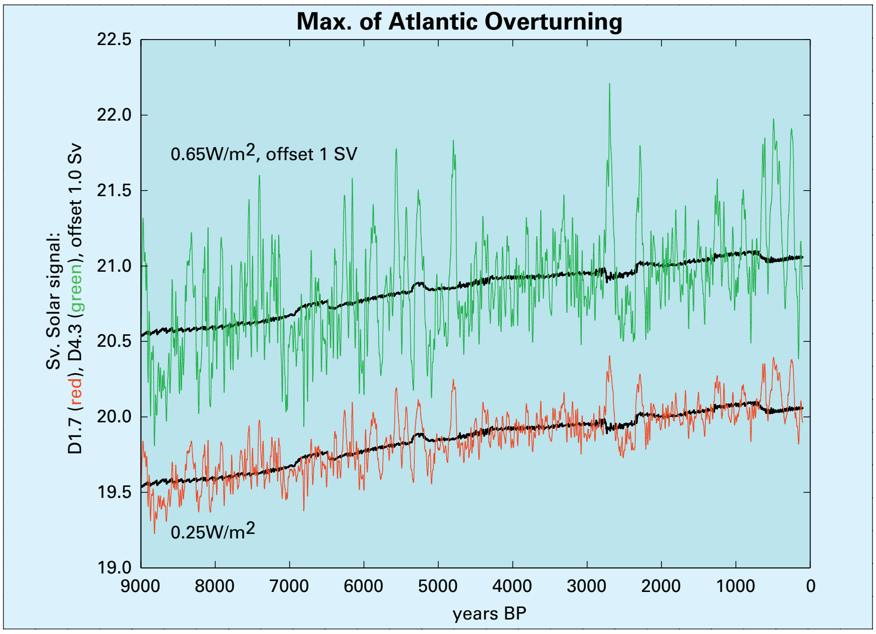

Figure 3: Maximum of North Atlantic overturning for the past 9,000 years, modeled by CLIMBER-2. The control runs (black lines) are compared to runs including solar irradiance changes, based on a parameterization by ^{14}C production rate changes, (Figure 1, red curve). Two scalings are shown, equivalent to a change in irradiance of 0.24% and 0.65% between the Maunder minimum and the present.

between the Maunder minimum and the present (Lean et al. 1995, Reid 1997). The atmospheric temperature in the North Atlantic area of CLIMBER-2 encompassing Greenland shows a linear, positive response to irradiance changes of 0.2 and 0.6°C, respectively, whereas the stream function is inversely related to solar forcing (overturning is enhanced during cooling events). Since CLIMBER-2 reveals no free decadal and centennial variability, the response of the climate system to changes in solar forcing is clearly seen in the model results. There appear to be some larger temperature excursions in response to solar variability, for example in the 3rd and 5th millennium BP, which could be as large as the climate changes in the Maunder and Spörer minima of the last millennium.

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Simulation of the Oxygen Isotope Ratio in Foraminiferal Carbonate During Heinrich Event 1: A Climate Model-Data Comparison

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The oxygen-18/oxygen-16 isotope ratio preserved in fossil carbonate shells of planktic and benthic foraminifera ($\delta^{18}O_c$) is one of the most important proxies in paleoclimatology. First, oxygen isotopes in general circulate through all components of the climate system (atmosphere, ocean and ice) and are fractionated whenever a phase transition occurs. Second, $\delta^{18}O_c$ in particular has been measured on a vast number of ocean sediment cores with high precision. Since a temperature-dependent fractionation of about 0.2‰/°C occurs during the formation of carbonate tests in foraminiferal shells, changes in $\delta^{18}O_c$ contain the clue for both changes in the temperature and the oxygen-18/oxygen-16 ratio of the ambient seawater ($\delta^{18}O_w$).

This opens up a unique opportunity to combine climate models and a wealth of paleoclimatic data: If oxygen isotopes are transported in an ocean model (e.g., Paul et al., 1999; Schmidt, 1999; Delayge et al., 2000; Paul and Schäfer-Neth, 2004), the isotopic composition of carbonate can be directly simulated. As a by-product, the relative importance of changes in seawater $\delta^{18}O$ and temperature can be assessed. By improving the match to the reconstructed climate history, a better understanding of the climate system can be gained. This, as well as the prospect of reducing uncertainties in forecasting future climate changes, lies at the heart of the DEKLIM program.

Like the two related DEKLIM-Paleo projects 'CliTrans' and 'Palaeo Isotopes', our 'Oxygen Isotopes' project focuses on the climate transition from the Last Glacial Maximum (LGM) to the Holocene (the last deglaciation). Here our special interest is on the oxygen isotope data from the Atlantic Ocean for the LGM and Heinrich Event 1 (H1), which reflect the extent and speed of climate change during this period. For the first time we employed a climate model of reduced complexity that followed the oxygen isotopes through all stages of the hydrological cycle.

As compared to the LGM, the H1 data shows large negative anomalies. At the sea surface, these anomalies are thought to reflect the discharge of icebergs from the Northern Hemisphere ice sheets, accompanied by a rapid cooling in the north and a warming in the south (the 'bipolar seesaw', Broecker, 1998). Large negative anomalies have also been reconstructed in the thermocline and down to a depth of 1,500 m in the South Atlantic Ocean, as well as in the vicinity of the Greenland-Iceland-Scotland Ridge (GISR). We argue that the common cause of all these changes is the slow-down and subsequent recovery of the meridional overturning circulation (MOC) in the Atlantic Ocean in response to meltwater input to high northern latitudes.

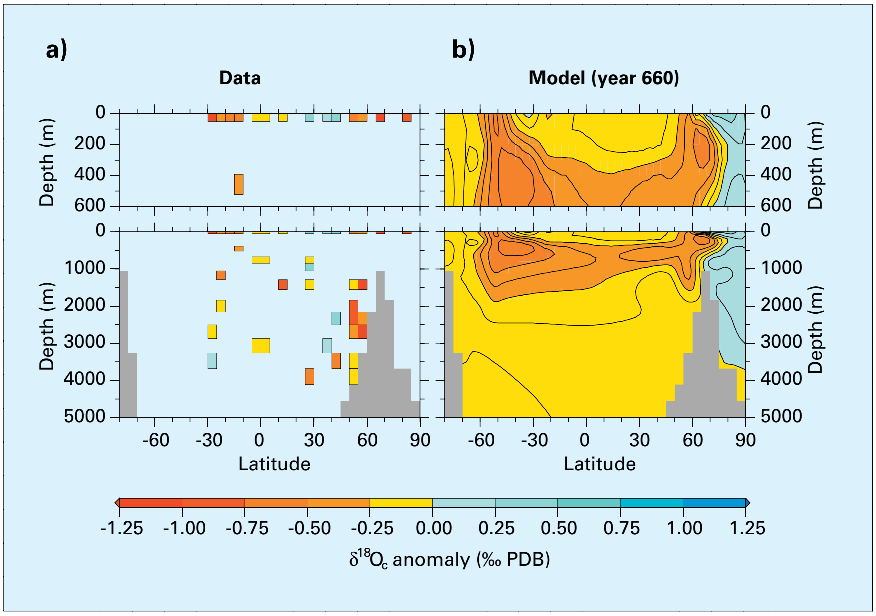

Fig. 1: The zonal-mean $\delta^{18}O_c$ anomaly in the Atlantic Ocean (Heinrich event 1 minus LGM, ‰ PDB) as found in (a) our collection of planktic and benthic high-resolution oxygen isotope records, and (b) the ocean component of the 'Hanse' climate model. The data were binned into the latitude-depth grid cells of the model. To account for the global $\delta^{18}O_w$ anomaly, 0.12‰ (after Fairbanks, 1989) was added to the data and 0.03‰ to the model output. The data is from Zahn (1986), Bard et al. (1989), Jansen and Veum (1990), Winn et al. (1991), Bickert (1992), Sarnthein et al. (1994), Jung (1996), Abrantes et al. (1998), Marchitto et al. (1998), Kiefer (1998), Kirst (1998), Richter (1998), Arz et al. (1999), Knies and Stein (1999), Rühlemann et al. (1999), Vidal et al. (1999), Völker (1999), Hülls (2000), Shackleton et al (2000), Henderiks et al. (2002), Kim et al. (2002), and Mollenhauer (2002).

Methods

The data set presented here was derived from 30 benthic and 27 planktic high-resolution oxygen isotope records from the Atlantic Ocean between 28°S and 81°N (see references on PAGES website). The stratigraphy of the cores was based on AMS-radiocarbon dating of foraminiferal carbonate. To avoid species-specific offsets, we calculated the $\delta^{18}O_c$ anomalies (differences) between H1 (15.4-16.8 ka BP) and the LGM (19-23 ka BP).

The 'Hanse' model (Paul and Schulz, 2002) is a zonally averaged, coupled climate model of reduced complexity, which consists of an atmospheric energy balance model with a hydrologic cycle after Jentsch (1991a, b), a Wright and Stocker (1992)-type ocean model and a simple sea-ice model. The model domain is the Atlantic Ocean. The meridional resolution is 5°. There is one layer in the atmospheric and 20 layers in the ocean component. We included an atmosphere-ocean isotopic cycle with fractionation of oxygen isotopes upon evaporation and precipitation.

The initial state of our Heinrich experiment was a cold climate state reminiscent of the LGM (with orbital parameters characteristic of 21 ka BP, an atmospheric CO_2 concentration of 220 ppmv and a maximum of the MOC in the North Atlantic Ocean of 9 Sv). For 100 years, freshwater was input to mid-latitudes (40° to 50°N) at a rate of 0.1 Sv (1 Sv = 10^6 m^3 s^{-1}). It was given a $\delta^{18}O_w$ signature of -40‰ to mimic meltwater discharge from glaciers in the Hudson Bay area. After this perturbation, the 'Hanse' model was integrated until a quasi-steady state was reached. From the model output, we computed the equilibrium carbonate $\delta^{18}O$ using the paleotemperature equation of Mulitza et al. (2004) (see also Mulitza et al., 2003).

Results

In the tropical and subtropical South Atlantic Ocean, the reconstructed $\delta^{18}O_c$ change (H1 minus LGM) shows a strong decrease from the surface down to a depth of about 1,500 m (Fig. 1). In the North Atlantic Ocean, the situation is more complex: Between about 30° and 50°N, the surface water is characterized by a positive $\delta^{18}O_c$ anomaly. North of about 45°N, a tongue of negative anomalies extends from the surface to the deep water along the southern flank of the GISR.

Following the rapid slow-down at the beginning, the MOC in the 'Hanse' model slowly recovered from the meltwater input until it experienced a large overshoot in the year 660 of the experiment (not shown). Shortly after, a negative $\delta^{18}O_c$ anomaly appeared near the GISR. Furthermore, in accordance with the data, a large negative $\delta^{18}O_c$ anomaly had developed in the upper 1,500 m of the South Atlantic Ocean (Fig. 1b). A further experiment showed that the MOC did not recover for a meltwater input at a slightly higher rate of 0.15 Sv (not shown).

The temperature of the surface water in the northern mid-latitudes (at 37.5°N, Fig. 2) showed a rapid cooling and a first gradual, then step-like recovery, with an overshoot similar to the MOC. The central water in the southern tropics (at 12.5° S) experienced a pronounced warming that resulted in a net decrease in $\delta^{18}O_c$, in spite of an increase in $\delta^{18}O_w$.

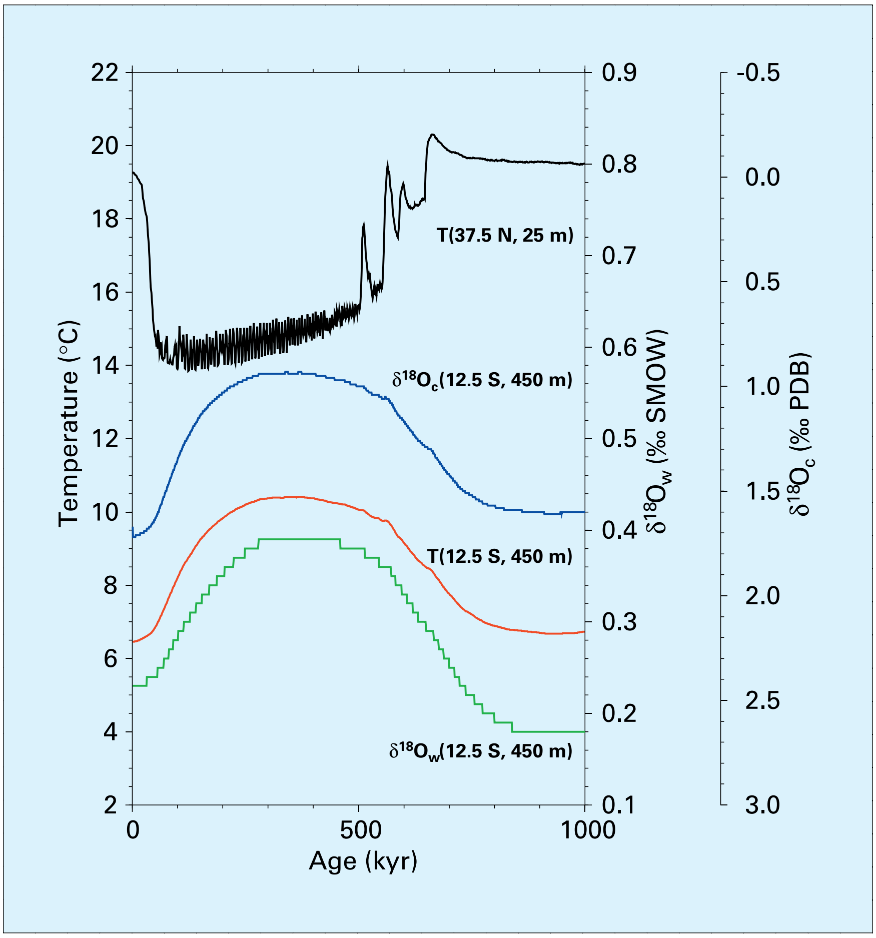

Fig. 2: Time evolution of temperature and the oxygen isotope ratios $\delta^{18}O_w$ and $\delta^{18}O_c$ at two locations in the ocean component of the 'Hanse' climate model: 37.5°N (at the surface) and 12.5°S (at 450 m depth). The first location is close to the core site of EN120 GGC1, the second to that of ODP 1078C (cf. Fig. 7 of Rühlemann et al., 2004).

Discussion

The bipolar seesaw (Broecker, 1998)—a rapid cooling in the north accompanied by a warming in the south—is a characteristic of a weakened MOC (Lohmann, 2003). In the tropics and subtropics of the 'Hanse' model, downward diffusion of heat from the surface won over horizontal advection and caused the warm anomaly to reach great depths (cf. Rühlemann et al., 2004). Correspondingly, a large negative $\delta^{18}O_c$ anomaly developed, which was only weakly influenced by changes in $\delta^{18}O_w$.

In contrast, the $\delta^{18}O_w$ signature of the glacial meltwater was the main reason for the negative anomaly near the GISR. In our simulation, the deep water temperature in this region changed by 0.5°C at most. However, the isotopically light meltwater that had been stored at the surface during the stagnation of the MOC was released upon its recovery and advected downward to about 2,000 m depth.

This 'store-and-advect' mechanism was first described by Lehman et al. (1993) in a modeling study of the Younger Dryas cold event. It possibly acts in combination with other mechanisms, such as the sea-ice brine mechanism invoked by Vidal et al. (1998) and van Kreveld et al. (2000) for low benthic $\delta^{18}O_c$ signals associated with Heinrich events in the northern North Atlantic Ocean (see also Bauch and Bauch, 2001).

We note that the sea-level change implied by our Heinrich experiment is in accordance with the simulation of a single surge of the Hudson Strait ice stream (Marshall and Clarke, 1997), but smaller than estimated for the entire H1 (approx. 10-15 m, cf. Hemming, 2004). To explain a larger sea-level change and the long duration of the low benthic $\delta^{18}O_c$ signals recorded in the sediments, either a sequence of small surges (Roche et al., 2003) or a stronger baseline MOC in the 'Hanse' model that permits a larger single surge might be required.

Conclusion and Outlook

Our simulation of the Heinrich event 1 with the 'Hanse' model of reduced complexity is a first step towards the direct simulation of paleo proxies. The model results suggest that the large negative $\delta^{18}O_c$ anomalies exhibited by the ocean sediment core data can be explained in terms of the slow-down and subsequent recovery of the meridional overturning circulation. The match to the reconstructed climate history can probably be improved by changes in the model forcing and internal parameters. Once the 'Hanse' model has been shown to be consistent with the paleoclimatic data from the last deglaciation, we intend to confront our results with projections of future climate changes (cf. IPCC 2001, Chapter 9).

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