Modeling the effect of freshwater pulses on the early Holocene climate: The influence of high-frequency climate variability

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The effect of freshwater pulses on the early Holocene climate is investigated with the ECBilt-CLIO global coupled atmosphere-sea ice-ocean model (*Opsteegh et al.*, 1998, *Goosse and Fichefet*, 1999). In the model an early Holocene (i.e. 8.5 ka BP) equilibrium climate state is perturbed by releasing a fixed amount of freshwater (4.67×10^{14} m^3) into the Labrador Sea at three different constant rates: 1.5 Sv (1 Sv = 1×10^6 m^3 s^{-1}) in 10 years, 0.75 Sv in 20 years, and 0.3 Sv in 50 years. For each rate, five ensemble experiments have been performed, varying in initial conditions. As expected, the freshwater pulses produce a weakening of the thermohaline circulation (THC). The perturbed state is in agreement with proxy evidence for the 8.2 ka cooling event (Figure 1), which is characterized by 1 to 2°C cooling in the North Atlantic region (*Alley et al.*, 1998) and widespread drying in Northern Africa (*Gasse*, 2000).

Two types of recovery of the THC occurred, differing in time-scale: (1) ≤200 years and (2) >200 years. In the experiments with 10-year and 20-year pulses, both types of recovery were observed (Figure 2, *Renssen et al.*, 2001, 2002). This suggests that the model response is unpredictable in the range of parameters studied here. It is hypothesized that the unpredictability is associated with annual-to-decadal climate variability.

Our findings have important implications for other model studies and for the interpretation of proxy records. In contrast to earlier suggestions (e.g., *Manabe and Stouffer*, 1995), the duration of a THC perturbation event may not be a simple function of the amount of meltwater released. In our simulations, releasing two identical (relatively moderate) pulses to the same steady state climate produced events of 1200- and 200-yr duration (i.e. 10-yr 'a' and 'd' cases). This may imply, for instance, that longer time-scale events such as the Younger Dryas (~1200-yr duration) is forced by the same kind of pulse as shorter time-scale events such as the Older Dryas or the 8.2 kyr BP event (several 100 years duration). Thus, looking at the paleorecord, it may not always be possible to explain the severity of a climatic event in terms of the magnitude of the forcing. In addition, our results indicate that modelers should be cautious when interpreting results of perturbation experiments, as a particular model outcome is not necessarily unique and other solutions may exist with the same kind of perturbation. Ideally, one should also perform ensemble experiments when studying dynamics of the coupled climate system on longer time-scales to ensure that the result is robust.

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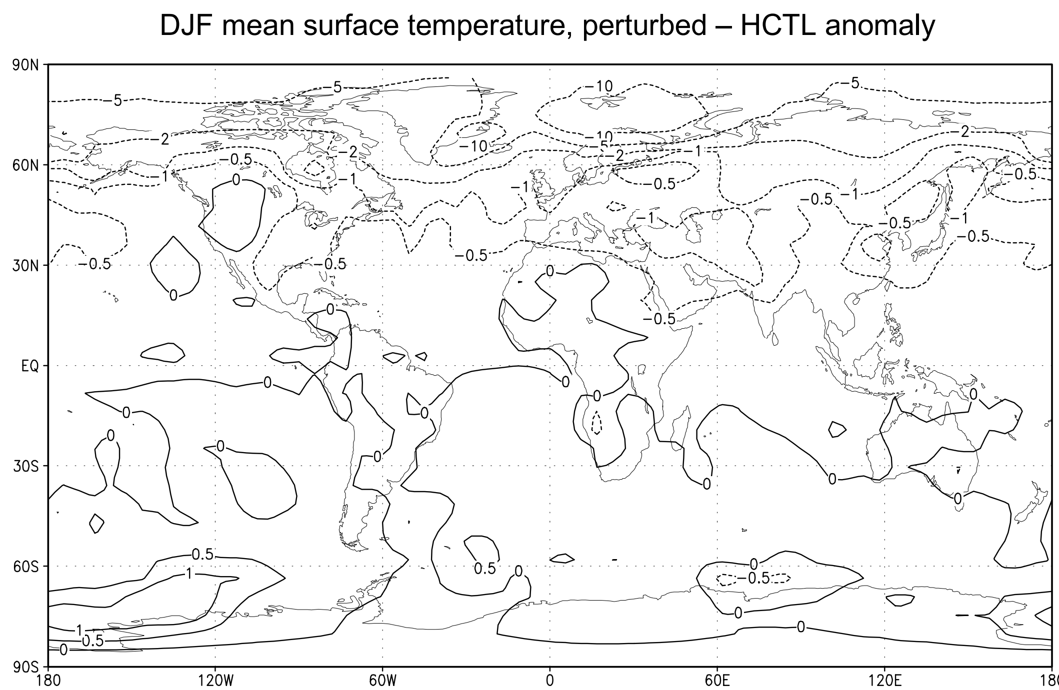

Figure 1: December-January-February mean surface temperature anomaly (20-yr [a] case, see Fig. 2), perturbed state minus early Holocene equilibrium state. Contours at –10, -5, -2, -1, -0.5, 0, 0.5, 1, 2, 5 and 10°C. Note that the strongest cooling is simulated over the Nordic Seas between Svalbard and Norway. This is the main location of deep convection in the early Holocene equilibrium climate state. In the perturbed state, deep convection occurs more to the South (near Southern Norway) and most of the Nordic Seas becomes perennially covered by sea ice (from Renssen et al. 2002).

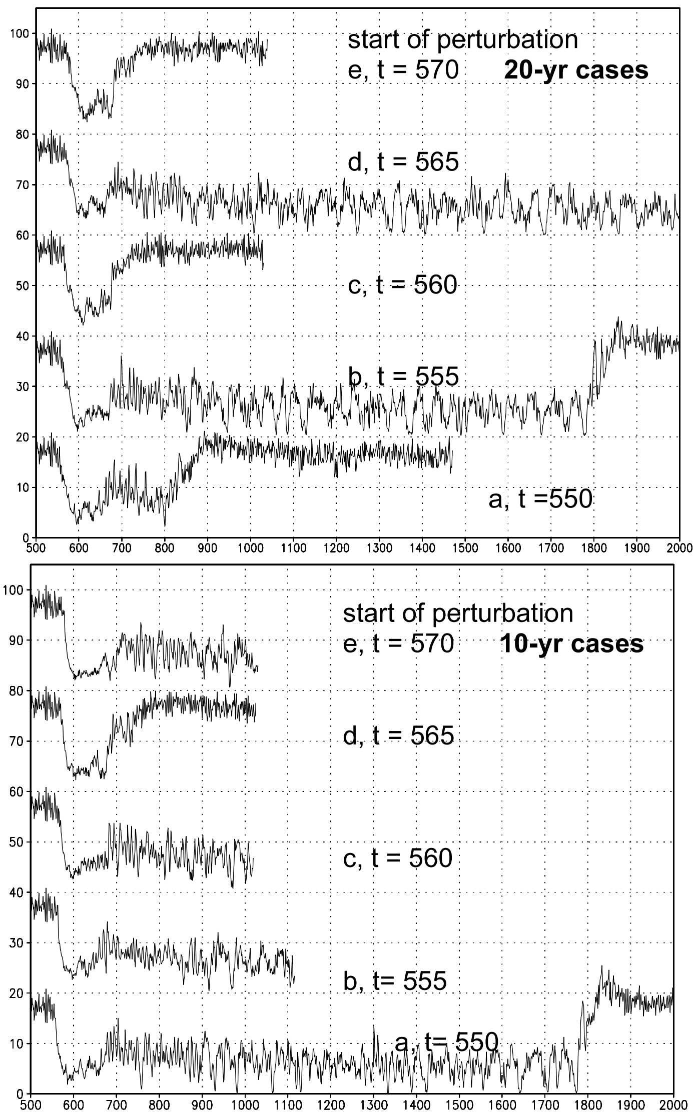

Figure 2: Maximum meridional overturning rate (Sv) in the Nordic Seas (i.e. between 60 and 80°N) plotted against time (years) for all ensemble experiments. Top: 20-yr cases, bottom: 10-yr cases. The start times of the ensembles (labeled 'a' to 'e') are indicated. The 8.5 ka BP equilibrium climate state is shown between t=500 and t=550. Note that, for convenience, values of the upper four curves have been elevated by 20, 40, 60 and 80 Sv (from Renssen et al., 2002)