

Supplemental Material for  
**On freshwater fluxes and the Atlantic Meridional Overturning Circulation**

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Here we include the following additional figures:

S1: Annually averaged freshwater fluxes for 0K simulation.

S2: Difference in overturning streamfunction between simulation with +2°K-equivalent freshwater forcing and the reference case, for Southern Ocean and Indo-Pacific.

S3: As Fig. 3 in the main text, but for depth-averaged overturning.

S4: As Fig. 3 in the main text, but for simulations that include sea-surface salinity restoring.

S5: As Fig. 4 in the main text, but for depth-averaged overturning.

S6: As Fig. 4 in the main text, but for simulations that include sea-surface salinity restoring.

S7: Change in the difference between the salinity at 750m in a box in the North Atlantic (40-58°N, 15-49°W) and a box in the South Atlantic (40-58°S, 1-35°W), versus change in temperature-equivalent freshwater fluxes. Regressions are forced through (0,0).

S8: As Figure S7, but for the entire North Atlantic vs. the rest of the ocean at 750m.

S9: As Figure 2a,d in the main text, but for the surface. a) 0K simulation, b) difference between +2K and 0K simulation.

S10: As Figure S7, but for sea surface salinity of the Atlantic vs. Pacific basins.

And the following additional text:

MITgcm configuration details: Global ocean bathymetry is derived from *Sloss* [1988]. The mesoscale eddy field is parameterized using the *Redi* [1982] and *Gent and McWilliams* [1990] schemes with an eddy diffusivity of 1000 m<sup>2</sup>/s. The diapycnal diffusivity is 5x10<sup>-5</sup> m<sup>2</sup>/s. Momentum is dissipated using a Laplacian viscosity with horizontal coefficient of 5x10<sup>5</sup> m<sup>2</sup>/s and a vertical coefficient of 1x10<sup>-3</sup> m<sup>2</sup>/s. Tracers are advected using a flux-limited second-order scheme. The acceleration technique of *Bryan* [1984] is used, with a momentum/vorticity timestep of 0.25 hr and a temperature/salinity timestep of 12 hr. While the tracer acceleration distorts the physics of transients, equilibrium solutions are virtually unaffected. Enhanced mixing in boundary layers and regions of low stratification is implemented using a kappa-profile parameterization mixing scheme [*Large et al.*, 1994]. Wind stress forcing is taken from *Trenberth et al.* [1989]. Heat fluxes are imposed by restoring (with a piston velocity of 42 cm/day) to a monthly varying effective atmospheric temperature as suggested by *Haney* [1971], which incorporates both a monthly heat flux climatology and a monthly sea surface temperature field [*Jiang et al.*, 1999; *Levitus and Boyer*, 1994b]. See *Lauderdale* [2010] for a more detailed description of the model.

Overtuning streamfunction:  $\Psi$  is defined to be the zonally integrated transport of water above the potential density surface  $\varrho$ , i.e.

$$\Psi(y, \varrho) \equiv \int_{z_{bot}}^0 \overline{V \mathcal{H}(\varrho - \sigma_1(x, y, z, t))} dz$$

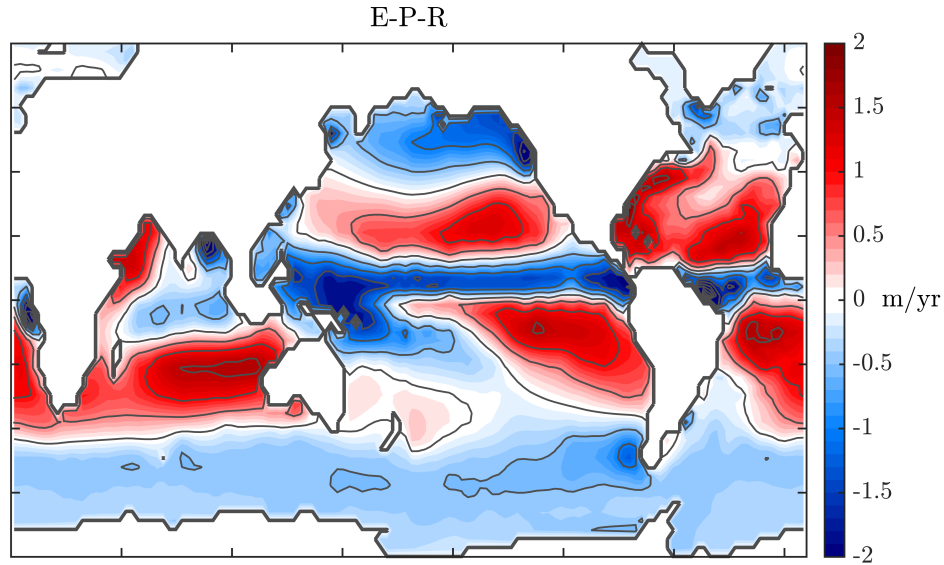
where the overline represents averaging in time and integrating longitudinally across the respective basin,  $V = v + v_{GM}$  is the sum of the resolved velocity and the eddy velocity from the Gent-McWilliams parameterization,  $z_{bot}$  is the depth of the ocean bottom,  $\mathcal{H}$  is the Heaviside step function, and  $\sigma_1$  is the potential density field referenced to 1km depth. Thus the potential density  $\varrho$  serves as the vertical coordinate, making  $\Psi$  an isopycnally-averaged streamfunction. For Figure 3 in the main text the isopycnally averaged overturning streamfunction is mapped back to physical space using the time and zonal mean potential density field.  $\Psi^z$  is defined to be the zonally integrated transport of water above a constant depth  $z$ , i.e.

$$\Psi^z(y, z) \equiv \int_z^0 \overline{V} dz$$

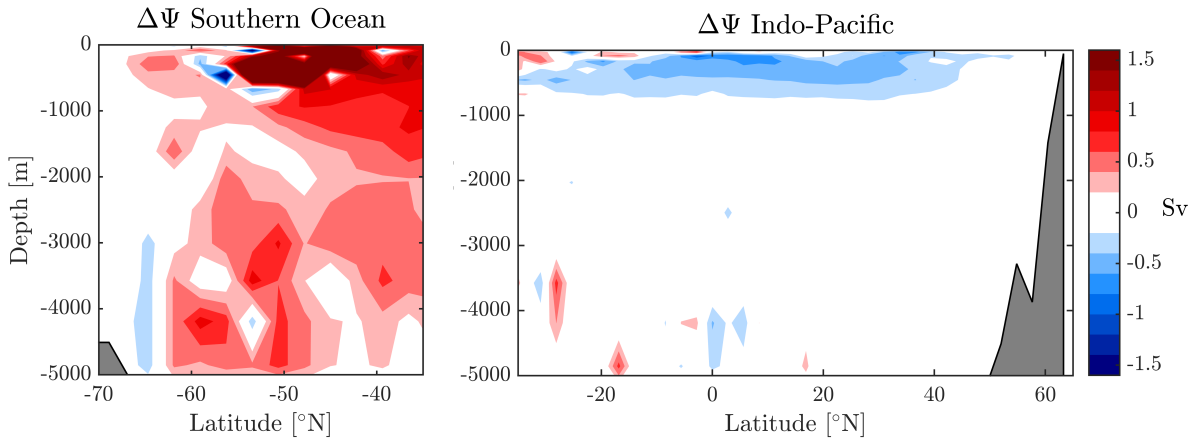
Scaling of AMOC changes: The sensitivity of the AMOC strength to small changes in the freshwater forcing amplitude across our simulations amounts to about 2 – 5% per 7% change in freshwater forcing. This result is broadly consistent with scaling arguments that suggest that, all else being equal, AMOC strength is proportional to a bulk buoyancy contrast across the AMOC [e.g. Gnanadesikan, 1999; Nikurashin and Vallis]. Assuming that changes in the relevant buoyancy contrast,  $\Delta\delta b$ , are dominated by changes in the salinity contrast,  $\Delta\delta S$ , whose relative change in turn is proportional to the relative change in freshwater forcing,  $\epsilon$ , (as predicted by Eq. 3 in the main text), we find

$$\frac{\Delta\Psi}{\Psi} \sim \frac{\Delta\delta b}{\delta b} \sim \frac{g\beta\Delta\delta S}{\delta b} \sim \frac{g\beta\delta S}{\delta b} \frac{\Delta\delta S}{\delta S} = \frac{g\beta\delta S}{\delta b} \epsilon,$$

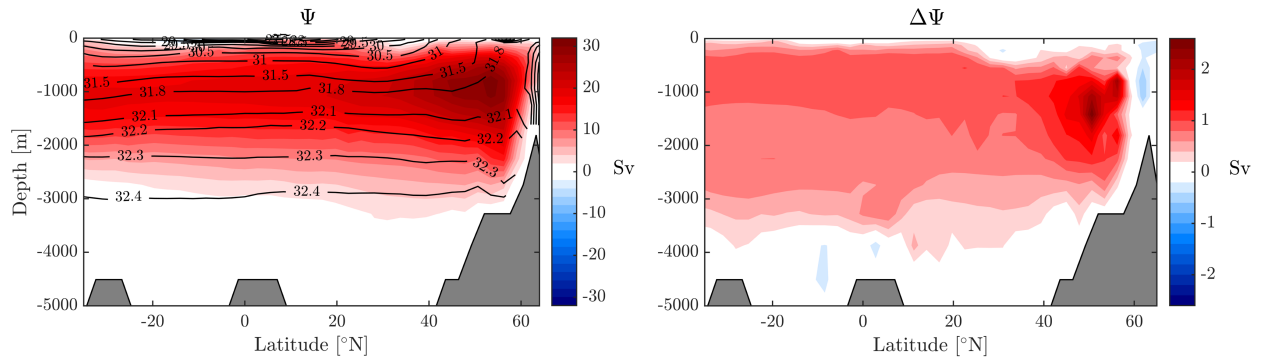
where  $g = 9.81 \text{ ms}^{-2}$  is the gravitational acceleration and  $\beta \approx 7.5 \times 10^{-4} \text{ psu}^{-1}$  is the haline contraction coefficient. The relative change in the AMOC transport hence is expected to scale with the relative change in the freshwater forcing times the relative contribution of salinity to the overall buoyancy contrast. A quantitative evaluation of the latter would require a specific definition of the relevant buoyancy contrast, but assuming a salinity contrast of around 0.5 psu (c.f. Fig. 2) and a total buoyancy contrast on the order of  $10^{-2} \text{ ms}^{-2}$  (equivalent to a temperature contrast of around  $5^\circ\text{C}$ ) gives  $\frac{g\beta\delta S}{\delta b} \sim 0.4$ . That is, the scaling would suggest the relative change in the AMOC to be around 40% of the relative change in the freshwater forcing, which is within the range of our model results.



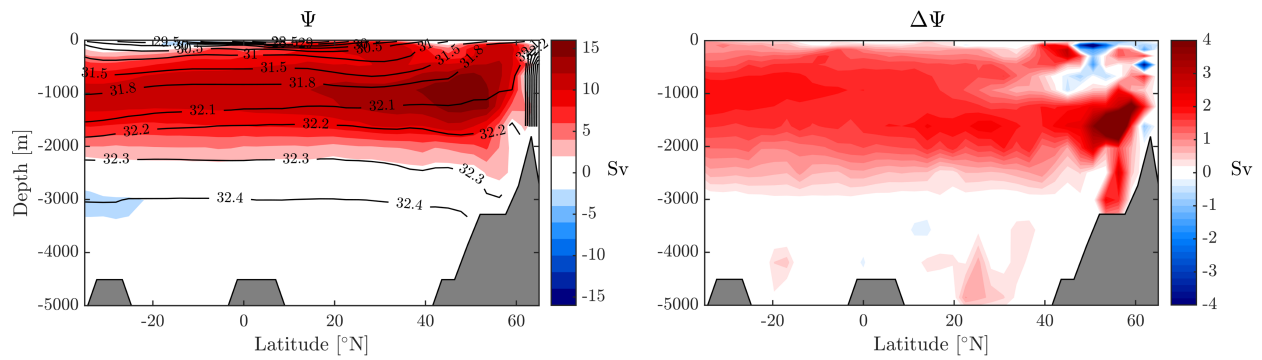
**Figure S1.** Annually averaged surface freshwater flux for 0K simulation, synthesized from observations by *Jiang et al.* [1999]. Positive values denote net evaporation, while negative values denote net precipitation and runoff. Notice that the colorbar is truncated towards large negative values, which reach up to  $-4.7\text{m/yr}$  in the Amazon outflow region.



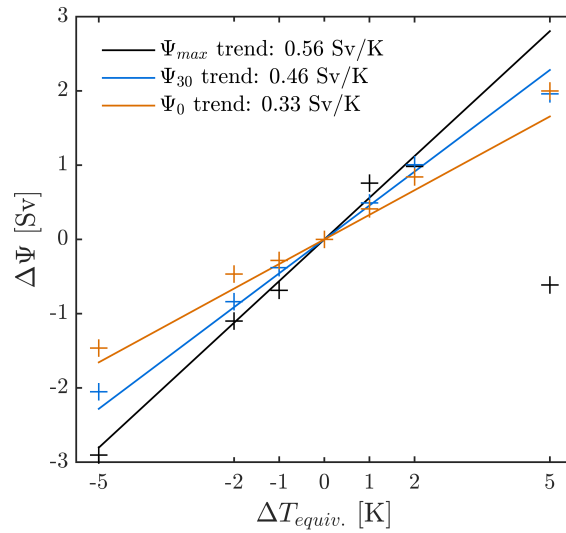
**Figure S2.** Difference in overturning streamfunction between simulation with  $+2^\circ\text{C}$  equivalent FW forcing and the reference case, for Southern Ocean (left) and Indo-Pacific (right). Notice that the colorscale is saturated in the upper Southern Ocean. Compare to the anomalous overturning in the Atlantic in Figure 3 of the main text.



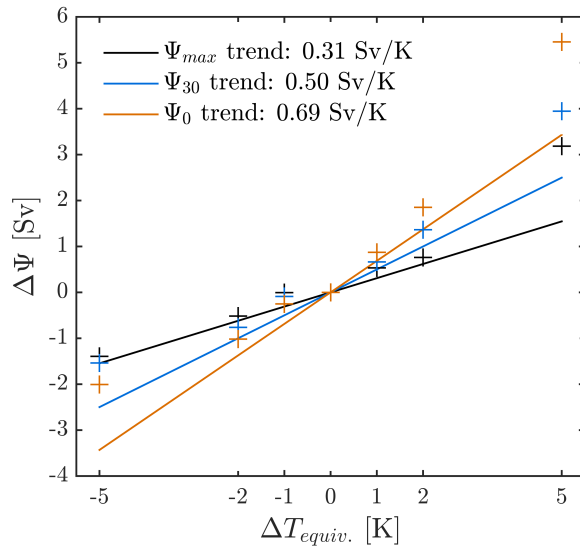
**Figure S3.** As Fig. 3 in the main text, but for depth-averaged overturning.



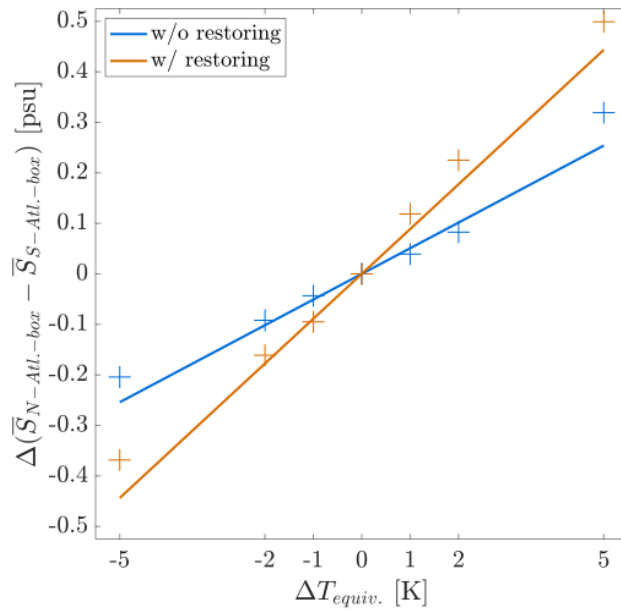
**Figure S4.** As Fig. 3 in the main text, but for simulations that include sea-surface salinity restoring.



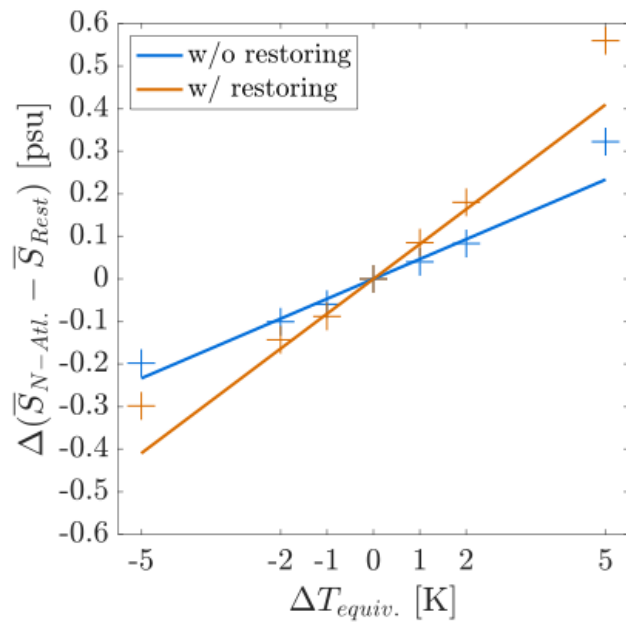
**Figure S5.** As Fig. 4 in the main text, but for the depth-averaged overturning.



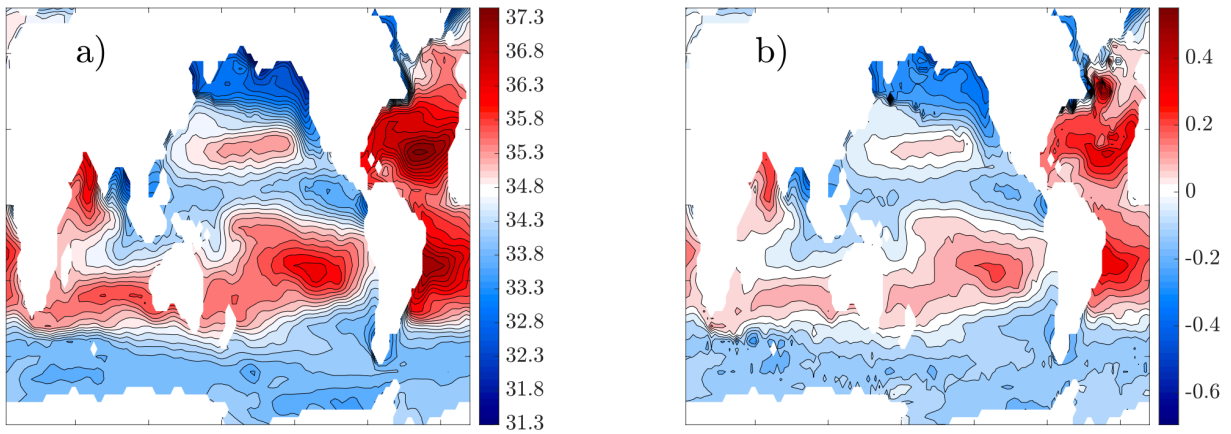
**Figure S6.** As Fig. 4 in the main text, but for simulations that include sea-surface salinity restoring.



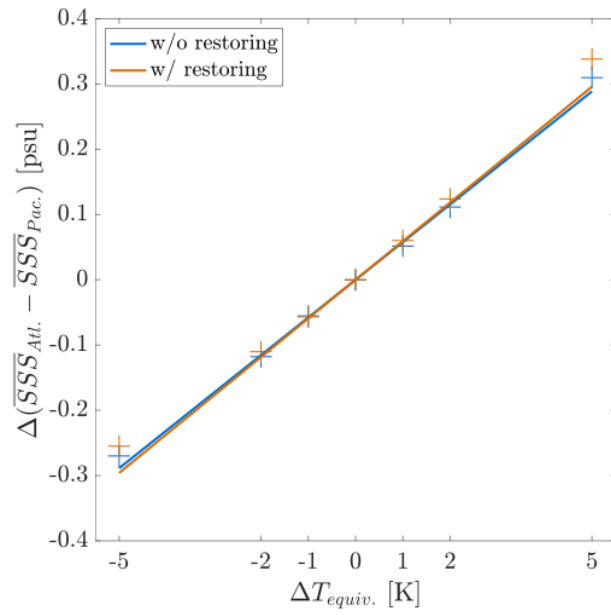
**Figure S7.** Change in the difference between the salinity at 750m in a box in the North Atlantic (40-58°N, 15-49°W) and a box in the South Atlantic (40-58°S, 1-35°W), versus change in temperature-equivalent freshwater fluxes.



**Figure S8.** As Figure S7, but for the entire North Atlantic vs. the rest of the ocean at 750m.



**Figure S9.** As Figure 2a,d in the main text, but for the surface. a) 0K simulation, b) difference between +2K and 0K simulation.



**Figure S10.** As Figure S7, but for sea surface salinity of the Atlantic vs. Pacific basins.