevertheless, the idealized experiments presented here provide new process-
level understanding of how temporary SAI deployment interacts with carbon dioxide removal during a temperature overshoot,
585 and identify deployment location as a key determinant of long-term Earth system evolution. Future work should extend this
framework to more realistic overshoot scenarios (Van Vuuren et al., 2026) that include evolving aerosol emissions, CO₂ and
non-CO₂ greenhouse gases, as well as activity-driven CDR (Schwinger et al., 2026), and should evaluate these responses across
multiple Earth system models to quantify structural uncertainties.



Appendix A: Quantification of sulfur injections

We estimate how much SAI is needed to bring down the global mean temperature in the CDR scenario to the no overshoot scenario (see Fig. 1d). Since these two experiments only vary in the CO₂ concentration (i.e. other forcings such as GHGs N₂O, CH₄, CFC-11, and CFC-12, land-use, volcanic, solar, etc. were kept constant), we estimate the radiative forcing (RF) difference using the expression from the IPCC Third Assessment report (Ramaswamy et al., 2001, their Table 6.2):

$$RF = 5.35 \ln \frac{[CO_2]}{[CO_2]_0}. \quad (A1)$$

Figure 1c shows the CO₂ concentration evolution in both experiments, which allows to calculate the RF induced by the CO₂. The CO₂ in CDR scenario is higher in the first 160 years, but lower between years 160 and 300, as also reflected in the global mean temperature evolution. To offset this excess warming in CDR, we need a negative forcing from SAI between year 0 and 160 (with its strongest forcing around year 90), followed by a small positive forcing afterwards in the year 160–350 (with the strongest positive forcing around year 180). As it is not possible to induce a positive forcing from SAI, we stop the SAI forcing in year 160.

To estimate the effective radiative forcing (ERF) from SAI injection between 22.5 and 23.5 km altitudes in the 30°S–30°N latitudinal band, we perform fixed-SST simulations (under preindustrial conditions), with varying emission strengths: 0, 2 and 5 Tg S yr⁻¹. Based on that we derived a sensitivity of around 1 W m⁻² per spatially uniform 1 Tg S yr⁻¹ injections in the 30°S–30°N and 22.5–23.5 km.

Combining the strength of the CO₂ RF with the sensitivity of the SAI allowed us to estimate the time-evolution of the SAI strength. Initial tests in fully-coupled setup indicated however that we needed 50 % stronger sulfur emissions to bring down the global mean temperature in the CDR scenario to the no-overshoot scenario (corresponding with 1 W m⁻² per spatially uniform 1.5 Tg S yr⁻¹). The actual time-evolution of the SAI emission strength used in the experiments can be seen in Fig. 1a.

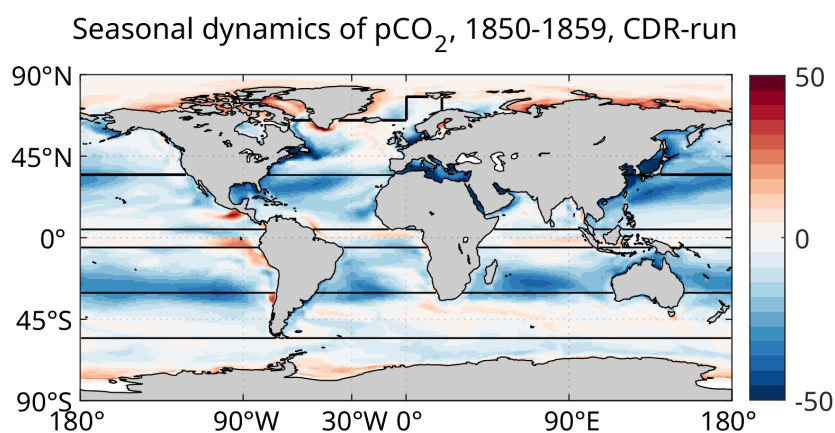


Figure A1. Map of mean seasonal amplitudes of oceanic $p\text{CO}_2$ in $\mu\text{atm per year}$ during the first 10 years of the CDR experiments, calculated as November to April minus May to October for the Northern Hemisphere and May to October minus November to April for the Southern Hemisphere. Black contours show the regional mask. Blue shading indicates thermal control while orange shading indicates that non-thermal processes are the drivers of the seasonality.

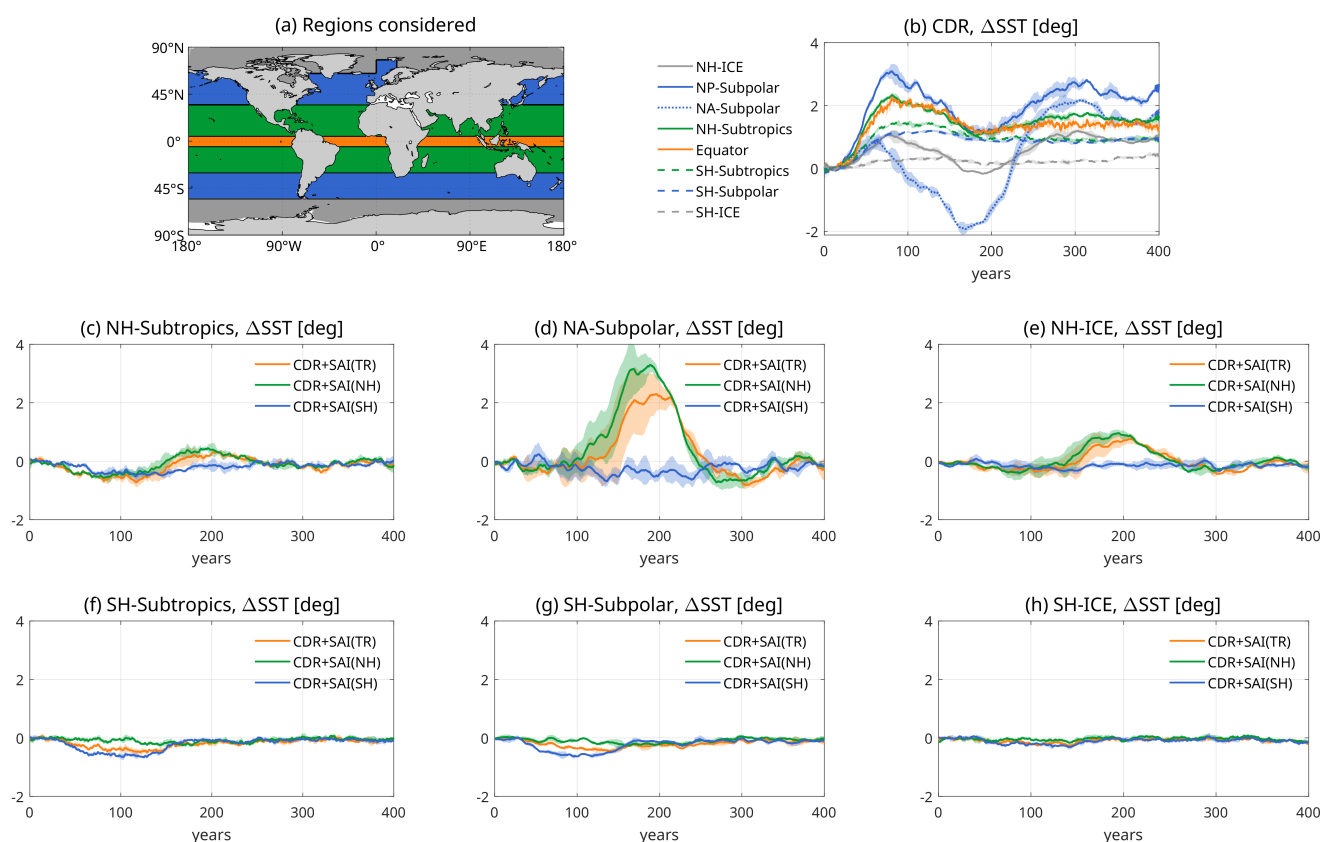


Figure A2. Time-series of regional sea surface temperature for all experiments. (a) Regional mask applied, (b) time-series of regional sea surface temperature in the CDR experiment expressed relative to the first 10 years of the CDR experiment and (c-h) time-series of regional sea surface temperature anomalies relative to the CDR experiment. Shading in (b-h) indicates the ensemble spread. All time-series have been smoothed by applying a (symmetric) 10-year running mean.

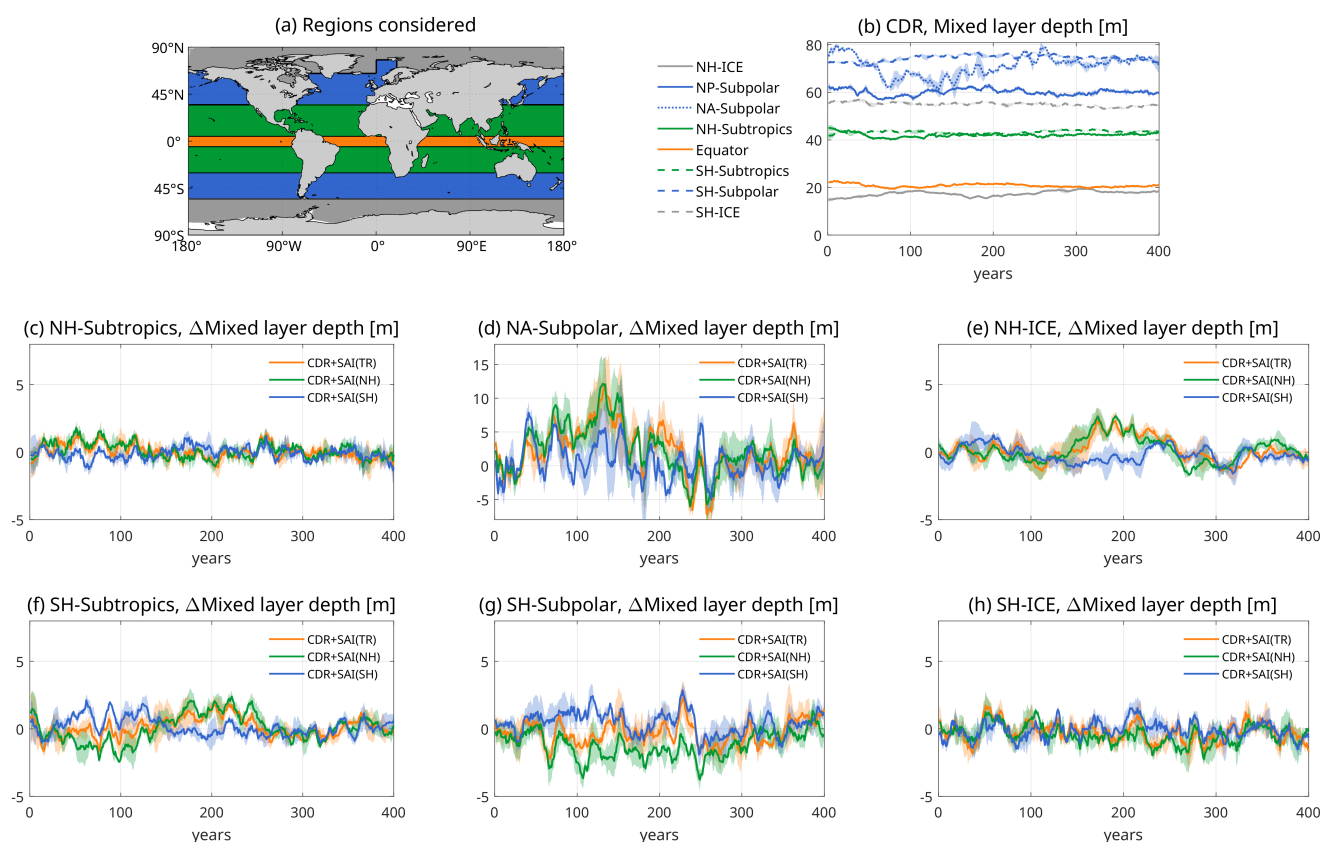


Figure A3. Time-series of regional winter mixed layer depth for all experiments (calculated as an average over January to March for the Northern Hemisphere and July to September for the Southern Hemisphere). (a) Regional mask applied, (b) time-series of regional mixed layer depth in the CDR experiment and (c-h) time-series of regional mixed layer depth anomalies relative to the CDR experiment. Shading in (b-h) indicates the ensemble spread. All time-series have been smoothed by applying a (symmetric) 10-year running mean.

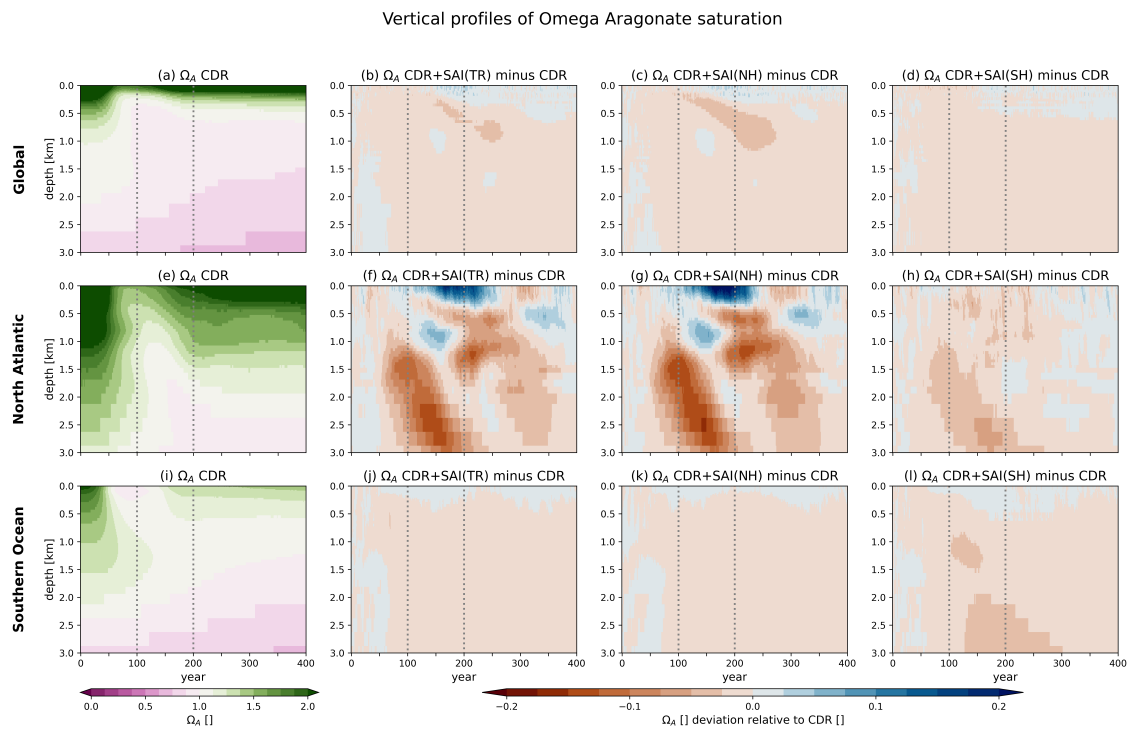


Figure A4. Temporal evolution of vertical (surface to 3 km depth) aragonite saturation state (Ω_A) profile in the ocean, averaged for the (a) global ocean, (e) North Atlantic ($>30^\circ\text{N}$), and (i) Southern Ocean ($>45^\circ\text{S}$) for experiment CDR. Deviation relative to CDR are shown for experiments (b,f,j) CDR+SAI(TR), (c,g,k) CDR+SAI(NH), and (d,h,l) CDR+SAI(SH).

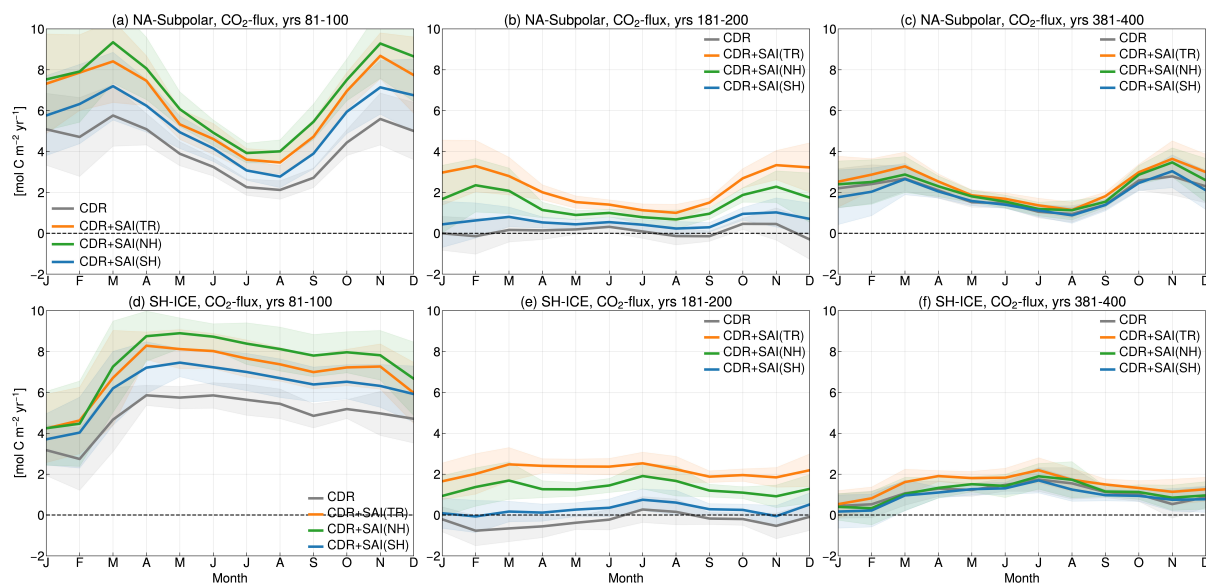


Figure A5. Mean seasonality of air-sea CO₂ fluxes for all experiments in the (a-c) subpolar North Atlantic (NA-Subpolar; 45°–75°N, 60°W–20°E) and (d-f) Southern Ocean (SH-ICE; >60°S), averaged for years 81–100, 181–200, and 381–400. Positive values depict ocean uptakes.

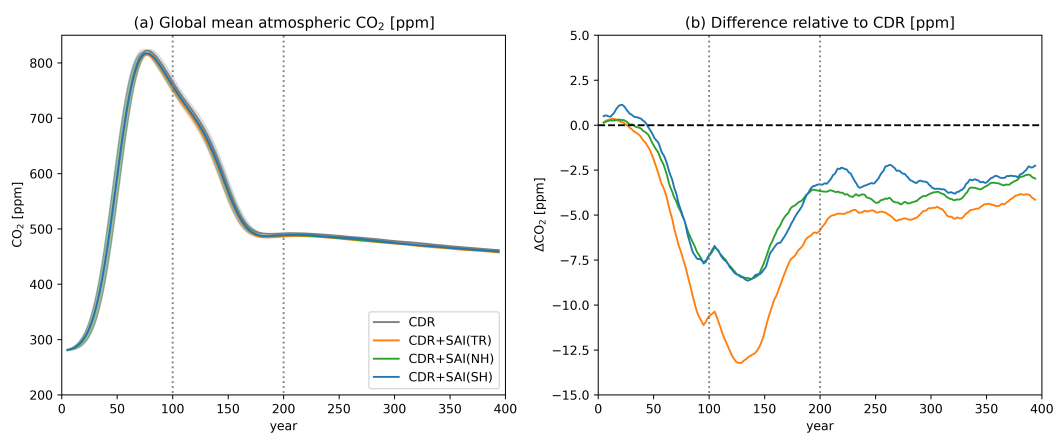


Figure A6. Timeseries of global mean (a) atmospheric CO₂ concentrations for all experiments and (b) the difference between experiments with SAI and baseline CDR experiment.



Code and data availability. The source code of NorESM2-LM is available at <https://doi.org/10.5281/zenodo.3905091> (Seland et al., 2020b).

610 The model data (CDR and CDR+SAI(XX)) generated in this study are available through the Norwegian Research Data Archive and can be accessed at <https://doi.org/10.11582/2022.00012> (Schwinger et al., 2022b) and <https://doi.org/10.11582/2026.kymtop3b> (Tjiputra and Olivié, 2026).

Author contributions. JT conceived the study. JT, DO, and JS designed and implemented the model experiments. All co-authors analyzed and interpreted the model data, and wrote the manuscript.

615 *Competing interests.* The authors declare no competing interests.

Disclaimer. The views, opinions, and practices used to produce this dataset/software are, however, those of the author(s) only and do not necessarily reflect those of the European Union or European Research Executive Agency. Neither the European Union nor the granting authority can be held responsible for them.

620 *Acknowledgements.* The authors were supported by the Research Council of Norway through the project NAVIGATE (no. 352142). JT, RF and JS acknowledge the project NorESM4CMIP7 funded through the Research Council of Norway (grant no. 352204). JT also acknowledges the project OceanICU (Horizon Europe, grant agreement no. 101083922). Super-computing and storage resources were provided by the Norwegian Research Infrastructure Services (projects nn1002k, ns1002k, nn10054k, and ns10054k). HL acknowledges the National Research Foundation (NRF) of Korea grant RS-2025-25428567.



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