Supporting Information for

On timescales and reversibility of the ocean's response to enhanced Greenland Ice Sheet melting in comprehensive climate models

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Methods

Freshwater perturbation

We derive the freshwater (FW) perturbation from the meltwater and calving flux rates provided by Bamber et al. (2012) averaging over the years 1991-2010 while maintaining the seasonal cycle inherent to this monthly data set as well as its spatial heterogeneity (Fig. S1). However, we scale the local fluxes so that an overall annual mean of 0.05 Sv results. The perturbation is simplified as it has no inter-annual variability, does not depend on the modeled climate state and is prescribed as a step-function from year 0 to 99 of the perturbation experiments extending over 200 years in total.

In past hosing experiments an artificial FW flux of typically 0.1 to 1.0 Sv was applied (e.g. Hu et al., 2011; Jackson and Wood, 2018) on basin-wide scales (Kageyama et al., 2013; Jackson and Wood, 2018), focused on just the SPNA deep convection regions (Hu et al., 2011), more realistically as an additional coastal runoff influx (Swingedouw et al., 2013; Böning et al., 2016) or even accounting for iceberg melt patterns (Berk et al., 2021).

A total FW flux of 0.05 Sv is more than twice the anticipated increase in GrIS mass loss by the end of the 21st century (Golledge et al., 2019) but only half of the classic hosing flux of ±0.1 Sv, which in many simulations yields a strong reduction or even shutdown of the AMOC. Taking also past rapid climate change events into consideration, recent simulations with the MPI-ESM-ISM (including an interactive ice-sheet model) yield a FW input to the ocean of about 0.05 Sv by ice surges from instabilities of the Laurentide Ice Sheet entirely due to internal variability of the coupled system (Ziemen et al, 2019) supporting our choice. Similarly, the meltwater applied to the subpolar North Atlantic for simulations of the last deglaciation is also in this range of up to 0.06 Sv (Obase and Abe-Ouchi, 2019).

Figure S1: The enhanced Greenland runoff (EGR) forcing used in the sensitivity experiments. The freshwater flux is derived from the Greenland mass balance data provided by Bamber et al. (2012). We computed a long-term mean climatology maintaining spatial heterogeneity and seasonal cycle of the freshwater flux using the years 1991–2010 including the spatial pattern of the recent melt increase.
We also note, that a small or moderate FW perturbation of less than 0.1 Sv is rather unlikely to push the climate system to its bifurcation point. The actual tipping point, i.e. the forcing magnitude causing irreversible climate change or bifurcation of the transient climate trajectory, depends on the sensitivity of the climate (or model) system (Rahmstorf et al., 2005) but this is beyond the focus of the present study.

The FW perturbation is applied in addition to the runoff simulated by the climate model in its equilibrium state. A noteworthy difference among the ocean models is that in AWICM the FW perturbation is applied as a virtual salt flux whereas MPI-ESM and FOCI prescribe real FW. While in all models the FW is entering the ocean through the surface, we find it rapidly mixed downward in the water column on the Greenland shelf resembling the vertical distribution of the fjord export in the real world. Further, the lack of interactive ice sheet components in our models does neither prohibit this study nor limit its outcomes because rapid ice-sheet melt or disintegration occurs on a much faster timescale than ice sheet growth. We thus assume errors related to the model insufficiencies discussed here on the large-scale modifications presented to be small.

**Sea level change decomposition**

For the discussion on sea-level change (SLC) we follow the nomenclature of Gregory et al. (2019) and further theoretical considerations and explanations of Griffies and Greatbatch (2012) and Griffies et al. (2014) for Boussinesq ocean models. From such models we can diagnose stero-dynamic sea-level (SdynSL) change consisting of variations in regional ocean dynamic sea level (DSL) and a global-mean thermosteric sea level (GMTSL) anomaly. We note that the contribution of halosteric SLC to the global mean is negligible—even in freshwater-release scenarios such as studied here—but may be significant on regional scales (Gregory et al., 2019, App. 2). DSL has zero global mean though ocean model diagnostics often includes the global mean barystatic sea level in the related output field, i.e. the contribution by water inflow through the ocean’s boundaries from land and atmosphere. SdynSL can be divided into a steric and a manometric component relating the ocean volume change to changes in local density and mass (or bottom pressure). We diagnose steric sea-level change

$$SSLC = -\frac{1}{\rho_0} \int_{-H}^{\eta} (\rho - \rho_{ref}) \, dz$$

(1)

based on local density differences from the long-term mean local density $\rho_{ref}(x,y,z)$ of the reference run integrated over the full ocean depth $z=-H(x,y)$ to the surface $z=\eta(x,y,t)$ using a representative ocean density of $\rho_0=1035 \text{ kg/m}^2$ (cf. Griffies et al., 2014, Eq. 3). Density with respect to surface pressure is computed from annual mean salinity and potential temperature model output. For computing thermosteric sea level change the local long-term mean salinity of the reference run is applied. Manometric sea-level change

$$mSLC = DSLA + gmbSLC + gmtSLA - SSLC$$

(2)

is computed from $DSL$, the regional dynamic sea-level anomaly from the long-term mean of the reference run (with zero global mean by definition), the global-mean barystatic sea-level change $gmbSLC$ due to the FW input, and $gmtSLA$, the global-mean thermosteric sea-level anomaly. The sum of $DSL$ and $gmbSLC$ were available as model
diagnostic; \( \text{gmtSLA} \) is based on the global integral of Equation (1) with \( \rho = \rho(T, S_{\text{ref}}, p) \) where the local long-term mean salinity of the respective reference run was used. In our analysis of manometric SLC we omit changes on the continental shelves, because here the signal is compromised by large internal variability, and we focus on the regional response of the deep ocean (depth > 500 m) instead.

Model output processing and statistical methods

All our analysis is based on annual averages derived from monthly mean model output, because the relevant robust changes we present and found distinguishable from internal variability occur on decadal to centennial timescales. By analyzing 7 simulations from 3 models, we find that internal variability, especially in (near-)surface quantities, is often masking the emergence and exact timing of the response. We thus use long averaging periods of 50 years to show an “equilibrium” response to the perturbation on maps and apply an 11-year running mean to time series where necessary (noted in figure captions). All quantitative results presented are statistically significant at \( p=0.99 \), if not stated otherwise. We tested the mean (total varied distance) of the perturbed state averaged over 50 years against the reference run using bootstrap Monte-Carlo simulations with 200 and 1000 random samples for each grid point on maps and for spatially integrated timeseries, respectively. Similarly, the time of emergence of FW perturbation induced change is defined as the year in which total varied distance exceeds the 0.99 percentile.

References


Supporting Figures

**Figure S2:** Low-pass filtered (11-yr running mean) time series of regional sea-level change by dynamic (left), steric (middle) and manometric (right) effects for the subpolar North Atlantic (40°-65°N, bold solid lines), the subtropical North Atlantic (15°-40°N, dashed) and the equatorial plus southern Atlantic (30°S-15°N, thin solid). Note, for the manometric SLC we considered only the deep ocean basin (depth > 500 m); neglecting the continental shelf regions greatly reduces internal variability and helps to emphasize the very low-frequency evolution. Here, the ensemble mean is shown for MPI-ESM and FOCI. Graphs in the right column are merged in Figure 1b.

**Figure S3:** Manometric sea-level change (m) averaged over years 150-199.
Figure S4: Density of passive tracer tagging freshwater released from Greenland. Ensemble means averaged over years 90-109 are shown for three depth levels: surface, ~550m and ~2100m. Note, logarithmic color scale.

Figure S5: Long-term mean barotropic streamfunction (Sv) of the reference run (colored contours) and change due to FW perturbation (yellow contours, 2 Sv increments starting at -1 Sv [dashed] and +1 Sv [solid]) averaged over years 75-125 of the experiments. Ensemble mean differences are shown for MPI-ESM and FOCI. Note, ±10 Sv contours of the mean state are depicted as solid and dashed red lines in Figure 2a, so is the -1 Sv contour of the change in yellow (solid in Fig.2a, first dashed line here).
Figure S6: Long-term mean of mixed-layer depth (MLD) in March from reference simulations (a) and its change due to the freshwater perturbation in meters (b) and in percent of reference state (c). The change is computed as difference between overall mean of reference run and averaged over model years 50-99 of experiments. For MPI-ESM and FOCI ensemble means are shown.

Figure S7: Time series of 2m air temperature change over cooling patches of the subpolar North Atlantic (see non-hatched areas in Fig. 2b. Ensemble spread in MPI-ESM and FOCI experiments is indicated by shading the ±1σ area around the ensemble-mean curves (bold lines).
Figure S8: Changes in potential temperature (a) and salinity (b) in the deep ocean at 2000m (upper panel) and 3100 m (lower panel) in response to the FW perturbation. Changes are computed subtracting the long-term mean reference state from the temperature averaged over years 75-125 of the perturbation experiments. Note, the additional FW alone in the perturbation experiments cannot explain the salinity response, feedbacks in circulation play a role, too. For this, see the similarity of passive tracer distributions in MPI-ESM and FOCI at 2000 m in Figure S4 (bottom panel) but differences in salinity here.