Ice cores

High-resolution archive of rapid climate changes

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Snow falling on the central parts of polar ice sheets is compressed to firn by the weight of the overlaying snow, eventually turning to ice through sintering and recrystallization (Herron and Langway 1980). The isotopic composition of the snow deposited on the surface (a proxy for the local condensation temperature) and impurities (such as aerosols and dust particles) are preserved in this process. In addition, air is trapped in bubbles during the transition from firn to ice, and can be extracted to study, for example, past greenhouse gas concentrations.

Accordingly, the top 50-120 m of an ice sheet consists of porous snow and firn, which allows the air to circulate between the surface and the top of the firn column, while diffusive processes dominate further down the firn column (Fig. 1). Once the transformation from firn to ice is completed, the air is trapped in the ice, and the age distribution and composition of the air in the bubbles are no longer changed.

In the firn column, thermal and gravitational diffusion leads to isotopic fractionation. The isotopic composition of the gas trapped in the ice can thus be used as a temperature proxy during fast temperature changes at the surface of an ice sheet, e.g. by analyzing the $^{15}$N/$^{14}$N ratio of nitrogen gas (Huber et al. 2006; Kindler et al. 2014; Severinghaus et al. 1998). Through the mixing of air in the firn column, the air entrapped in the ice is considerably younger than the surrounding ice matrix (Fig. 1). This age difference is usually denoted as $\Delta$age, and vice versa. Therefore, maximum $\Delta$age values are observed in Antarctica e.g. in the Vostok ice core during the Last Glacial Maximum about 20,000 years ago ($\Delta$age of approx. 5000 years), while in Greenland $\Delta$age values are considerably lower (up to 1400 years during the Last Glacial Maximum).

Dansgaard-Oeschger events
During the last glacial period, North Atlantic climate was not stable. The cold stadial periods were interrupted by warmer interstadial periods of durations from 100 to several thousand years. The interstadials, also called Dansgaard-Oeschger (D-O) events, generally show a common shape in time - at the beginning, temperature increases rapidly, and subsequently decreases first slowly, then abruptly to reach stadial values again (Fig. 2a). Other climate parameters mimic this pattern. This Northern Hemisphere temperature pattern is linked to Antarctic temperature by means of the bipolar seesaw (Stocker and Johnsen 2003). This concept proposes that a reduced Atlantic Meridional Overturning Circulation (AMOC) leads to heat accumulation in the southern hemisphere (Southern Ocean) until temperature increases rapidly in the north, whereafter temperature decreases again in the south (EPICA Community Members 2006).

Due to their outstanding temporal resolution and well-constrained chronologies throughout the entire last glacial period, Greenland ice-core records are perfectly suited to investigate fast climate variations in the North Atlantic region (e.g. Huber et al. 2006; Steffensen et al. 2008), while CH$_4$ synchronized Antarctic ice cores can be used to reconstruct mechanisms which link both hemispheres during past abrupt climate changes through the bipolar seesaw (EPICA Community Members 2006).
Duration and rates of change during D-O onsets

In Figure 2b, we stack δ¹⁸O and Ca²⁺ at the onsets of D-O events 2-20. It is evident that the changes in δ¹⁸O and Ca²⁺ are equally abrupt between stadials and interstadials. The transition from stadial to interstadial conditions of δ¹⁸O takes place within 1-2 steps of the 20-years-resolution record. Within the data resolution, no significant lead or lag of Ca²⁺ relative to δ¹⁸O can be observed. The mean duration of the climate transition for Ca²⁺ is also in the order of 40 years. Within these four decades, Ca²⁺ concentrations decrease by one order of magnitude, and δ¹⁸O increases by 3.8% on average.

The Ca²⁺ record is primarily reflecting changes in dust source conditions, most likely from Central Asian desert regions (Biscaye et al. 1997; Svensson et al. 2000), and transport effects. Thus, changes in Ca²⁺ concentration indicate reorganizations of wind fields and atmospheric circulation patterns at regional to hemispherical scale. The close relative timing of δ¹⁸O and Ca²⁺ changes indicates that the rapid changes in Greenland atmospheric dust loading and in δ¹⁸O may be linked to the same large-scale circulation changes.

Gas concentrations stored in ice cores change more slowly than δ¹⁸O and Ca²⁺ because they are well mixed in the atmosphere and have residence times of a decade (CH₄) or more (e.g. CO₂ and N₂O). Due to gas diffusion in the firm column and the slow bubble enclosure process, fast changes of atmospheric gas concentrations are further smoothed in ice cores (Fig. 2c). Huber et al. (2006) calculated an average duration of δ¹⁰N increase of 225±50 years for D-O events 9-17. δ¹⁰N is controlled by the width in the age distribution of the air enclosed in the ice and by the slow heat conductance in the firm column, which gets rid of the thermal diffusion signal. From the stack of D-O 2-20, we calculate a mean temperature jump of 10.1°C, and a mean CH₄ concentration increase of 70 ppb (Fig. 2c). The increase of atmospheric CH₄ concentration at the onset of a D-O event shows a slight lag of approximately 50 years, relative to the temperature increase recorded in δ¹⁰N in line with the findings of Huber et al. (2006). A direct comparison of gas parameters and ice parameters is difficult, because Δ¹⁸O uncertainty (50-100 years) is comparable to the observed differences of the start of the increase.

The durations of the fast D-O onsets discussed above can be translated into rates of change in CH₄ concentrations and temperature. If we assume that δ¹⁰O documents the temporal change in surface temperature at the ice-core site (approx. 40 years; e.g. Steffensen et al. 2008) and take the δ¹⁰N-derived average temperature increase of 10.1°C, this results in an average temperature increase of 2.5°C/decade. Fig. 2c shows that the increase of CH₄ is slightly faster than δ¹⁰N. Assuming a rise time of atmospheric CH₄ concentration of about 30 years and an increase of 70 ppb, this results in an average rate of change of 23 ppb/decade, however, delayed by a few decades relative to the temperature increase. Comparing these values with modern rates of change (temperature: 0.15°C/decade (global) and 0.46°C/decade (Arctic), last 40 years; CH₄: 48 ppb/decade, last 30 years) shows that Greenland temperature increased considerably faster at the onsets of D-O events than modern temperature does, but modern atmospheric CH₄ concentration is increasing substantially faster than it did during D-O events. This stresses the strength of the anthropogenic CH₄ perturbation in recent decades compared to the most severe natural CH₄ changes, and at the same time illustrates how fast earth climate system variations can occur under glacial boundary conditions.