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Delayed Southern Ocean heat release challenges climate stabilization policies

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Abstract

Net-zero CO₂ emissions are widely considered necessary and sufficient to stabilize global-mean temperature, although radiative equilibrium is reached only centuries to millennia after emissions cease. Until then, the ocean is expected to absorb most excess heat. Here we use an intermediate-complexity Earth system model to show that, after centuries of stable surface temperatures, Southern Ocean deep convection can abruptly release a substantial fraction of ocean-stored excess heat to the atmosphere. Continued subsurface warming erodes the stratified, cold and fresh surface layer from below, initiating convection. This heat release occurs across model configurations and emissions scenarios, except with implausibly weak ocean-interior mixing or implausibly high cumulative emissions. Despite sustained net-zero emissions, it abruptly raises global mean surface air temperature by 0.4 °C and temperatures near Antarctica by more than 2 °C. Ocean heat storage may therefore be less durable than assumed, and net-zero targets alone may not ensure long-term temperature stabilization.

1. Introduction

A key element of current climate policies is the target of “net-zero” emissions in order to arrest the increase in global-mean surface air temperature (SAT), as explicitly called for in the 2015 Paris Agreement (UNFCCC 2015). Its empirical base is the linear proportionality of the increase in SAT and cumulative emissions of anthropogenic CO₂ (Matthews et al., 2009; Allen et al., 2009). Such a linear relationship is supported by the historical record (Millar and Friedlingstein, 2018) and has, with some small variations caused primarily by non-CO₂ climate forcings, been confirmed in various simulations by a large number of climate models (Williams et al., 2020). It implies that any temperature target corresponds to a certain CO₂

emission budget and that global-mean SAT should no longer increase once net-zero CO₂ emissions are reached and maintained (MacDougall et al., 2020; Allen et al., 2022).

The near constancy of global-mean SAT under net-zero emissions despite a remaining radiative imbalance caused by legacy anthropogenic CO₂ in the atmosphere relies on the ocean taking up anthropogenic heat and carbon in just the right proportion (Goodwin et al., 2015). Expected consequences are committed long-term future environmental impacts on the ocean, such as warming, acidification, deoxygenation and thermohaline sea level change (Solomon et al., 2009; Clark et al., 2016; Oschlies, 2021; Palazzo Corner et al., 2023).

Modeling studies investigating climate impacts of net-zero emissions typically cover multi-decadal to centennial timescales (MacDougall et al., 2020). Much longer timescales have been addressed primarily with Earth system models of intermediate complexity (e.g., Archer et al., 2009; Eby et al., 2009; Solomon et al., 2009; Frölicher and Paynter, 2015; Ehlert and Zickfeld, 2017; Oschlies, 2021), but increasingly also with state-of-the-art Earth system models (Frölicher et al., 2020; King et al., 2024). Interestingly, many of these millennial-scale model simulations show signatures of intermittent abrupt warming events with global SAT increases by several tenths of a degree long after anthropogenic CO₂ emissions have ceased. These sudden warming events have not yet been studied in detail, but were typically linked to the emergence of Southern Ocean deep convection (Archer et al., 2009; Eby et al., 2009; Frenger et al., 2025). We here present a first comprehensive analysis of the underlying mechanisms and their controls using an intermediate complexity Earth system model (Weaver et al., 2001; see Methods), examine the sensitivity to ocean-mixing parameters and emission scenarios and discuss implications for long-term net-zero scenarios and responsible climate policies.

2. Results

2.1 Sudden atmospheric warming under net-zero emissions

Consistent with theory and modeling studies (Allen et al., 2022), an idealized simulation driven by historical CO₂ emissions until 2020 and zero emissions thereafter shows SAT increasing by about 1.1 °C until 2020 and remaining essentially constant thereafter (Fig. 1b). This occurs despite declining atmospheric pCO₂ and continued net heat uptake under a positive top-of-atmosphere (TOA) radiation balance (Fig. 1a). After a few centuries of stability, with variations below 0.04 °C, simulated global-mean SAT raises abruptly by more than 0.4 °C within a few decades after 2650, despite sustained net-zero emissions. Subsequently, SAT declines somewhat, but remains about 0.2 °C above its pre-warming level. Global-mean sea surface temperature (SST) follows a similar trajectory, but with smaller amplitude (Fig. 1b).

Before the abrupt warming, nearly all excess energy associated with the positive TOA radiation balance is absorbed by the ocean. This uptake stabilizes SAT after emissions cease, while subsurface ocean temperatures continue to rise (Fig. 1c). Around year 2235, the increase in global-mean thermocline warming at 500 m exceeds surface warming by 0.1 °C. Although initial warming rates generally decline with depth, the laterally averaged warming becomes increasingly uniform and exceeds 0.5 °C at all depths by 2650 (Fig. 1c).

From year 2020 to 2650, the model ocean absorbs 2.6 YJ (2.6×10^{24} J), nearly three times the heat absorbed during the historical period from 1800 to 2020 (0.89 YJ; Fig. 2a). The majority of this excess heat uptake takes place in the southern hemisphere, in agreement with observations (Cheng et al., 2024) and modeling studies (Huguenin et al., 2022). Simulated storage of excess heat remains positive in both hemispheres, with a largely stable interhemispheric partitioning throughout the simulation: 60 $\pm$ 3% is stored in the southern hemisphere, closely matching its $\sim$ 60% share of global ocean volume. Different interhemispheric shares of anomalous surface heat flux and heat storage are connected via an anomalous northward heat transport within the ocean (Fig. 2b). During the period 2020 to 2650, the transport reaches a maximum of more than 0.25 PW at 44 $^{\circ}$ S, while showing a gradual increase over time in the northern hemisphere between 15 $^{\circ}$ N and 55 $^{\circ}$ N and a slight weakening between 15 $^{\circ}$ S and 55 $^{\circ}$ S.

After 2650, a complete reorganization of the anomalies in ocean heat storage, meridional heat transport and overturning circulation occurs: The anomalous meridional oceanic heat transport reverses from northward to southward, initially south of 60 $^{\circ}$ S, then propagating northward to reach the northern hemisphere mid-latitudes within decades (Fig. 2b). This reversal coincides with a rapid intensification of the Southern Meridional deep Overturning Circulation (SMOC), whose volume transport more than doubles within 20 years (Fig. 2c). This occurs after an abrupt increase in Southern Ocean maximum mixed layer depths drives buoyancy loss, increases the meridional pressure gradient across the Antarctic Circumpolar Current (ACC), and accelerates the ACC by about 20% (Fig. 2d), consistent with other models (Pedro et al., 2016). A few decades after the increase in the SMOC, the Atlantic meridional overturning circulation (AMOC) begins to weaken, eventually by about 25% by 3000 (Fig. 2c). Between 2650 and 3000, the ocean releases 0.82 YJ to the atmosphere, lowering steric sea level by 9 cm (Figure 2a) and nearly matching the historical heat uptake of 0.89 YJ during 1800–2020.

2.2 Ocean warming patterns

During the historical period of rising anthropogenic emissions and global mean SAT, zonally averaged oceanic temperature shows warming from the surface, strongest in low- and mid-latitude waters, and limited warming in polar regions where sea ice acts as a thermostat (Fig. 3a). Subsurface warming exceeding 0.25 $^{\circ}$ C extends below 2000 m only in the Southern Ocean and in the North Atlantic Deep Water (NADW) formation region around 60 $^{\circ}$ N (Fig. S1). By 2020, simulated upper-ocean salinity decreases at high latitudes and increases in the subtropics (Fig. 3d), in good agreement with observations (Gould and Cunningham, 2021) and linked to an intensification of the hydrological cycle (Held and Soden, 2006).

Over this period, both the clockwise northern hemispheric overturning associated with NADW and the anticlockwise SMOC associated with the northward Antarctic Bottom Water (AABW) flow weaken (Figs. 3a, 2c). The simulated AMOC declines by 15%, from 18.1 Sv in 1800 to 15.7 Sv in 2020, while the northward AABW transport in the Atlantic decreases from 3.1 Sv at 16 $^{\circ}$ N in 2000 to 2.8 Sv in 2020, both in good agreement with observational estimates (Jackson et al., 2022; Bilo et al., 2024; Li et al., 2026).

After net-zero emissions are imposed in 2020, upper-ocean temperatures change little, consistent with stabilized SAT, whereas subsurface warming exceeds 0.5 °C across large deep-ocean regions by 2650 (Fig. 3b). The AMOC begins strengthening shortly after 2020, recovers by 2250, and exceeds the pre-industrial level by over 10% by 2650 (Fig. 2c), as also seen in other net-zero simulations (MacDougall et al., 2022). The stronger AMOC and persistently weak SMOC promote accumulation of warmer and saltier deep waters (Figs. 3b,e), especially in the Atlantic (Fig. S2).

After 2650, zonally averaged temperature and salinity anomalies change substantially (Fig. 3c): Deep-ocean temperatures decline by up to 1 °C, particularly in the southern hemisphere, indicating enhanced ventilation from the south. These waters freshen the deep ocean (Fig. 3f), while Subantarctic Mode and Intermediate Waters get warmer and saltier in the upper 1000 to 2000 m, especially in the Indian and Pacific Oceans (Fig. S3). Ocean heat loss to the atmosphere between years 2650 and 3000 originates primarily below 1500 m (Figs. 1c, 3c, S4).

2.3 Mechanisms

The rapid deepening of mixed layers south of the ACC after 2650 (Fig. 2d) together with patterns of SAT increase (Fig. S5) and deep-ocean cooling (Fig. 3c) identifies Southern Ocean deep convection as likely driver of the abrupt release of deep-ocean heat to the atmosphere. In the following we therefore focus on waters south of 60°S, where zonally averaged SST increases by less than 0.1 °C until 2020 (Fig. 4a), tenfold less than the global-mean SST increase (Fig. 1), and in general agreement with observations that even show a slight cooling in recent decades (Haumann et al., 2020). Southern Ocean SSTs increase by another 0.05 °C between 2020 and 2650. In contrast, subsurface warming is more than fivefold greater by 2020, about twice the historical global-mean warming rate at 1000 m (Fig. 1c; Gille, 2002), and exceeds 0.7 °C from 1000 m to the sea floor by 2650 (Figure 4a). This warming reduces density by more than 0.08 kg/m³ (referenced to 2000 dbar) between 1000 and 4000 m depth (Figure 4c), which alone would correspond to a vertical displacement of more than 3000 m in the weakly stratified Southern Ocean.

Simulated Southern Ocean surface salinities freshen by -0.008 psu/decade until 2020, at the low end of observed rates (-0.011 +- 0.004 psu/decade; de Lavergne et al., 2014), and change little from 2020 to 2650 (Fig. 4b). In the model, roughly equal contributions from an intensified hydrological cycle and reduced equatorward sea-ice export cause this freshening (Fig. S6), which strengthens surface stratification (Fig.4d). Subsurface salinity increases through weakened mixing of warmer and saltier waters into the surface layer (Olivier and Haumann, 2025) and enhanced replacement by saltier northern-sourced waters at depth (Fig. 3e), but only for slightly more than a century after emissions cease (Fig. 4b), whereas subsurface warming continues until 2650 (Fig. 4a).

Deep convection begins with an increase in surface heat loss, declining sea ice near the Kerguelen Plateau (~90°E) and close to the Antarctic coast in 2665 (Fig. S7). This region also initiates deep convection in a number of sensitivity experiments with different mixing parameterizations and emission scenarios. Topographic control, a cyclonic Weddell Sea gyre to the west and a smaller cyclonic recirculation to the east favour subsurface heat

accumulation. Within two decades, convective activity, heat loss and sea-ice decline propagate westward into the Weddell Sea near 45°W, resembling the observed westward progression of historical polynyas from Maud Rise to the Weddell Sea (Cheon & Gordon, 2019). Consistent with simulations of Weddell Sea polynya onset (Rheinlaender et al., 2021), elevated mixing first develops at mid-depth between about 1500 and 2000 m (Fig. S8) where subsurface warming peaks and erodes stratification from below (Fig. 4a,c).

Once initiated, deep convection and elevated ocean heat loss persist for decades to centuries (Fig. 2d). Sea ice disappears year-round between 60°E and 115°E and between 120°E and 145°E (Fig. S7c), but remains seasonal elsewhere around Antarctica. Annual-mean ocean heat loss reaches more than 100 Wm⁻² in convective regions south of 60°S (Fig. S7b).

2.4 Controls

To investigate controls on subsurface heat accumulation, erosion of stratification from below and the onset of Southern Ocean deep convection, we constructed two dedicated model ensembles. The ‘scenario ensemble’ branches off from an extended high-emission scenario (Methods) and switches to zero emissions at different years between 1990 and 2300 (Fig. S9a). All simulations show multi-centennial periods of near-constant global SAT after reaching net zero (Fig. S9c), with Southern Ocean SST increases (Fig. 5b) much smaller than global SAT increases. They also show continued subsurface ($z = 2000$ m) warming south of 60°S after emissions cease (Fig. 5c). In simulations that reach net-zero emissions by 2100, this subsurface warming exceeds Southern Ocean surface warming, eroding stratification from below and ultimately triggering deep convection. By contrast, the 2150 and 2300 net-zero simulations show no deep convection. In these cases, Southern Ocean SST continues to increase after emissions cease and outpaces subsurface warming, thereby strengthening stratification. Although deep convection does not occur within the integration period, the SMOC gradually strengthens a few centuries after emissions cease, partially recovers (Fig. S10b), and arrests subsurface warming well below surface warming (Fig. 5). The AMOC does not subsequently weaken in these simulations (Fig. S10a).

The temperature-salinity diagram (Fig. 5a) illustrates that surface freshening dominates the buoyancy gain of Southern Ocean surface waters under anthropogenic CO₂ emissions until the mid 21st century (Fig. 5a, grey arrow pointing to the left). With increasing cumulative emissions, however, surface warming becomes increasingly relevant for the surface buoyancy gain: It remains below 0.1 °C by 2020, but reaches 0.6 °C by 2100 and 2.6 °C by 2300 under continued high emissions (Figs. 5a,b). Antarctic sea ice declines correspondingly by 2%, 19% and 60%, respectively (Fig. S10c). In contrast, Southern Ocean deep waters become lighter almost exclusively by warming (Fig. 5a, grey arrow pointing upward), with a tendency, but no strict relationship, towards greater subsurface warming at higher cumulative emissions (Fig. 5c).

Deep convection begins a few hundred years after emissions cease in all simulations where it occurs, suggesting a leading role for diffusion rather than the more rapid response expected from waves or advection. We therefore performed a ‘mixing ensemble’ (Methods), varying background diffusivity from $k_v = 0.01$ to 0.3 cm²/s, with $k_v = 0.15$ cm²/s being used in all other simulations.

The following picture emerges: For identical cumulative emissions, stronger background mixing leads to faster and more intensive warming of the ocean interior and deep Southern Ocean (Fig. 5e), whereas pre-convection Southern Ocean surface temperatures (Fig. 5d) and global mean SAT (Fig. S11) are relatively insensitive to variations in k_v . Higher ocean-interior mixing results in an earlier onset of Southern Ocean deep convection and increases the associated abrupt warming of Southern Ocean surface waters and global SAT. No deep convection occurs for the lowest vertical mixing at $k_v = 0.01 \text{ cm}^2/\text{s}$ (not even by year 5000), whereas all other runs undergo abrupt Southern Ocean deep convection and excess heat release. The onset ranges from year 3410 for $k_v = 0.05 \text{ cm}^2/\text{s}$ to 2480 for $k_v = 0.3 \text{ cm}^2/\text{s}$. Ocean-interior mixing is thus a key control on the timing and intensity of Southern Ocean deep convection and the release of transiently stored ocean heat.

3. Discussion

Contrary to the prevailing expectation that global mean temperature stabilizes once net-zero emissions are reached, our simulations reveal an abrupt warming of several tenths of a degree within centuries after emissions cease. This response occurs consistently across scenarios that reach net zero by 2100 or earlier, corresponding to cumulative emissions of up to 8,500 Gt CO_2 , which was also the threshold for sudden warming in an emission-pulse ensemble (Eby et al., 2009). Nearly all scenarios used in current Climate Model Intercomparison Projects fall below this threshold; only high-emission scenarios generally considered implausible exceed it (Hausfather and Peters, 2020; van Vuuren et al., 2026). Southern Ocean deep convection triggers the warming by releasing heat comparable to all excess heat absorbed by the global ocean to date. Warming exceeds 2°C near Antarctica, potentially accelerating ice-sheet melt and associated cascading responses not represented in the current model.

Deep convection requires heat accumulation beneath a stratified Southern Ocean surface layer. During the historical period, surface freshening (Fig. 4) strengthens stratification and weakens upwelling of warmer, saltier waters (Haumann et al., 2020; Olivier and Haumann, 2025). Meanwhile, deeper layers associated with Circumpolar Deep Water warm faster and more strongly than the surface, eroding stratification from below until convection begins. Across both model ensembles, deep convection is avoided only with extremely weak ocean-interior mixing ($k_v = 0.01 \text{ cm}^2 \text{ s}^{-1}$) or extremely high cumulative emissions, when roughly half of the southern hemisphere sea ice is lost and Southern Ocean surface warming outpaces warming at depth (Fig. 5). This is in line with another intermediate-complexity model study, which simulated post-emissions deep-ocean flushing under present-day conditions with seasonal sea ice, but not under ice-free Palaeogene conditions (Ridgwell and Schmidt, 2010).

Deep convection strengthens the ACC and SMOC, shifts deep-water formation from predominantly northern to predominantly southern sources that flush the deep ocean with cold and fresh waters. As these bottom waters spread northward, they induce rapid cooling throughout the Atlantic to the Greenland–Scotland Ridge. Following centuries of gradual warming, cold bottom-water temperatures decrease by more than 1°C within a decade in the southern and equatorial Atlantic, with a weaker and delayed decline north of the

equator (Fig. 6). Rapid cooling is accompanied by increased dissolved oxygen (Oschlies, 2021) and pH (not shown). Thus, even full implementation of current climate policies and net-zero emissions could substantially perturb surface and deep-ocean environments.

These results question the durability of ocean storage of excess heat, on which the post-net-zero SAT stabilization—and, therefore, the adequacy of current net-zero targets—depends. They also challenge our understanding of ocean heat storage and redistribution following temporary greenhouse-gas emissions. In multimillennial climate-model simulations forced by a quadrupling of atmospheric CO₂, warming ranges from near-uniform throughout the water column (Li et al., 2013), resembling our simulation in 2650 (Fig. S4), to surface-intensified warming (Yamamoto et al., 2018), as in our simulation by 3000.

Simulations extended by another 4,000 years show enhanced Southern Ocean deep mixing and heat loss persisting for approximately 1,000 years, until the deep-ocean excess-heat reservoir is depleted. The deep ocean north of the ACC then cools, weakening the meridional density gradient across the ACC and reducing the strength of the ACC and SMOC, until the AMOC again dominates global overturning. Southern Ocean stratification subsequently re-establishes, NADW formation becomes more important, and a stronger AMOC temporarily accumulates heat at depth (Figs. S12 and S13). Deep-water formation consequently oscillates between the North Atlantic and Southern Ocean at approximately 1,400-year intervals, similar to behavior previously reported for the UVic model (Meissner et al., 2008) and fully coupled climate models (Martin et al., 2013). In these simulations, deep-ocean warming is transient, with excess heat being released during subsequent Southern Ocean deep convection episodes. By contrast, the extended simulation reaching net zero in year 2300 shows no deep convection and retains surface-intensified warming (Fig. S13b). Vertically homogeneous ocean warming is therefore not a stable response to anthropogenic CO₂ emissions in our model.

Reaching and maintaining net-zero CO₂ emissions is widely regarded as necessary and sufficient to stabilize global mean SAT. This expectation underpins current climate policy. Our results support the suitability of net-zero targets for near-term temperature stabilization, but indicate that a substantial fraction of excess heat stored in the ocean could eventually be released rapidly, causing abrupt surface warming, a reorganization of global ocean circulation and deep-ocean environmental change. Our results indicate that the threat of delayed release of excess heat via Southern Ocean deep convection renders net-zero emission targets insufficient for long-term climate stabilization. Understanding the durability of oceanic storage of excess heat is therefore essential for responsible long-term climate policy.

Methods

The University of Victoria (UVic) Earth System Model of intermediate complexity version 2.8 (Weaver et al., 2001) was used in the configuration described by Oschlies, 2021. It was spun up for more than 10,000 years under pre-industrial atmospheric CO₂ (280 ppm), then forced with historical CO₂ emissions from year 1765 until 2010, emissions corresponding to the RCP 8.5 high-emission scenario (Nakicenovic et al., 2000) until 2020, and zero emissions

thereafter until the end of the simulation in year 3000. Simulations in the ‘scenario ensemble’ use the same emission forcing but branch off to net-zero emissions in different years. Some simulations were continued under net-zero emissions until year 7000.

In the ‘mixing ensemble’, the runs differ only in the employed background vertical diffusivities, taking on values of $k_v = 0.01, 0.05, 0.10, 0.15, 0.2, 0.3 \text{ cm}^2/\text{s}$, respectively. This range covers values commonly employed in general ocean circulation models (Duteil and Oschlies, 2011) and observational estimates for ocean-interior mixing (Ledwell et al., 1998; Kunze et al., 2006), with $k_v = 0.15 \text{ cm}^2/\text{s}$ being used in all other simulations in this study. Stemming from a series of model simulations performed earlier, the runs of this ensemble branch off from the historical emissions to net zero emissions in 2010. In addition to the background vertical diffusivity, all runs also employ a bottom-intensified representation of tidally driven mixing, k_{tidal} (Simmons et al., 2004), and a vertical diffusivity of $k_{SO} = 1 \text{ cm}^2/\text{s}$ applied everywhere south of 40°S in order to account for unresolved intense small-scale mixing processes in the Southern Ocean (Naveira Garabato et al., 2004; Saenko, 2008). The total vertical diffusivity applied at any grid point and time step is $k_{tot} = k_v + k_{tidal} + k_{SO}$. In the stratified ocean interior away from rough bottom topography and north of 40°S , the background vertical diffusivity k_v is generally the dominant component.

Factors not included in our intermediate-complexity Earth system model affecting Southern Ocean stratification are changes in wind forcing and meltwater input from the Antarctic ice sheet. Apart from a temperature-based regional geostrophic wind correction (Weaver et al., 2001; Fig. S14), the current model employs prescribed wind fields, which, for all simulations above are taken as monthly climatological means over the second half of the 20th century (Kalnay et al., 1996). Model runs and reanalysis products using winds that specifically include a projected intensification and poleward shift of the southern hemispheric westerly winds (Getzlaff et al., 2016; Narayanan et al., 2026) display enhanced upwelling of warmer and saltier waters that reduce the stratification of the surface layer of the Southern Ocean and may also reduce the accumulation of subsurface heat. Fully coupled Earth system models that dynamically simulate winds, show both a sudden intensification of Southern Ocean deep convection (Yamamoto et al., 2015; Frölicher et al., 2020) and a weakening (King et al., 2024). Future studies will need to explore the impact of changing winds as well as meltwater input from the Antarctic ice sheet onto the occurrence of Southern Ocean deep convection.

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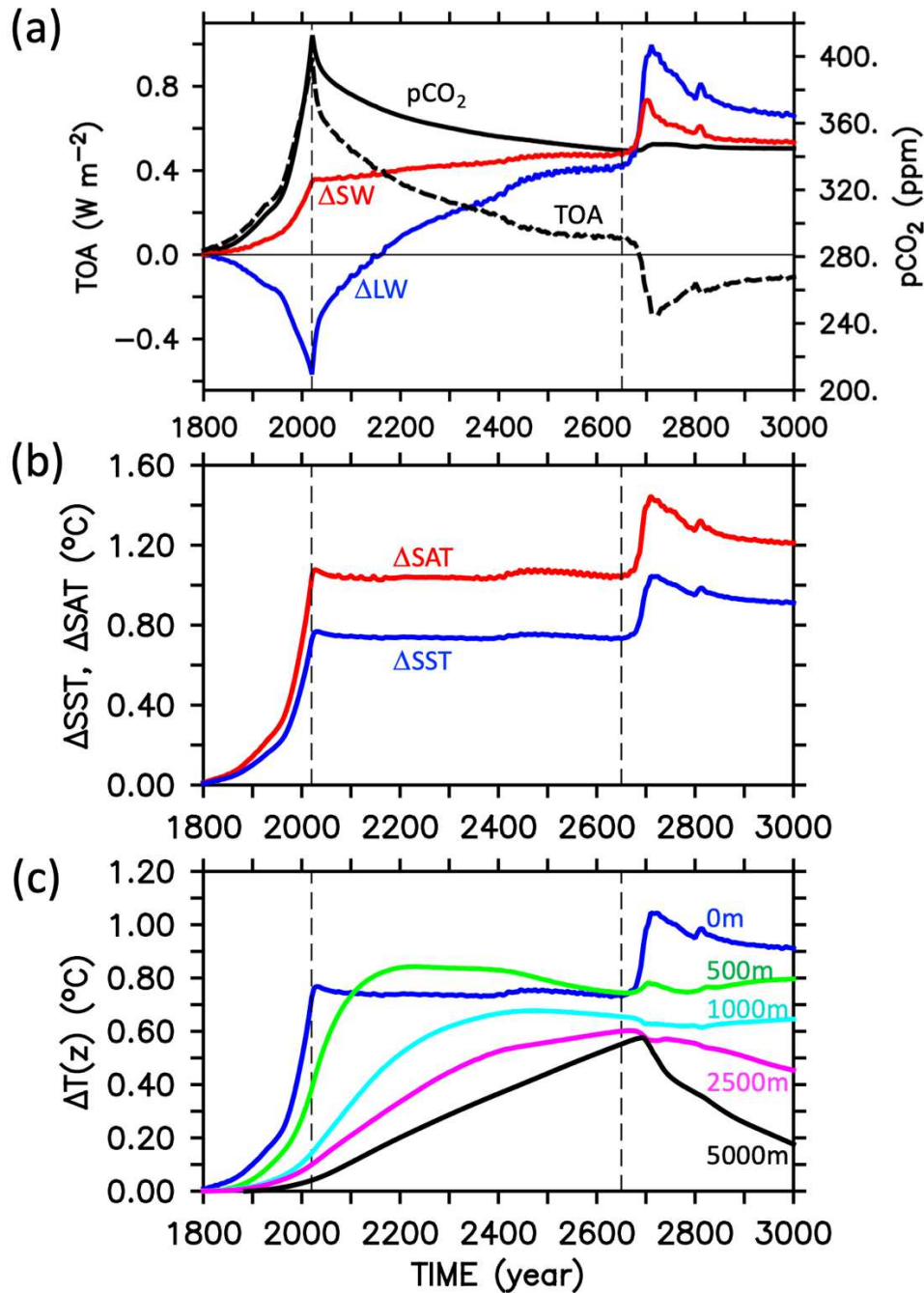


Figure 1. (a) Simulated atmospheric pCO_2 (black solid) and anomalies with respect to year 1800 of Top-of-Atmosphere (TOA) radiation (black dashed), net incoming short wavelength radiation (ΔSW , red) and outgoing long wavelength radiation (ΔLW , long). Anomalies with respect to year 1800 of (b) global mean surface air temperature (ΔSAT) and sea surface temperature (ΔSST) and (c) globally averaged ocean potential temperatures at depth of 0 m (blue), 500 m (green), 1000 m (cyan), 2500 m (pink) and 5000 m (black). Vertical dashed lines denote the stop of emissions in year 2020 and the year 2650 prior to the onset of Southern Ocean deep convection.

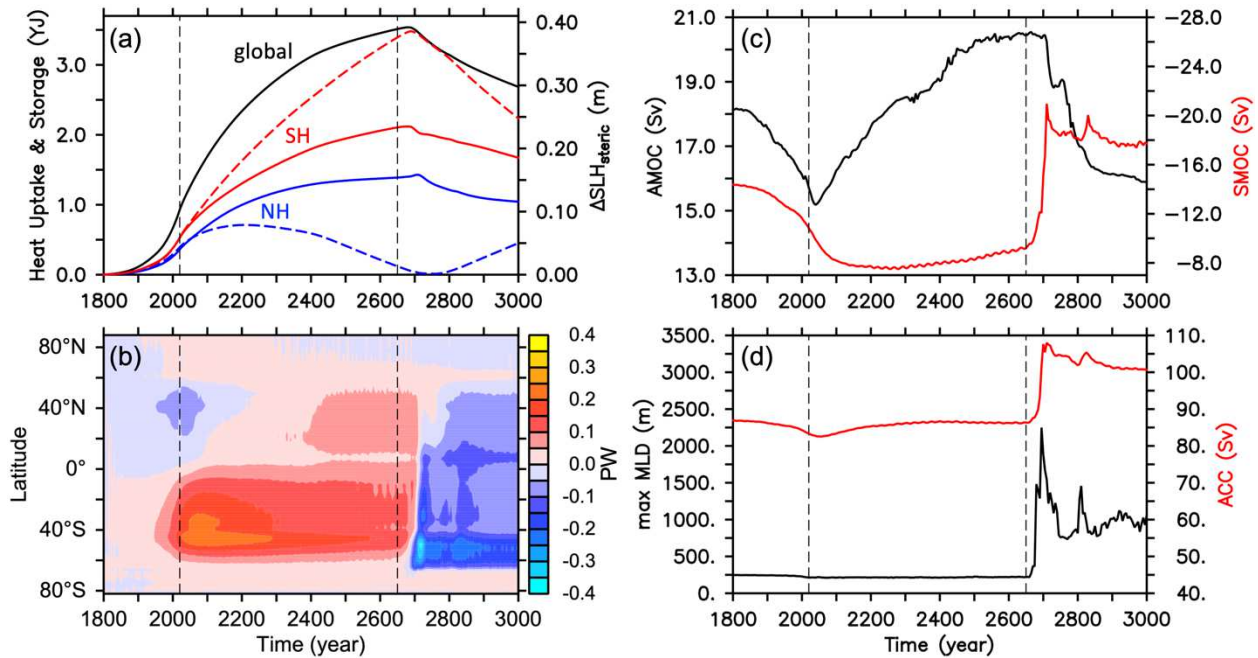


Figure 2. (a) Ocean heat storage relative to year 1800 globally (black) and in the northern (blue) and southern (red) hemisphere. Cumulative anomalous air-sea heat flux with respect to year 1800 in each hemisphere is denoted by the dashed lines. Units are YJ (10^{24} J). The corresponding thermosteric sea level change is indicated on the right axis. (b) Hovmöller plot of zonally and vertically integrated northward oceanic heat transport relative to year 1800 in PW. (c) Time series of the maximum (within 200 and 4000 m, 26° N and 71° N) of the Atlantic Meridional overturning stream function (black) and negative maximum (within 200 and 5500 m, 86° S and 44° S) of the Southern Ocean overturning stream function (red). Both stream functions are integrated zonally along geopotential surfaces and include the bolus velocity. (d) Time series of annually averaged maximum mixing depth south of 60° S (black), which is found equivalent to the maximum mixing depth in the southern hemisphere, and volume transport through Drake Passage (red). Vertical dashed lines in all panels denote years 2020 and 2650.

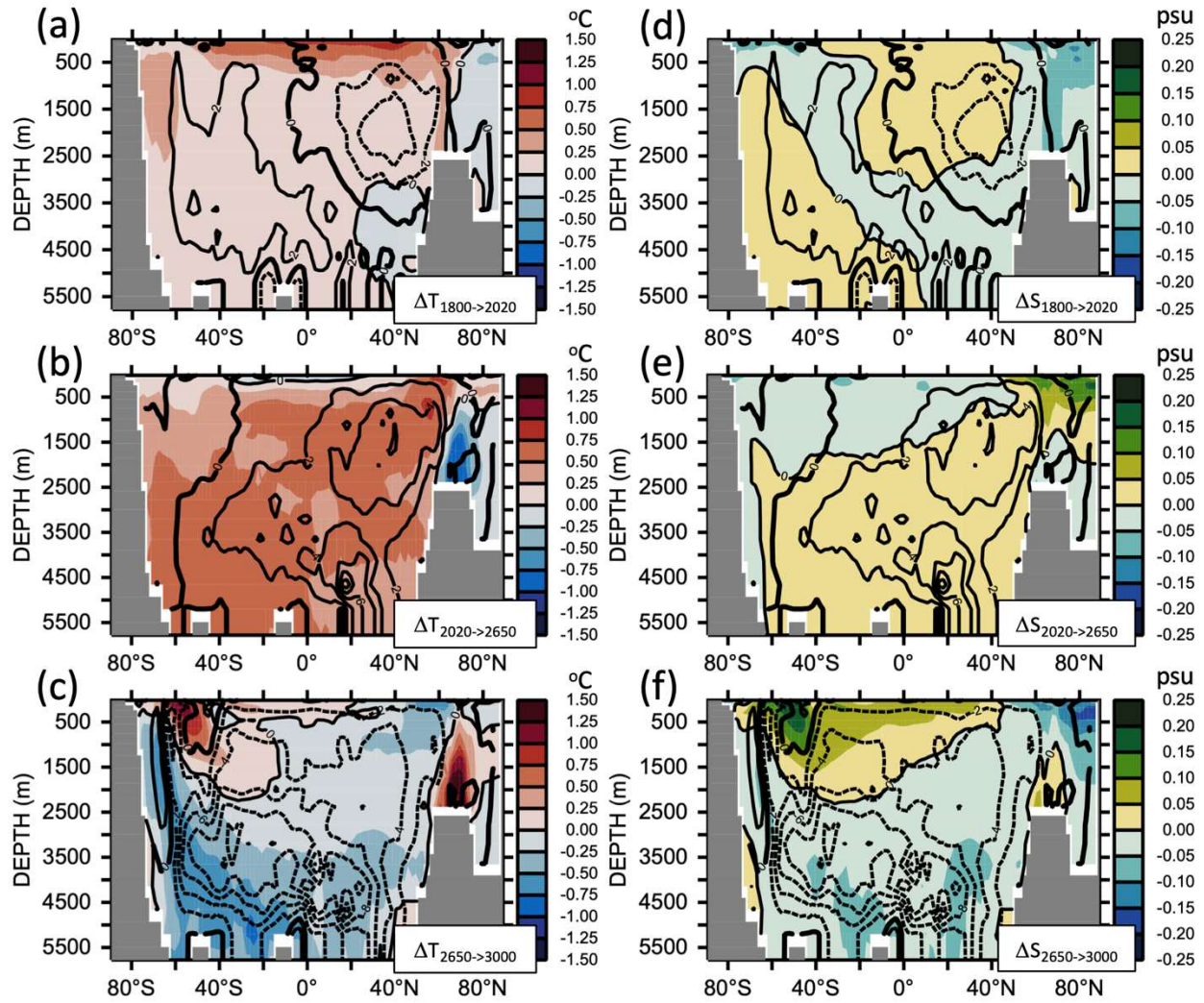


Figure 3. Zonally averaged potential temperature (left) and salinity (right) difference for (a,d) year 2020 minus 1800, (b,e) year 2650 minus 2020, and (c,f) year 3000 minus 2650. Contours denote the corresponding temporal changes in the global meridional overturning of the Eulerian plus bolus velocities, contour interval is 2 Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$).

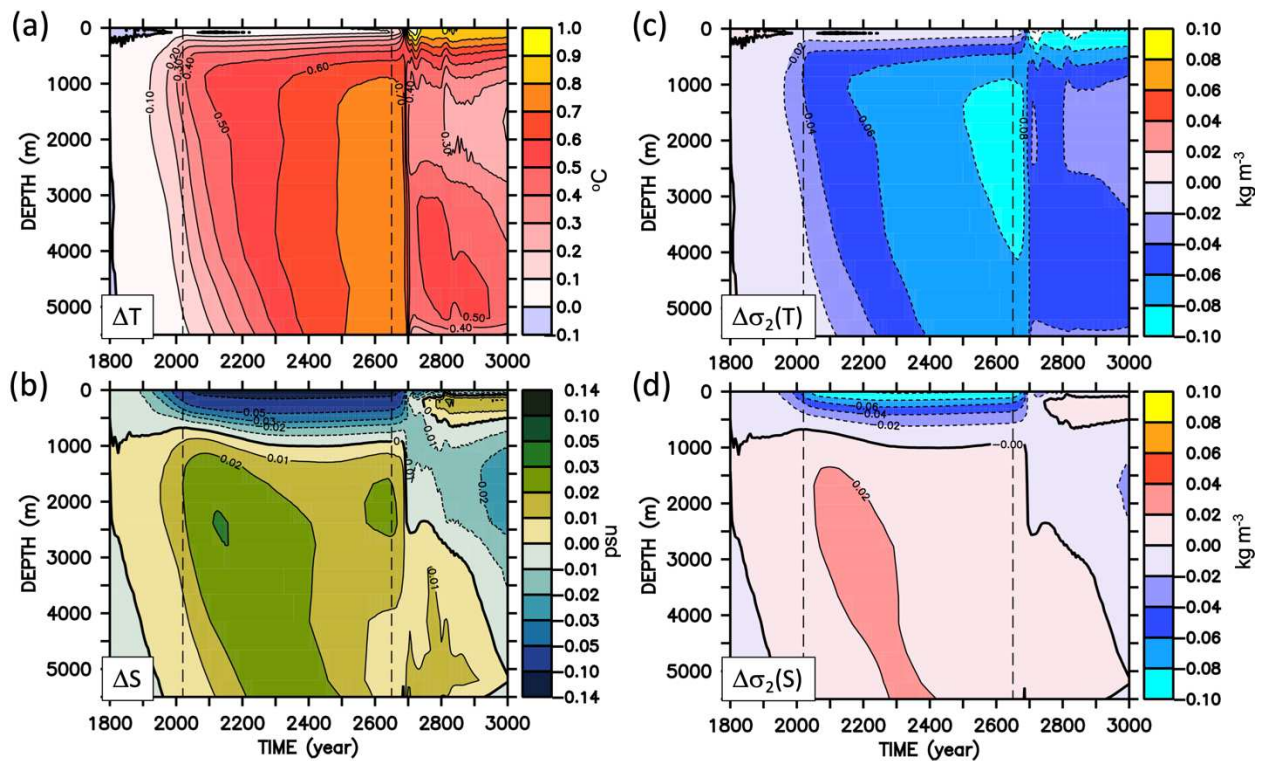


Figure 4: Temporal evolution with respect to year 1800 of (a) potential temperature, (b) salinity anomalies with respect to year 1800, (c) temperature-driven part of potential density, and (d) salinity-driven part of potential density, all averaged over the Southern Ocean south of 60°S . Units are degrees Celsius (a), practical salinity units (b), kg m^{-3} (c,d). Vertical dashed lines indicate the beginning of the net-zero emission period in year 2020 and year 2650 before the onset of Southern Ocean deep convection.

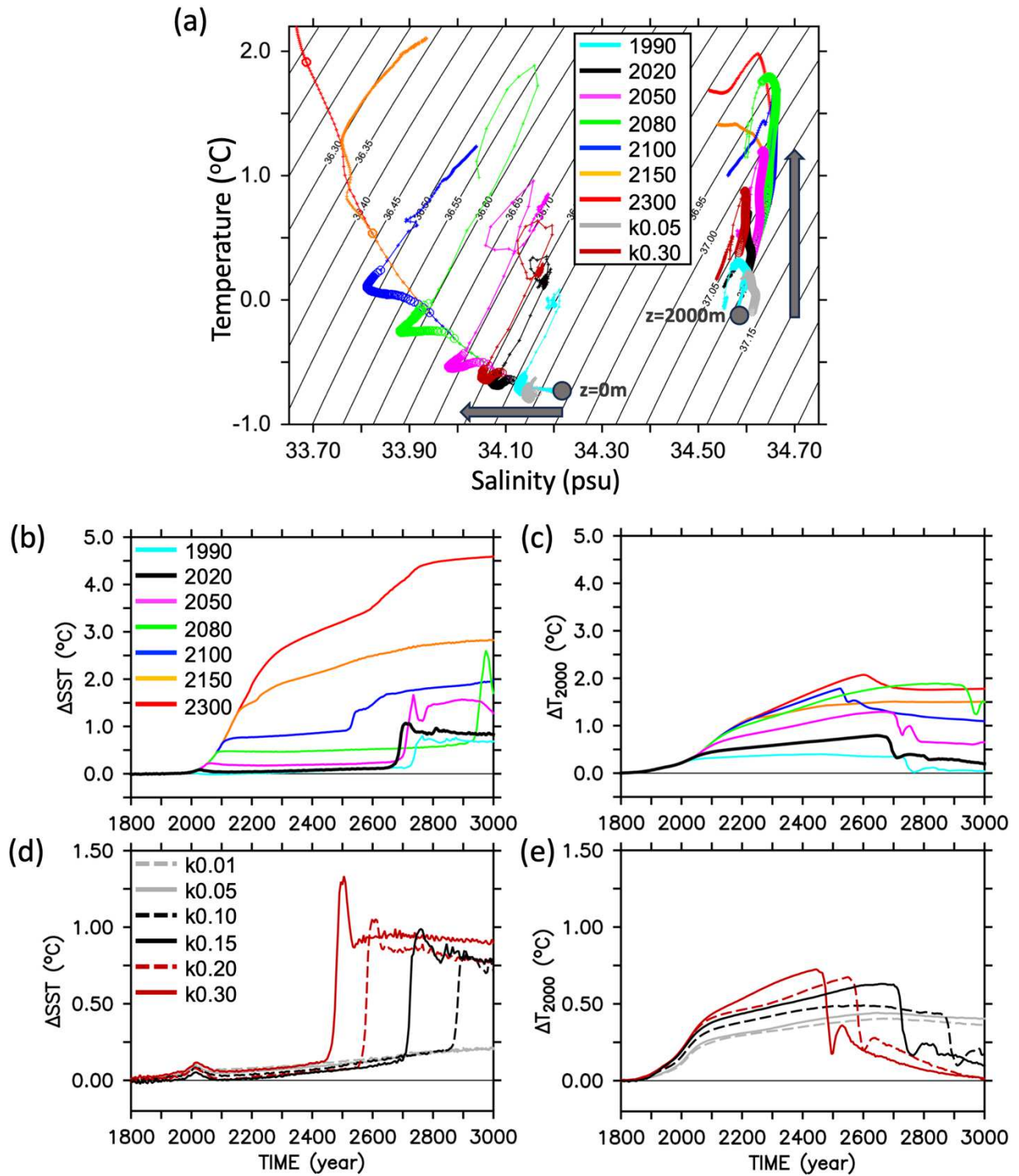


Figure 5: (a) Temperature-salinity (TS) diagram displaying simulated annual mean temperatures and salinities of the surface layer and at 2000m depth averaged laterally over the region south of 60°S. Shown are annual means every 5th year from 1800 to 3000. Year 1800 is at the lower end of each ensemble of trajectories marked by a solid grey circle for the surface layer (bottom) and the deep waters at z=2000m (right), respectively. For each model simulation, large circles on the trajectory mark the time of switching to net-zero emissions and continue, if existent, until the

onset of deep convection, e.g., from year 2020 to year 2650 in the standard run with emission stop in year 2020. After the onset of the deep convection event, trajectories of surface and deep TS properties continue as thin lines with small dots. Potential densities (σ_2) are referenced to 2000 dbar and labeled with units kg/m^3 . The thick grey arrow at the bottom indicates a density decrease occurring by surface freshening, whereas the thick grey arrow on the right indicates a density decrease by the same amount driven by warming at $z = 2000$ m. Lengths of the arrows correspond to the surface density changes of the simulation with emissions stop in year 2050 (pink circles) until the onset of the deep convection event. Results of simulations labeled k0.05 ($k_v = 0.05 \text{ cm}^2/\text{s}$) and k0.30 ($k_v = 0.30 \text{ cm}^2/\text{s}$) are included as grey and dark red trajectories. (b) Simulated changes in sea surface temperature and (c) potential temperature at $z = 2000$ m for simulations of the scenario ensemble, (d) changes in sea surface temperature and (e) potential temperature at $z = 2000$ m for simulations of the mixing ensemble, all averaged south of 60°S and referenced to the pre-industrial state in year 1800. Different colors refer to simulations of the scenario ensemble switching from the extended high-emission RCP 8.5 scenario to net-zero emissions in different years indicated by the legend in panel (b) and for different background vertical diffusivities given in cm^2/s in panel (d). The black lines show the standard experiment with background diffusivity of $0.15 \text{ cm}^2/\text{s}$ and zero emissions starting in year 2020 in (b) and (c) and, to align with the experimental design of the mixing ensemble, in year 2010 in (d) and (e). Note that vertical scales are identical in (b) and (c) as well as in (d) and (e).

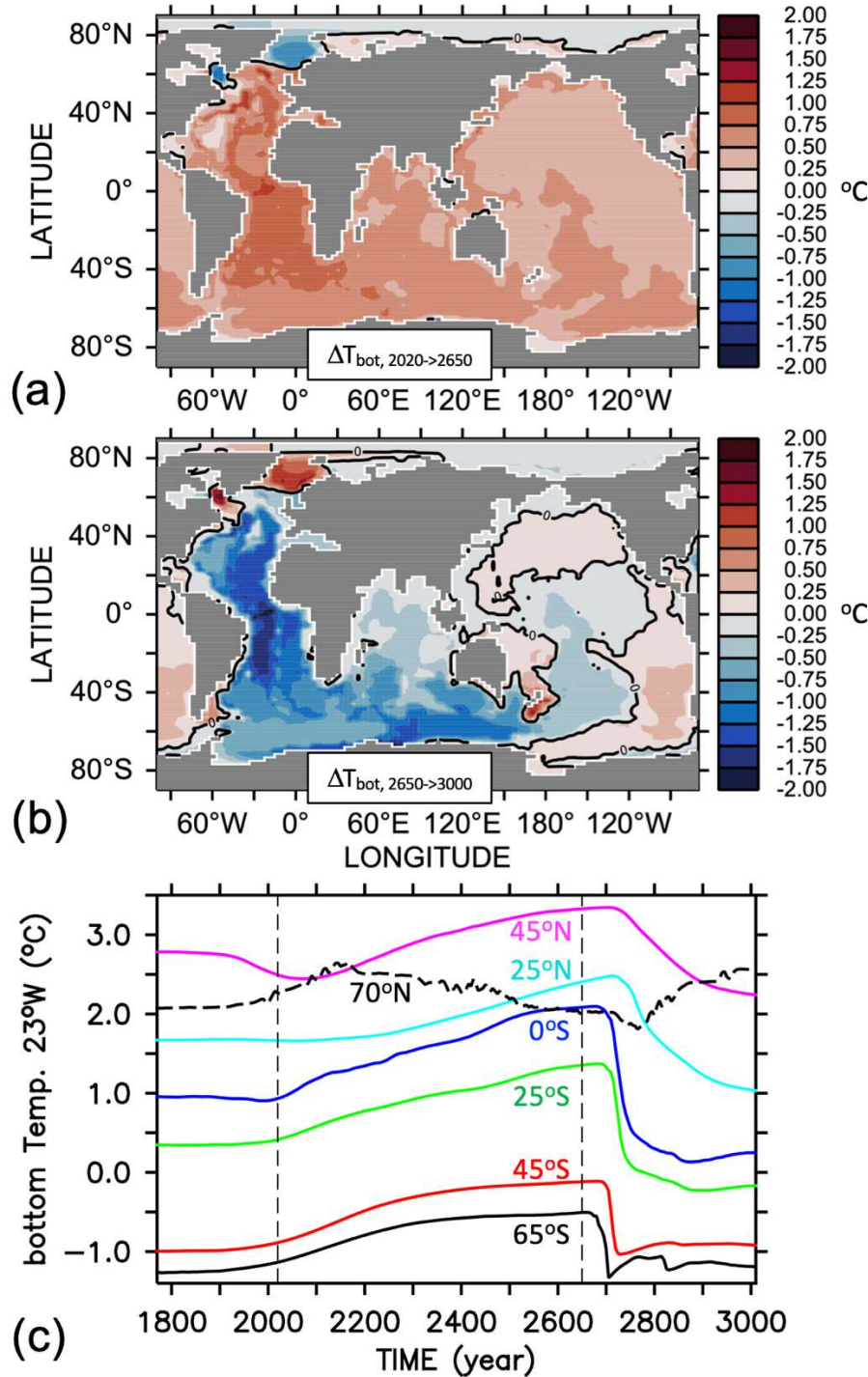


Figure 6: Simulated evolution in bottom temperature (a) between the net-zero emission years 2020 and 2650, (b) between years 2650 and 3000, and (c) at selected latitudes indicated by labels along 23°W. Vertical dashed lines in (c) refer to year 2020 when emissions stop and year 2650 prior to the deep convection event.

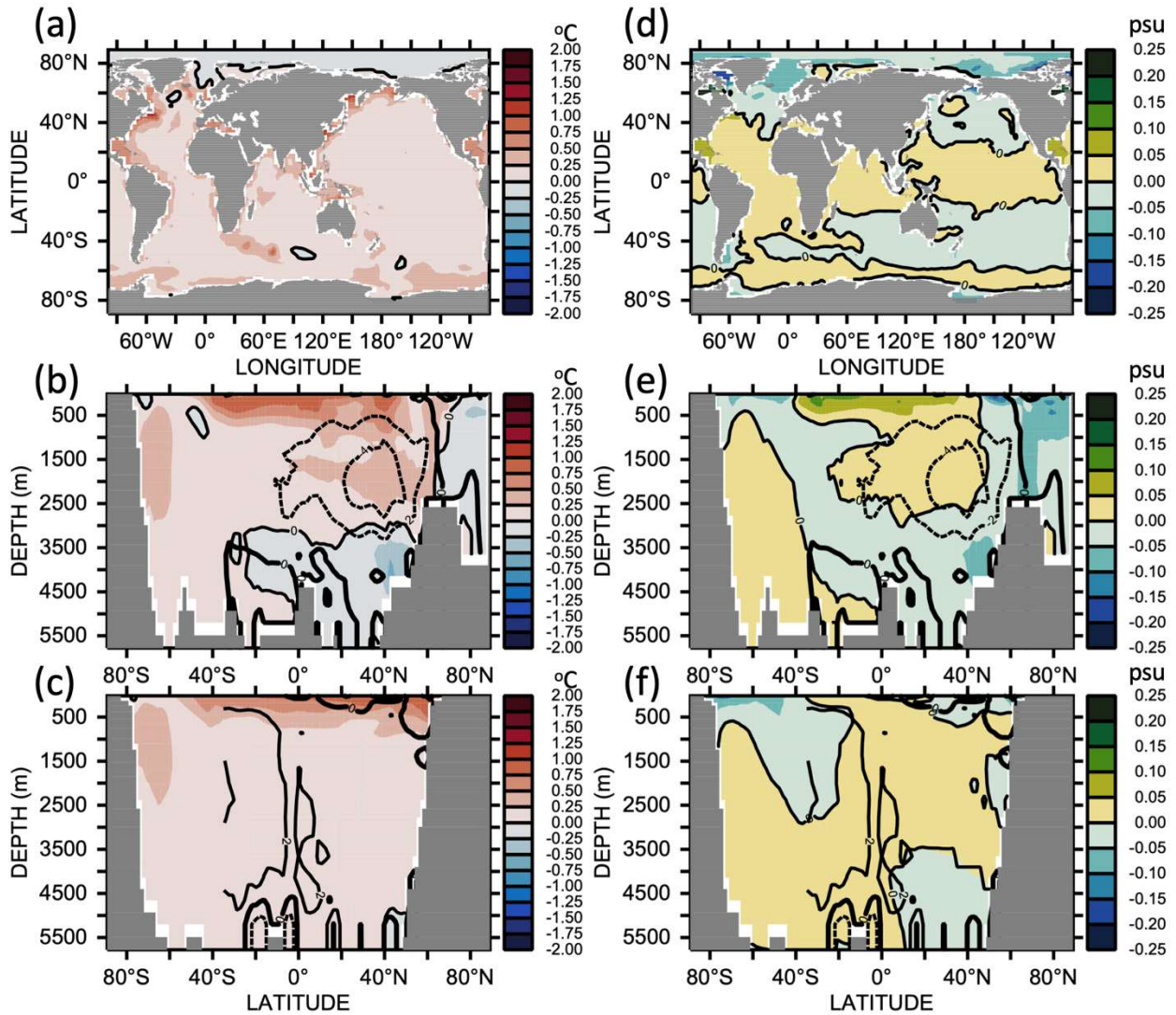


Figure S1. Left column: Simulated changes in potential temperature from year 1800 to year 2020 (a) vertically averaged, (b) zonally averaged over the Atlantic Ocean sector, (c) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Right column: Simulated changes in salinity from year 1800 to year 2020 (d) vertically averaged, (e) zonally averaged over the Atlantic Ocean sector, (f) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Contours denote the corresponding temporal changes in the global meridional overturning of the Eulerian plus bolus velocities, computed in the respective basins north of 36°S with contour interval of 2 Sv.

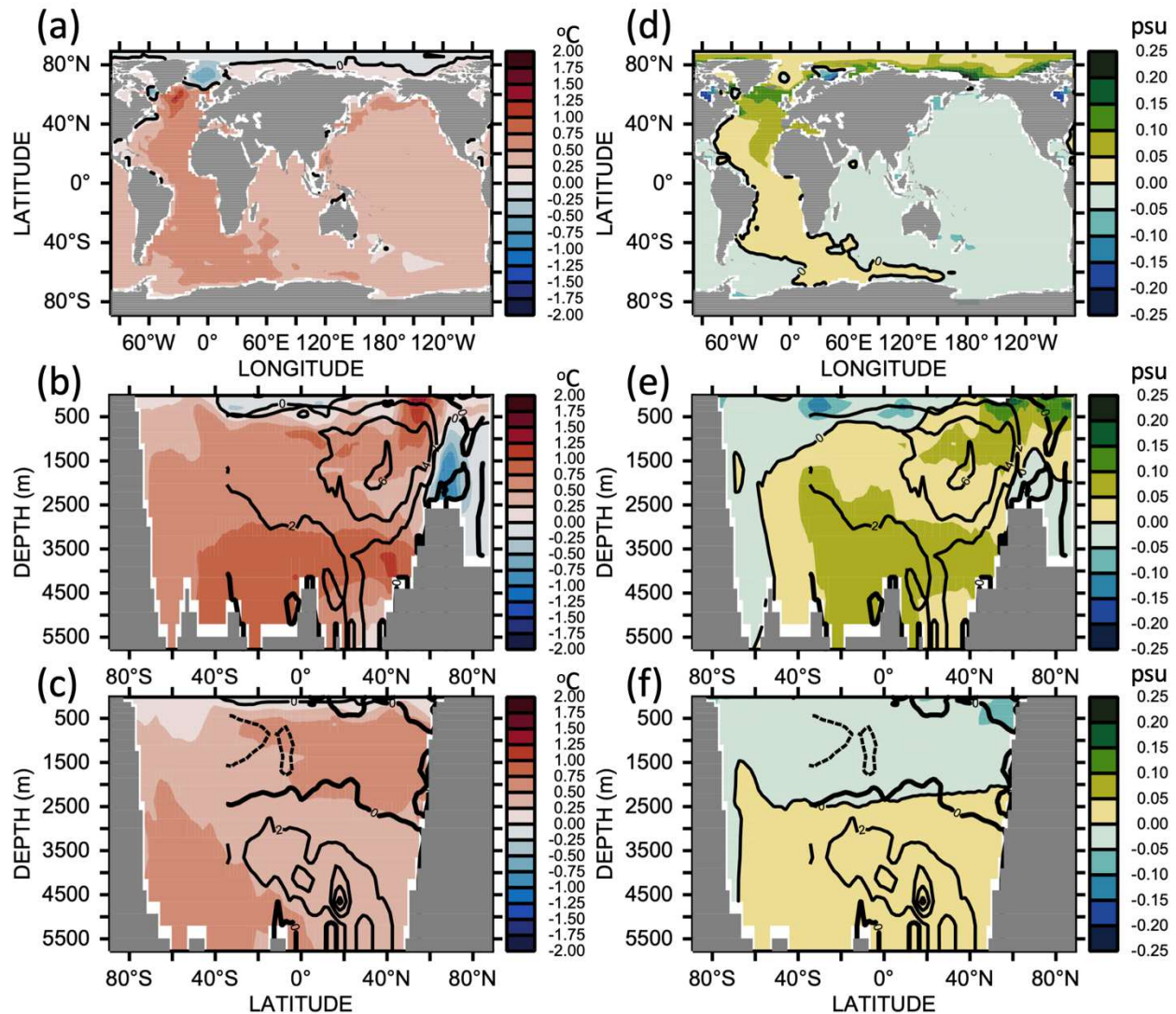


Figure S2. Left column: Simulated changes in potential temperature from year 2020 to year 2650 (a) vertically averaged, (b) zonally averaged over the Atlantic Ocean sector, (c) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Right column: Simulated changes in salinity from year 2020 to year 2650 (d) vertically averaged, (e) zonally averaged over the Atlantic Ocean sector, (f) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Contours denote the corresponding temporal changes in the global meridional overturning of the Eulerian plus bolus velocities, computed in the respective basins north of 36°S with contour interval of 2 Sv.

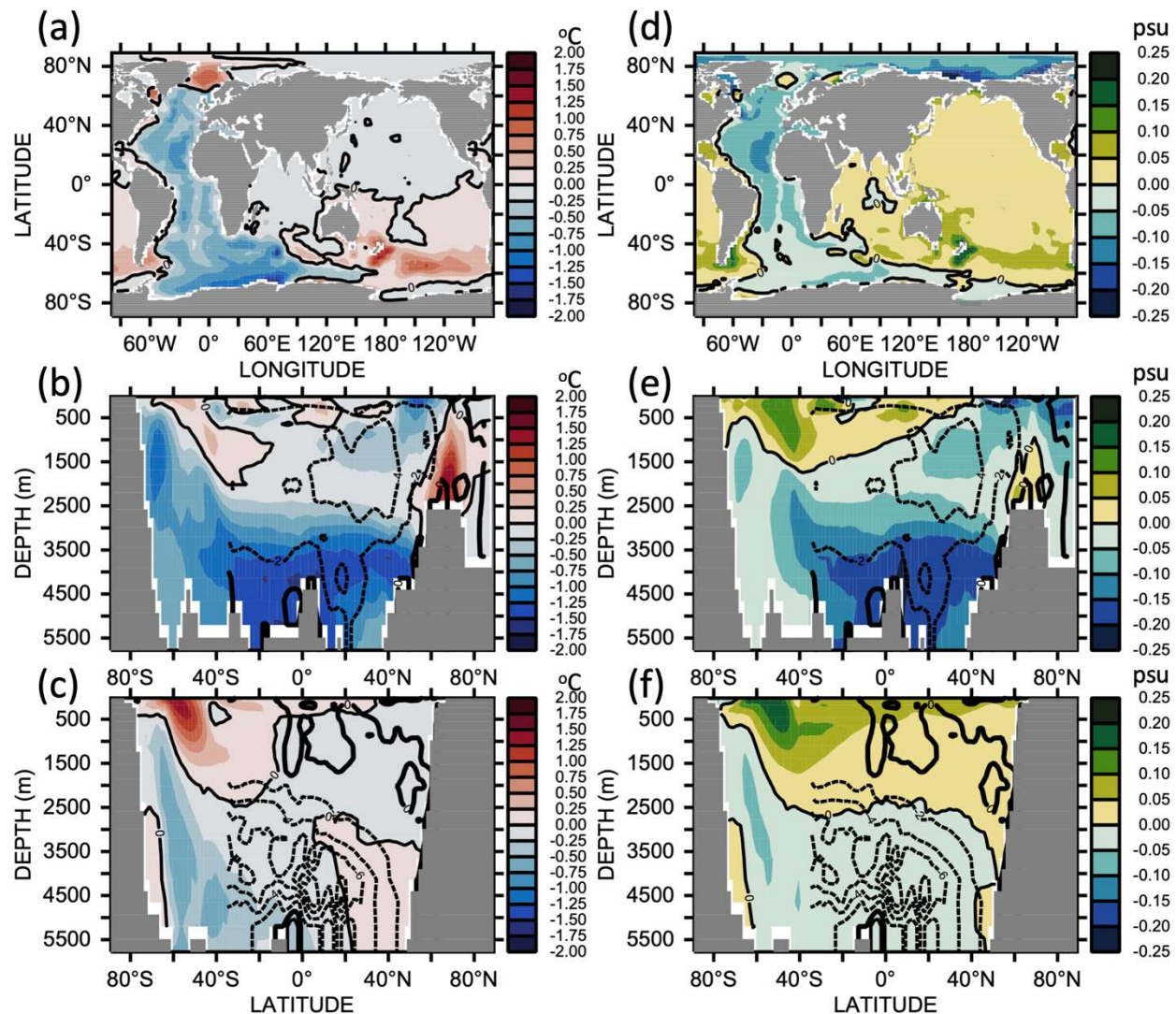


Figure S3. Left column: Simulated changes in potential temperature from year 2650 to year 3000 (a) vertically averaged, (b) zonally averaged over the Atlantic Ocean sector, (c) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Right column: Simulated changes in salinity from year 2650 to year 3000 (d) vertically averaged, (e) zonally averaged over the Atlantic Ocean sector, (f) zonally averaged over the Indian Ocean and Pacific Ocean sectors. Contours denote the corresponding temporal changes in the global meridional overturning of the Eulerian plus bolus velocities, computed in the respective basins north of 36°S with contour interval of 2 Sv.

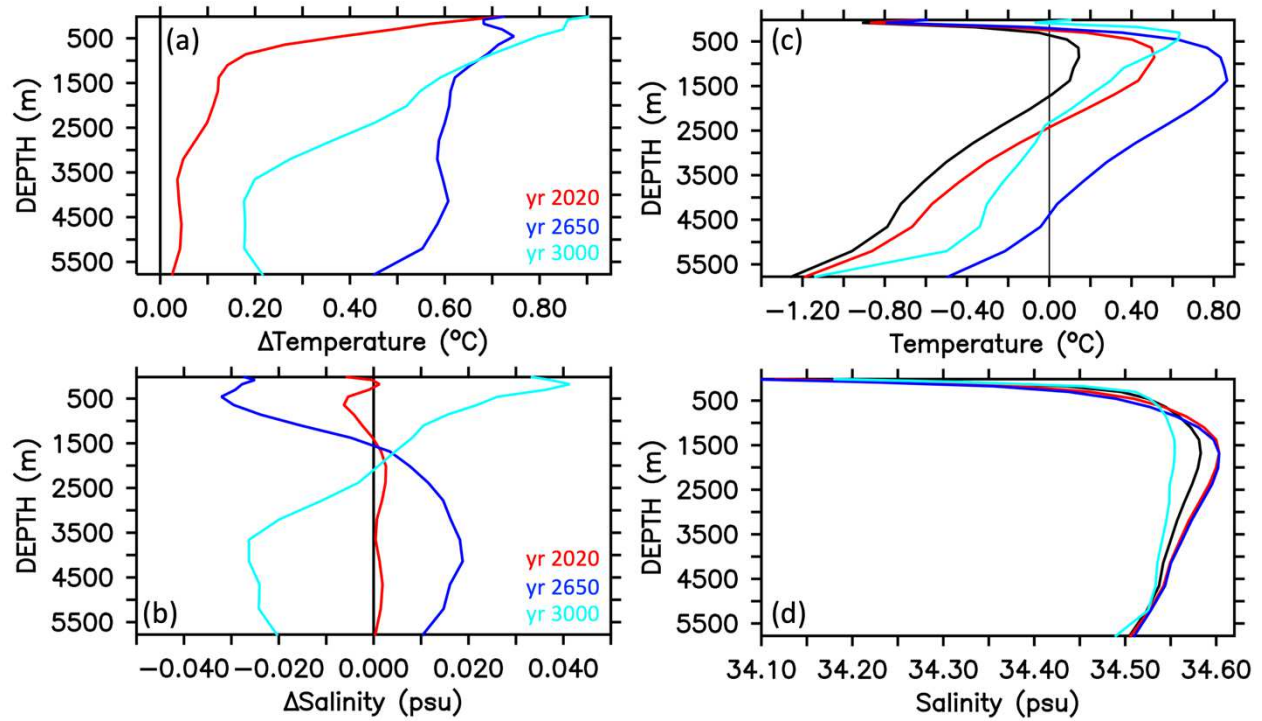


Figure S4: Profiles of simulated (a) global mean temperature relative to year 1800, (b) global mean salinity relative to year 1800, as well as (c) temperature and (d) salinity averaged laterally over the entire Southern Ocean south of 60° S. Color scale is black for year 1800, red for year 2020, blue for year 2650, and cyan for year 3000.

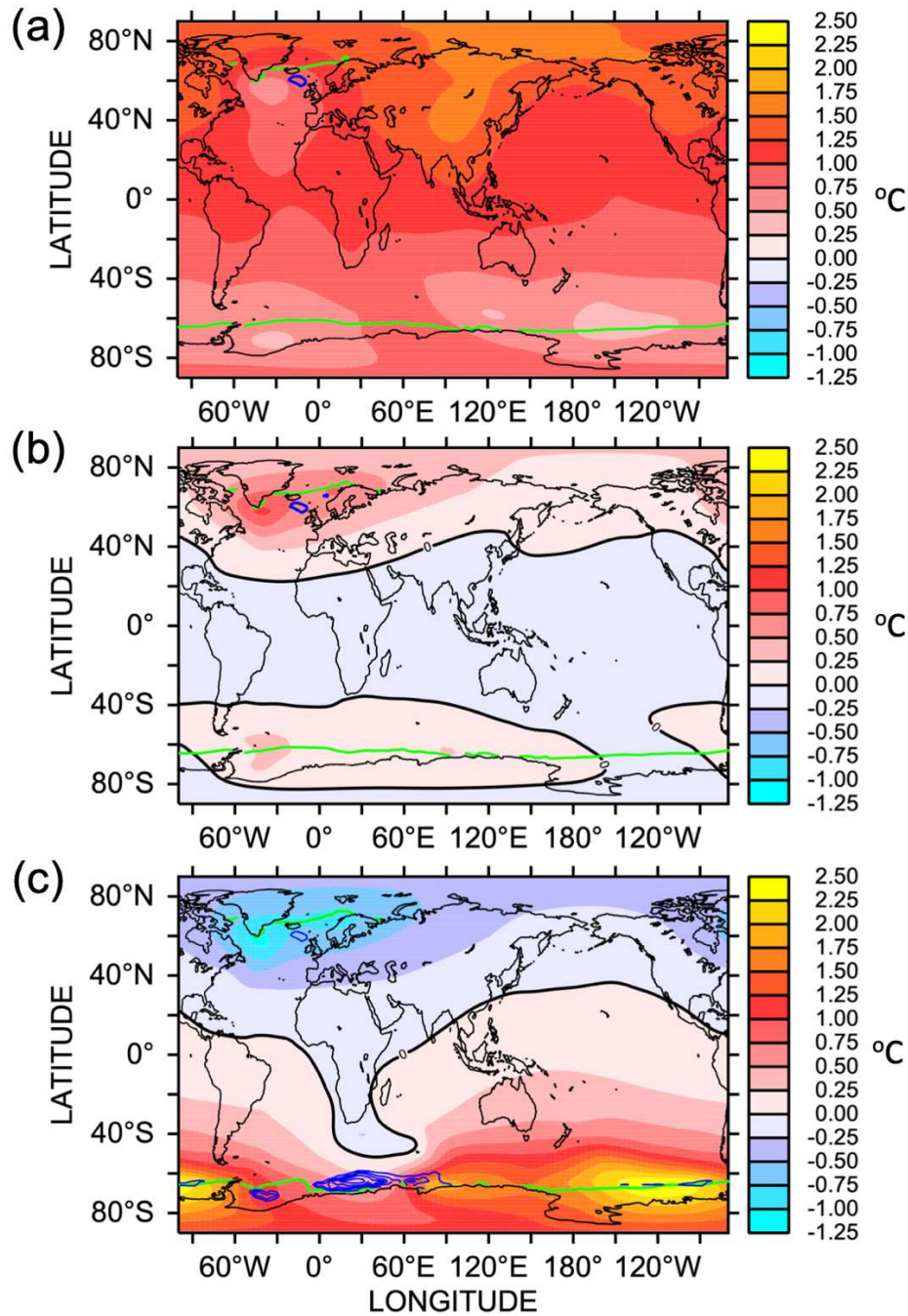


Figure S5: Simulated difference in annual mean surface air temperature over the time period (a) year 2020 minus 1800, (b) year 2650 minus 2020, and (c) year 3000 minus 2650. Units are degrees Celsius. Blue contour lines show the maximum annually averaged depth of the surface mixing layer during the respective time periods. Increments are 400m, and annually averaged mixing deeper than 400m is present in the North Atlantic between Iceland and Scotland in all panels, and in the Southern Ocean in (c) only. Green lines mark the regions where annual-mean sea ice fraction exceeds 0.5 in the respective time periods.

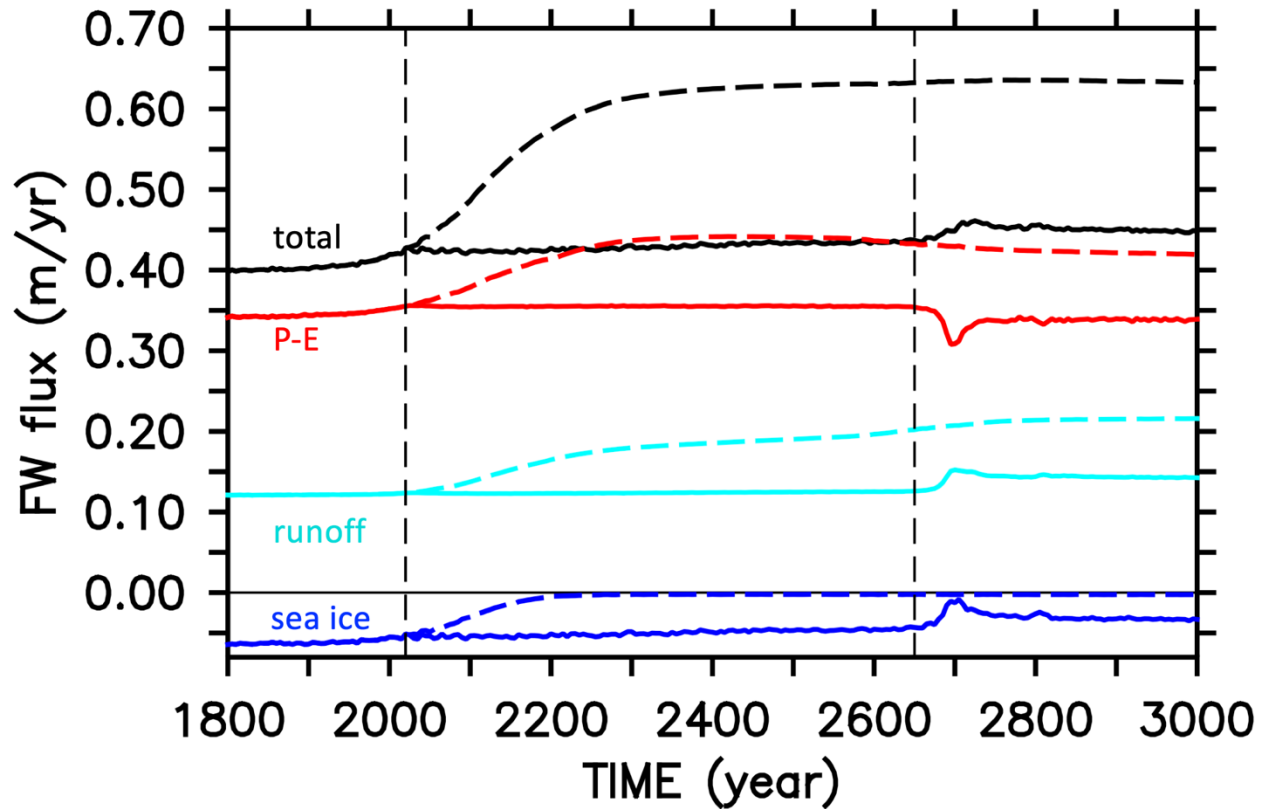


Figure S6: Time evolution of fresh water fluxes into the Southern Ocean south of 60°S for the standard experiments with emissions stopped in year 2020 (solid) and the extended RCP8.5 simulations with zero emissions only after year 2300 (dashed). The total fresh water flux into the area (black is the sum of precipitation minus evaporation (red), runoff from Antarctica (cyan) and freshwater loss by equatorward export of sea ice (blue). Precipitation minus evaporation increases by 0.014 m/yr from 0.34 m/yr in year 1800 to 0.35 m/yr in 2020, freshwater loss by sea ice export across 60°S decreases by 0.013 m/yr from 0.065 m/yr in year 1800 to 0.051 m/yr in 2020.

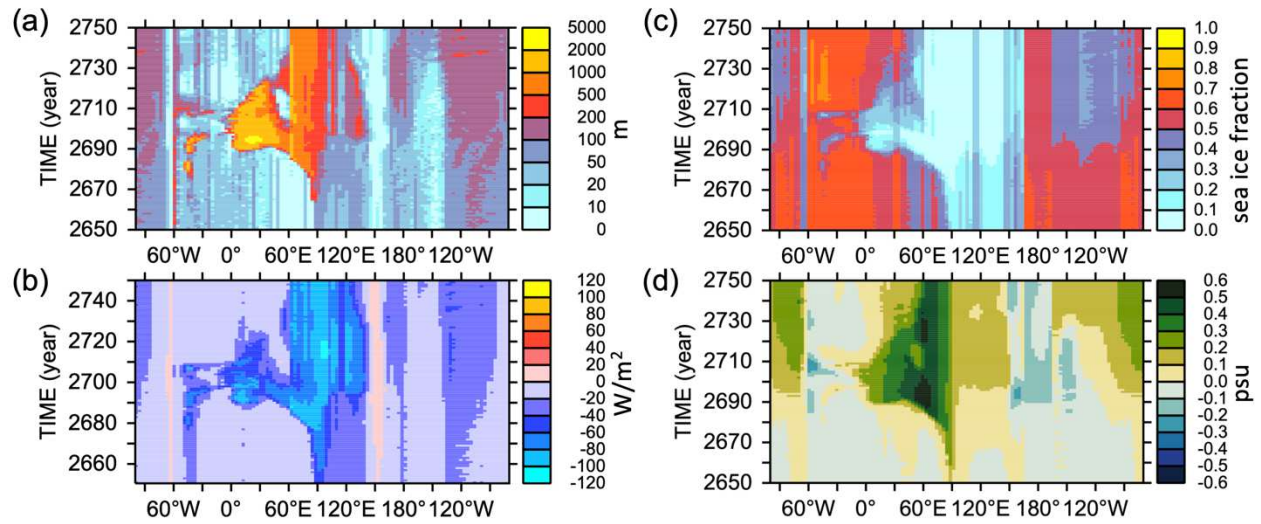


Figure S7: Hovmöller plot of (a) maximum depth of annual mean surface mixing layer, (b) annual mean air-sea heat flux, (c) annual mean sea-ice fraction, and (d) change in surface salinity with respect to year 2020, all meridionally averaged south of 60°S for the period year 2650 to 2750.

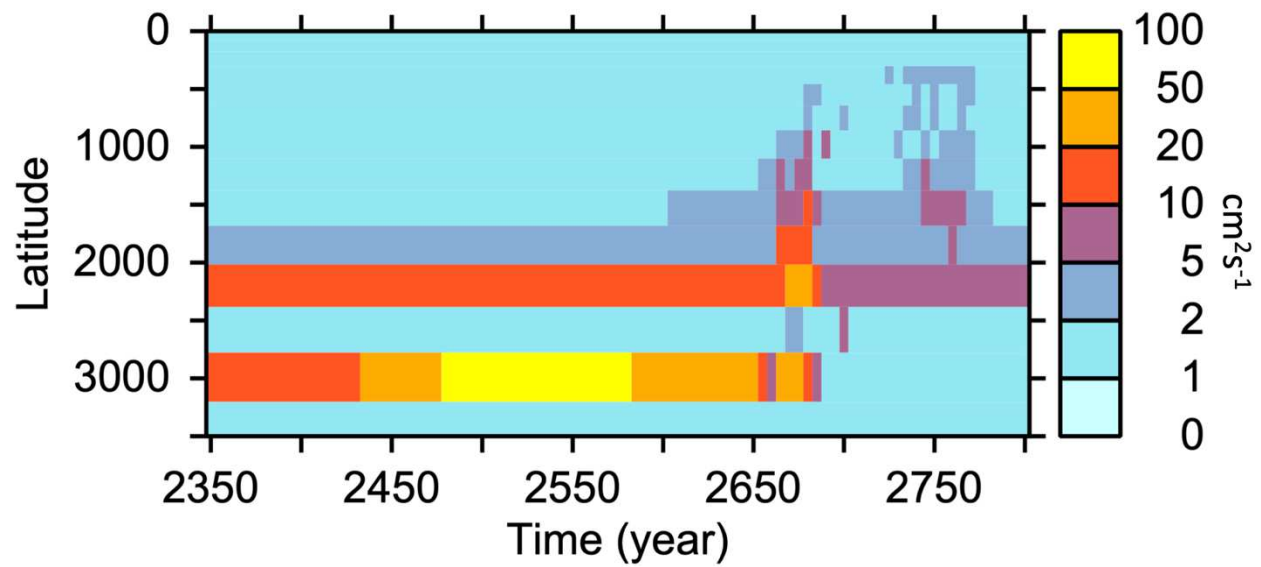


Figure S8: Vertical mixing coefficient k_v , in $\text{cm}^2 \text{s}^{-1}$, which includes a stratification dependent component increasing with decreasing static stability, laterally averaged over the region 82.8°E - 90°E , 66.5°S - 61.2°S where deep convection develops first (Figure S7).

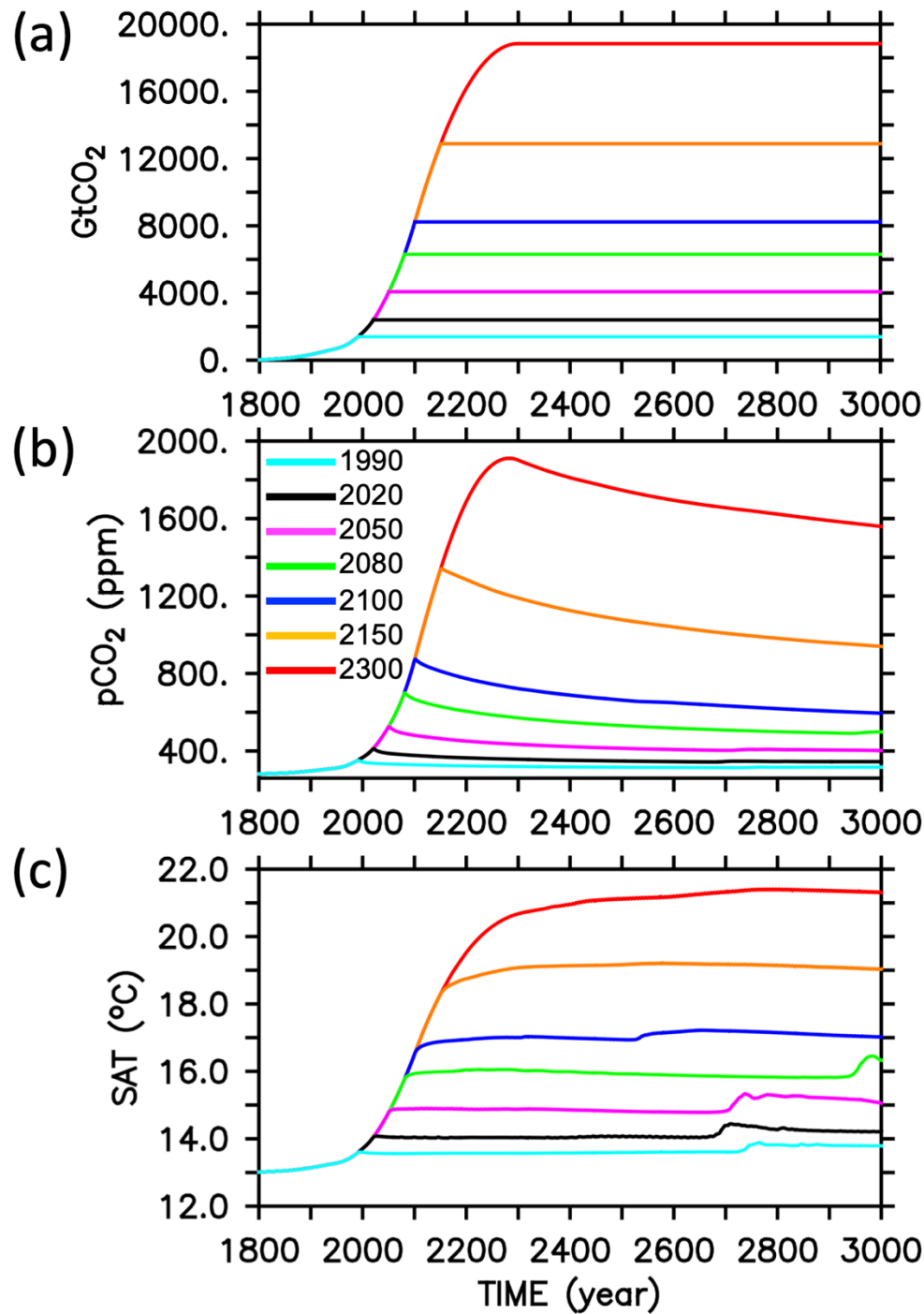


Figure S9: Time evolution of (a) cumulative anthropogenic CO₂ emissions, (b) atmospheric pCO₂ and (c) global mean surface air temperature for model simulations branching off from an extended high-emission RCP 8.5 scenario to net-zero emissions in different years as indicated in the legend of panel (b).

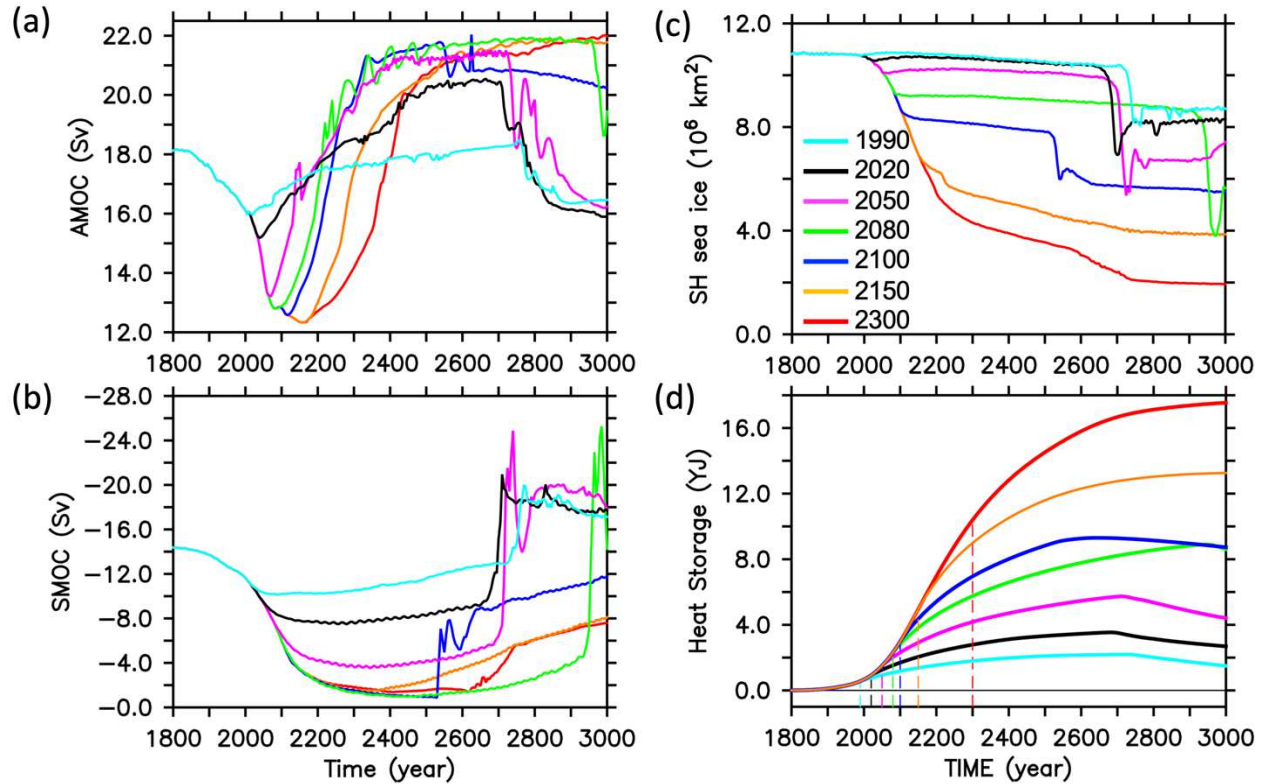


Figure S10: Time series of (a) the maximum (within 200 and 4000 m, 26°N and 71°N) of the Atlantic Meridional overturning stream function and (b) negative maximum (within 200 and 5500 m, 86°S and 44°S) of the Southern Ocean overturning stream function for model simulations branching off from an extended high-emission RCP 8.5 scenario to net-zero emissions in different years indicated in the legend of panel (c) that shows the simulated southern hemispheric annual mean sea ice area. Panel (d) shows the global ocean heat storage relative to year 1800, with the vertical dashed lines indicating the time of the switch to net-zero emissions in the respective runs. Both stream functions shown in (a) and (b) are integrated zonally along geopotential surfaces and include the bolus velocity.

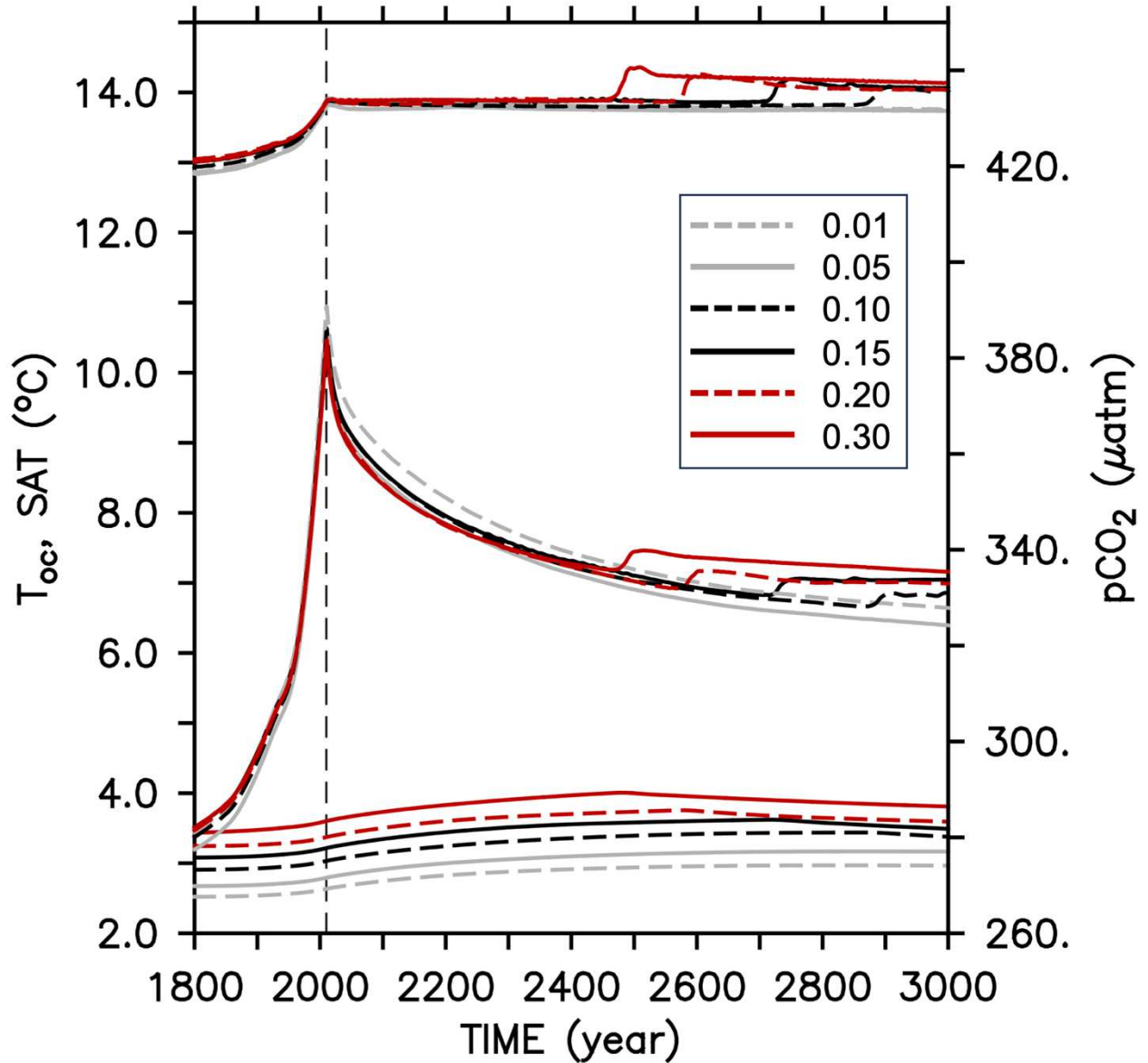


Figure S11: Time evolution of global mean surface air temperature (top, left axis), atmospheric $p\text{CO}_2$ (middle, right axis) and ocean temperature (bottom, left axis) for simulations with historical emissions and a switch to zero emissions in year 2010 for different background vertical diffusivities k_v with values given in cm^2/s in the inset. The simulation with diffusivity $k_v=0.01 \text{ cm}^2/\text{s}$ does not show any Southern Ocean deep convection event until year 5000. Simulations with $k_v=0.4 \text{ cm}^2/\text{s}$ and $k_v=0.5 \text{ cm}^2/\text{s}$ had also been performed, but due to an incomplete spin up did not stabilize SAT upon reaching net zero emissions, and were therefore excluded from the analysis.

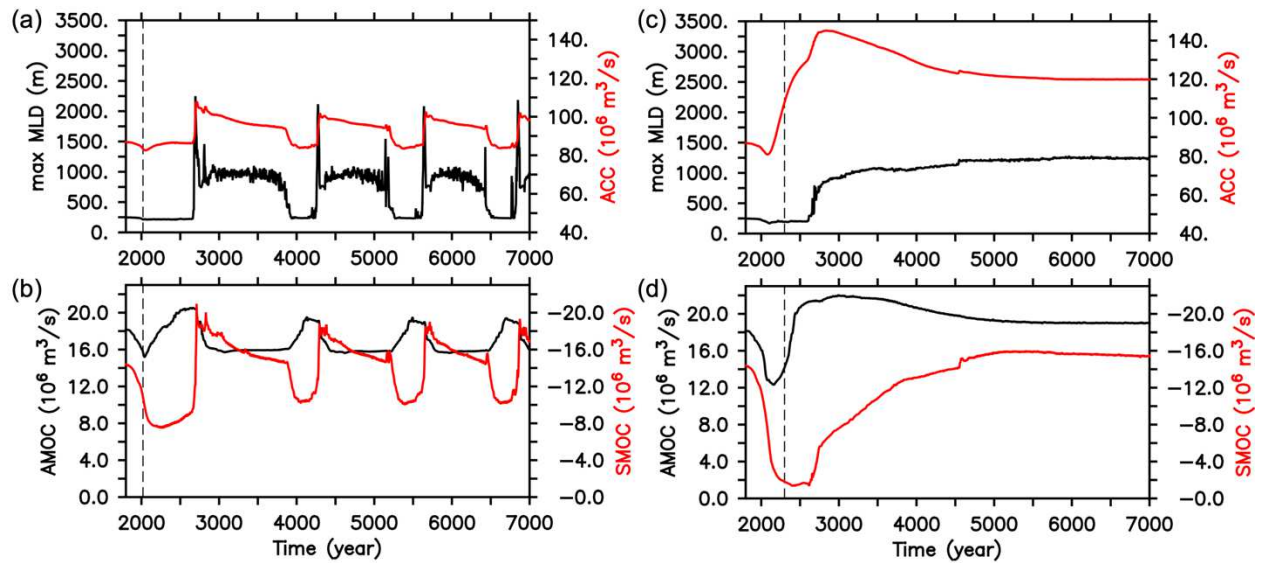


Figure S12. Left hand column shows results of the standard run with switch to net-zero emission in year 2020, right hand column of the sensitivity simulation that reaches net zero in year 2300. (a,c) Time series of annually averaged maximum mixing depth south of 44°S (black) and volume transport through Drake Passage (red). (b,d) Maximum (within 200 and 4000m, 26°N and 71°N) of the Atlantic Meridional overturning stream function (black) and negative maximum (within 200 and 5500m, 86°S and 44°S) of the Southern Ocean overturning stream function (red). Both stream functions include the bolus velocity. Time of reaching net-zero emissions indicated by vertical dashed lines.

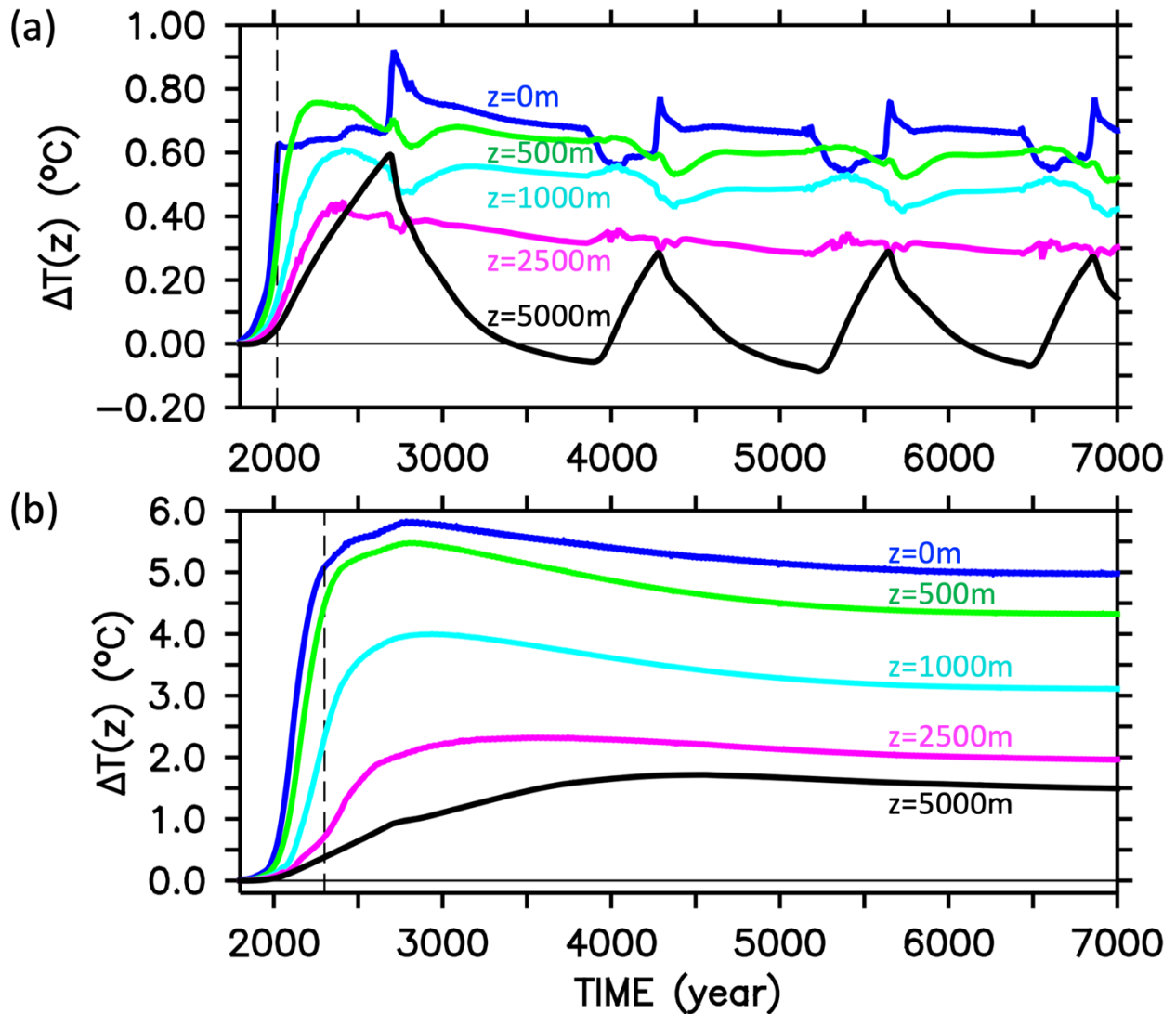


Figure S13. Time series of annually and globally averaged temperature anomalies relative to pre-industrial at the sea surface (blue), at 500m (green), 1000m (cyan), 2500m (pink) and 500m depth (black) for (a) the reference simulation that switches to net-zero emissions in year 2020, and (b) the simulation that reaches net zero in year 2300. Time of reaching net-zero emissions indicated by vertical dashed lines.

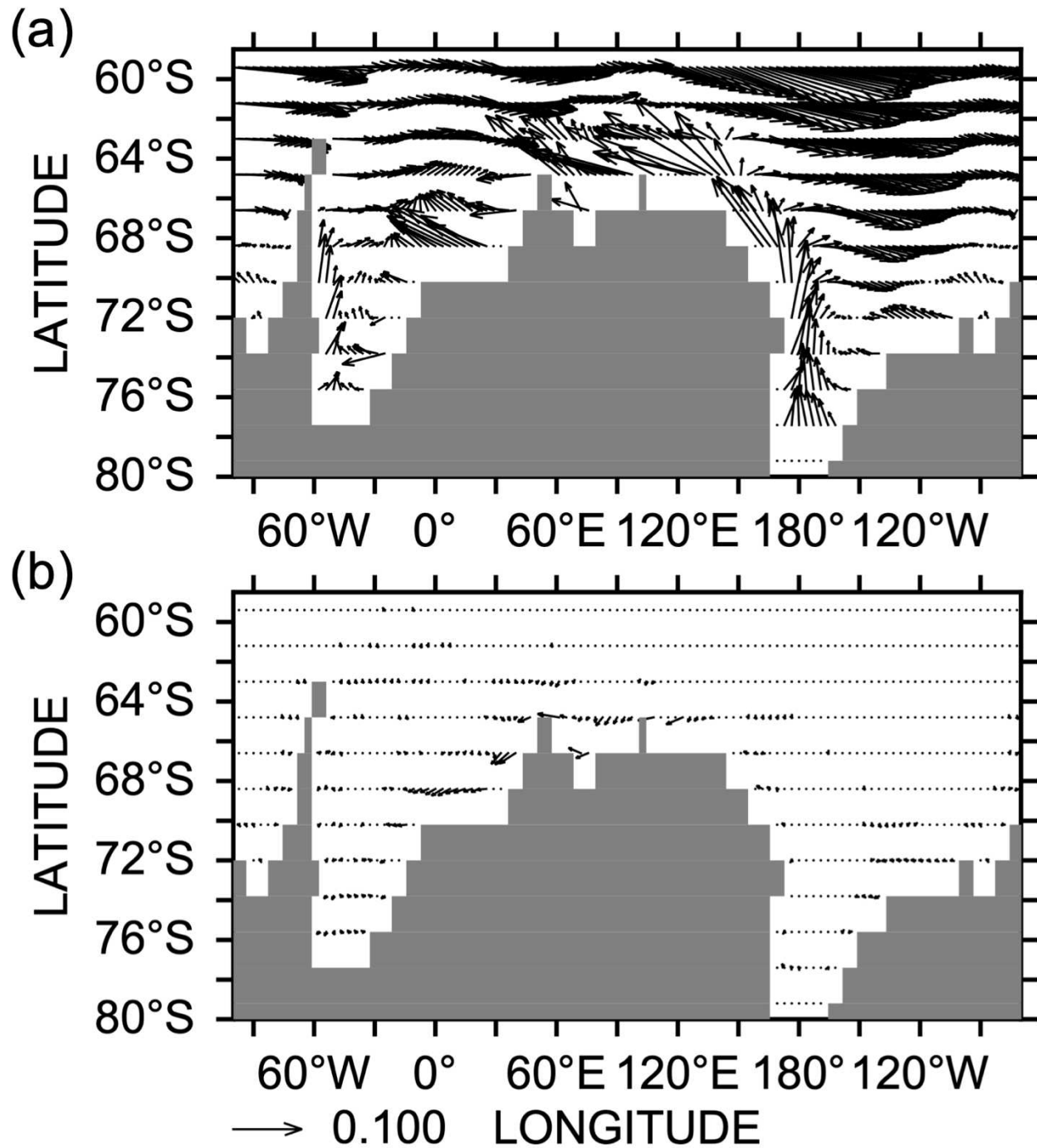


Figure S14: Southern Ocean annual mean wind stress vectors (shown is every 3rd grid point in zonal direction). Panel (a) shows wind stress for year 2650 and (b) the difference between year 2700 minus year 2650, surrounding the deep convection event. The scaling vector at the bottom denotes a wind stress of 0.1 Pa.