A short brief on ice core science

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1. Introduction

![Temporal resolution and coverage of different climate archives](image)

Ice sheets are formed by the accumulation of past precipitation events. In the interior of polar ice sheets and also on some high altitude glaciers, where dry snow prevails, a stratigraphically ordered sequence of precipitation events is archived. Using ice cores this climate archive can be accessed.

In principle three different strains of information can be derived from ice cores:

- the ice matrix itself (grain size, crystal orientation, snow accumulation, water isotopic composition)
- particulate and dissolved tracers in the ice
- air in bubble enclosures in the ice

Because annual snow accumulation integrates over many precipitation events, seasonal resolution of ice related climate records is possible in many ice cores (Fig. 1), where the annual snow accumulation is sufficient (typically above 5-10 cm water equivalent (WE)/yr). The gas records allow only a resolution of more than decades because of the slow bubble enclosure process (see below). The maximum time covered by ice cores (Fig. 1) is dependent on the ice thickness, the local accumulation rate, basal melting and the undisturbed flow of the ice at the drill site. The oldest stratigraphically ordered ice to date is derived from the EPICA (European Project for Ice Coring in Antarctica) ice core at Dome C (Fig. 2) and is about
800,000 years old [EPICA community members, 2004; Jouzel, et al., 2007]. The oldest ice in Greenland is “only” 125,000 years old and comes from the NGRIP (North Greenland Ice core Project) ice core. With these features and the wide range of physical and chemical parameters, ice cores represent the ideal climate archive to reconstruct climate variability and climate forcing factors such as greenhouse gases and solar activity.

Due to glacier flow (and in the top 50-100 m due to firnification) annual layers in the ice become thinner with depth (Fig. 3). This restricts the resolution of ice core parameters in deeper ice and makes dating of ice cores (a crucial prerequisite for the interpretation of climate records) a difficult task. The following dating techniques are applied:
Fig. 4: Comparison of a simple Nye and Dansgaard-Johnsen model with constant A and an independently counted age for the last 50,000 years of the GISP 2 ice core.

- Ice flow models: Using assumptions on the thinning rate \( dw/dz \) or the vertical velocity \( w \) with depth and using reconstructions of the snow accumulation rate \( A \) in the past the age of the ice can be calculated according to:

\[
\int = \frac{1}{H} \int w dz
\]

Where \( H \) is the thickness of the ice sheet after correction for density changes. The most simple Nye model assumes a constant thinning rate and constant accumulation rate. Then

\[
w(z) = -\frac{A}{H} z
\]

And equation (1) can be easily integrated leading to

\[
t(z) = \int \frac{H}{A} \int \frac{1}{z} dz = -\frac{H}{A} \ln \frac{z}{H}
\]

A better approach takes into account that the thinning rate has to decline to zero close to bedrock if the ice is frozen to the bedrock. Assuming that the thinning rate linearly declines to zero below a height \( h \) this Dansgaard-Johnsen model leads to

\[
t(z) = \frac{2H - h}{2A} \ln \frac{2H - h}{2z - h}; \ 0 \leq z \leq h
\]

\[
t(z) = \frac{2H - h}{2A} \ln \frac{2H - h}{h} + \frac{2H - h}{A} \left( \frac{h}{z} - 1 \right); \ h \leq z \leq H
\]

where \( A \) was still assumed to be constant in time to allow for analytical integration of equation (1). Since in reality \( A \) was not constant in the past, integration has to be performed numerically. Generally, the Dansgaard/Johnsen model leads to higher ages for deeper ice than the Nye model (Fig. 4).

- Counting of seasonal cycles in ice cores (\( \delta^{18}O \), chemical tracers (Fig. 12))
• Identification of time markers (volcanic eruptions (Fig. 5), magnetic anomalies, nuclear weapon tests etc.)
• Matching with other already dated climate archives or orbital parameters. This makes climatologically only sense if the phase relationship of the matched parameters is known or a physical process connecting local insolation with the ice core parameter can be firmly established.
• Reliable and accurate radioisotopic clocks for ice cores do not exist to date. $^{14}$C is hampered by in situ production in the air bubbles. Radioisotopic aerosol tracers such as $^{10}$Be have the problem of changing production rates in the stratosphere, changing deposition rates or unpractically long lifetimes. Currently the potential of radioisotope ratios such $^{36}$Cl/$^{10}$Be or $^{26}$Al/$^{10}$Be are explored, where all the production and deposition processes should cancel out and effective lifetimes would be shorter.
2. Stable water isotopes

The typical standard parameters measured on ice cores are the stable water isotope ratios $R$, i.e. the ratio of the components $\text{H}_2^{16}\text{O}$, $\text{H}_2^{18}\text{O}^{16}\text{O}$ and $\text{HDO}$. In the well mixed ocean the abundance of these components is approximately

$$1,000,000:2000:310$$

However, during phase transitions (evaporation, condensation) this ratio changes due to equilibrium (different water vapor pressure of the components) and kinetic (different diffusion constants of the components) fractionation processes. Equilibrium fractionation is reversible (i.e. evaporation followed by condensation at the same temperature leads to the same isotopic ratio), while kinetic processes are irreversible and always lead to a depletion of the heavier isotopes in the product of the process.

Because of technical reasons isotope ratios are not determined directly but relative to a standard water (SMOW: Standard Mean Ocean Water) using mass spectrometry. Accordingly, the so called $\delta$-notation is used

$$\delta D = \frac{R_D}{R_{\text{SMOW}}} - 1 \quad \text{bzw.} \quad \delta^{18}O = \frac{R^{18}}{R_{\text{SMOW}}} - 1 \quad \text{in} \quad [\%]$$

During evaporation of ocean water vapor pressure dependent equilibrium fractionation occurs depleting the heavier isotopes in the vapor with a temperature dependent equilibrium fractionation factor (Tab. 1)

$$\alpha_e = \frac{R_D}{R_v} = \frac{p_s}{p_s'}$$

<table>
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<th>Temperature (°C)</th>
<th>$\alpha_e$ (HDO)</th>
<th>$\delta$D (%)</th>
<th>$\alpha_e$ (H$_2^{18}$O)</th>
<th>$\delta$^{18}O (%)</th>
<th>$\delta$D/$\delta^{18}$O</th>
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</table>

In addition kinetic depletion occurs because of undersaturation of the water vapor above the ocean surface. This kinetic fractionation can be approximated by

$$\alpha_k = h + \frac{D}{D_{\text{heavy}}} (1-h)$$

Where h is the relative humidity and D are the diffusion constants of the normal and the heavy water isotopes which are dependent on the mol mass of the gases

$$\frac{D}{D_D} = \frac{(M + M_{\text{air}})M_D}{(M_D + M_{\text{air}})M}$$
The total fractionation factor is

$$\alpha = \alpha_e \alpha_k = \frac{R_{SMOW}}{R_{v,e}} \frac{R_v}{R_c}$$

During condensation of water vapor the opposite equilibrium fractionation occurs, i.e. the condensate is enriched in the heavier isotope compared to the vapor. In clouds where only liquid and water vapor coexist no kinetic fractionation occurs. However, if there is liquid, vapor and ice phase present, the ice is undersaturated relative to the vapor and diffusion, hence kinetic fractionation occurs also during snow formation which depletes the ice in the heavier isotope. This process becomes important for very cold temperatures.

When the water vapor is transported away from its source and cools new precipitate is formed and rained out of the cloud. This precipitate is enriched in the heavier isotope relative to the water vapor but leaves the vapor more depleted. The more water is rained out of the cloud the more depleted becomes the vapor and thus, also the freshly formed precipitate. Accordingly, the $\delta$-values of both vapor and precipitate become increasingly negative the more water vapor is removed from the cloud, i.e. the more the cloud is cooled. This process is called Rayleigh distillation (Fig. 6).

Assuming a constant fractionation factor and no kinetic fractionation ($R_c = \alpha_e R_v$) during condensation this problem can be easily solved using the balance of isotope changes

$$d(N_v + N_c) = dN_v + dN_c = 0$$

$$d(R_v N_v + R_c N_c) = N_v dR_v + R_v dN_v + N_c dR_c + R_c dN_c = 0$$

Where $N$ is the number of water molecules in the vapor and condensate and where we assume that all freshly formed condensate is removed from the cloud as rain ($N_c = 0$). This leads to the differential equations
Fig. 7: Geographic distribution of average δ¹⁸O-values in ice cores and the mean annual temperature at the ice core site [Dansgaard, et al., 1973].

\[
\frac{dR_v}{R_v} = \frac{(\alpha - 1)dN_v}{N_v} \quad \text{bzw.} \quad \frac{dR_c}{R_c} = \frac{(\alpha - 1)dN_c}{N_c} \\
\Rightarrow \quad \frac{R_s}{R_{s,0}} = \left( \frac{N_v}{N_{v,0}} \right)^{\alpha - 1} = f^{\alpha - 1} \quad \text{bzw.} \quad \frac{R_c}{R_{c,0}} = \frac{\alpha}{\alpha_0} f^{\alpha - 1}
\]

where \( \alpha \) was assumed to be constant during the distillation process. In reality the differential equation has to be integrated numerically specifying the temperature history, thus, the change of \( \alpha \) along the water vapor pathway.

With \( R_{c,0} = R_{SMOW} \) and \( R_{v,0} = R_{SMOW}/\alpha_0 \) follows

\[
\delta_v + 1 = \frac{1}{\alpha_0} f^{\alpha - 1} \quad \text{bzw.} \quad \delta_c + 1 = \frac{\alpha}{\alpha_0} f^{\alpha - 1}
\]

where

\[
f = \frac{N_v}{N_{v,0}} = \frac{p_s}{p_{v,0}} = e^{\frac{L\Delta T}{R T_0}} \Rightarrow \quad \delta_c + 1 = \frac{\alpha}{\alpha_0} \left( e^{\frac{L\Delta T}{R T_0}} \right)^{\alpha - 1}
\]

with \( p_s \) the saturation water vapor pressure. This implies that at first order

\[
\delta_c + 1 \approx \frac{\alpha}{\alpha_0} \left( 1 + (\alpha - 1) \frac{L\Delta T}{R T_0} \right)
\]

and, thus, that the δ-value is proportional to the temperature difference between the ocean and the ice sheet. Because the ocean temperature changes little compared to the ice sheet temperature, this temperature difference is usually dominated by changes in the ice sheet temperature. Accordingly, spatial changes in ice sheet temperatures are linearly
More information can be deduced from the water isotopes using the joint interpretation of $\delta^{18}$O and $\delta$D as combined in the deuterium excess $d = \delta D - 8 \delta^{18}$O. Due to the different temperature dependency of the equilibrium fractionation of deuterium and $^{18}$O during evaporation as well as due to the different kinetic fractionation of both species, the deuterium excess gives mainly information on the evaporation conditions of the water vapor (i.e. evaporation temperature and relative humidity over the sea surface). E.g. for sea surface temperatures warmer than 20°C the equilibrium fractionation (Tab. 1) leads to a positive deuterium excess (in Greenland the deuterium excess changes by about 0.35 ‰/°C [Johnsen, et al., 1989], Fig. 8). Using both $\delta$-values as well as simple isotope models this can be used to disentangle temperature changes at the ocean source and on the ice sheet [Stenni, et al., 2003].

The temperature dependence of the $\delta$-values is not only reflected in the geographical variation but also in a strong seasonal cycle and also in pronounced changes on glacial/interglacial time scales. Whether the spatial $\delta$/temperature gradient can be applied on temporal variations in the $\delta$-values can not be taken for granted. Modeling studies for East Antarctica show that the spatial gradient represents a very good calibration for temporal changes [Jouzel, et al., 2007] but that in Greenland this is not the case for glacial temperature changes. The reason is most likely a lack of winter precipitation for cold conditions leading to a bias of the $\delta$-record to warmer summer temperatures [Dahl-Jensen, et al., 1998; Masson-Delmotte, et al., 2005] as well as significant changes in the water vapor source temperature. Another complication for seasonal changes arises from diffusion of water molecules in the ice lattice and in the open pore space of the firm column. This leads to a strong smoothing of the $\delta$-record which is most pronounced for low accumulation sites.
The long-term temperature variation is reflected in the deep ice cores from Greenland and Antarctica. The EPICA Dome C ice core covers 8 glacial cycles and shows orbital climate variations with glacial/interglacial temperature changes of approximately 8°C and a tendency to cooler interglacials in the warm periods prior to 450,000 a BP (Fig. 9). Using the deuterium excess it could be shown that also the water vapor source regions were subject to temperature changes, which however are less pronounced (about 2-3°C) and not in phase with those on the ice sheet (Fig. 10). This implies that most likely obliquity driven changes in water vapor transport occurred which changed the latitude of the effective water vapor source for East Antarctic precipitation over time [Stenni, et al., 2003; Stenni, et al., 2001; Vimeux, et al., 1999].

The most conspicuous features of the Greenland record (Fig. 11) are the rapid climate changes (Dansgaard/Oeschger events) observed during the last ice age, while the Holocene is characterized by very subtle climate variations[North Greenland Ice Core Project members, 2004]. Due to the shift in precipitation seasonality it is difficult to directly ascribe temperature changes to these δ18O changes, however other methods (borehole temperature [Dahl-Jensen, et al., 1998] and gas thermometry (see below [Landais, et al., 2004]) show that the Dansgaard/Oeschger were connected to Greenland temperature changes of 10-15°C in only a few decades and that the glacial was about 20°C colder than the Holocene. The longest Greenland ice core from NGRIP reaches back to the very end of the Eemian, the penultimate warm age. Here, temperatures were about 5°C warmer than today [North Greenland Ice Core Project members, 2004].
Fig. 10: Reconstruction of site and source temperature using $\delta D$ and the deuterium excess from the EPICA Dome C ice core. Note, that the site reconstruction using the spatial $\delta$temperature gradient does not differ significantly from the deuterium excess corrected reconstruction [Stenni, et al., 2003].

Fig. 11: $\delta^{18}O$ profile of the North Greenland NGRIP ice core [North Greenland Ice Core Project members, 2004]
3. Aerosols in ice

Besides their active role on the radiation balance of the Earth, different aerosol species, which are archived in polar ice cores in particulate or dissolved form, represent tracers which can be used to reconstruct climate conditions in their source regions as well as large-scale atmospheric transport patterns. The life cycle of an aerosol particle starts with its formation. In the case of sea salt, aerosol production is caused by wind dispersion of sea water in liquid and frozen state. Dispersion of wave crests and the bursting of small air bubbles at the ocean surface producing small droplets represent an efficient way of producing sea salt particles, which may be uplifted by convective cells in the boundary layer. In high latitudes sea salt aerosol additionally derives from the sea ice surface. For example frost flowers, which form during the refreezing of open leads and polynyas within the sea ice cover, allow for the entrainment of sea salt particles into the atmosphere [Rankin, et al., 2000; Wagenbach, et al., 1998]. In the case of mineral dust, chemical and physical weathering of crustal material leads to the formation of small dust particles which may get entrained into the air during high surface winds. Larger particles may be horizontally accelerated by surface wind and collide with other particles, an avalanche process called saltation. The impact is able to provide vertical momentum to the smaller particles and leads to entrainment into the boundary layer and subsequently efficient uplift by convection. In contrast, sulfate aerosol is mainly produced by gas/particle conversion during the oxidation of the precursor gas SO\textsubscript{2} in the atmosphere, which is emitted by fossil fuel burning, volcanoes or by biogenic activity in the ocean and on land. In contrast to dispersion aerosol which is on the order of a few \( \mu \text{m} \), particles formed from gaseous precursors are on the order of nm and grow by coagulation. Distant to the initial aerosol source, e.g. on the polar ice sheets, “aged” aerosol particles are typically 1-2 \( \mu \text{m} \), where larger dispersion particles have already been removed by dry deposition and nucleation particles have substantially grown.

The sea salt and mineral dust load in the atmosphere is strongly influenced by the climate conditions in the source region and aerosol records provide information e.g. on cyclonic activity, wind speeds and, in the case of dust, on soil properties such as aridity or vegetation cover. In the case of the mineral dust sources, which are located in arid and semiarid regions in both hemispheres far away from the polar ice sheets [Prospero, et al., 2002], dust storm events provide the energy to lift dust particles to the high troposphere above the cloud level, where their atmospheric lifetime is large enough to survive long-range transport. Sea salt aerosol formation is closely linked to cyclonic activity, providing high wind speeds on the ocean surface for efficient sea salt aerosol formation.

The long-range transported aerosol is wet and dry deposited onto the ice sheets, where a stratigraphically ordered sequence of precipitation and aerosol deposition events is formed. The total average aerosol deposition \( J_{\text{tot}} \) is the sum of dry and wet deposition, which can be approximated by:

\[
J_{\text{tot}} = c_{\text{ice}} A = J_{\text{dry}} + J_{\text{wet}} = v_{\text{dry}} c_{\text{air}} + W c_{\text{air}} A
\]

and thus

\[
c_{\text{ice}} = W c_{\text{air}} + v_{\text{dry}} c_{\text{air}} \frac{1}{A}
\]
Where $c_{\text{ice}}$ is the measured average ice concentration, $c_{\text{air}}$ the atmospheric aerosol concentration, $v_{\text{dry}}$ the dry deposition velocity and $W$ an empiric scavenging efficiency for a specific aerosol. Formula (2) shows that even when $c_{\text{air}}$ remains constant a climatological change in $A$ (spatially or temporally) leads to a change in the flux and the ice concentration. Some aerosol species (such as nitrate) are not conservatively trapped in polar ice, i.e. they can partly reevaporate from the snow surface. For those species a quantitative interpretation of past aerosol changes is very difficult.

Due to changing atmospheric concentrations (e.g. in the case of sea salt due to increasing altitude above sea level or due to changes in the source strength) and changing deposition conditions (change in snow accumulation changes the relative contribution of wet and dry deposition) aerosol concentrations show pronounced spatial and seasonal variations. Accordingly, high resolution measurements of chemical aerosol species allow counting of annual layers and are of greatest importance for the dating of ice cores (Fig. 12).

Long-term records in aerosol species document changes in atmospheric aerosol concentrations such as the increase in anthropogenic sulfur and nitrogen emissions due to fossil fuel burning (Fig. 13) [Fischer, et al., 1998; Mayewski and Legrand, 1990]. Also on glacial/interglacial time scales sea salt and mineral dust aerosol show order of magnitude increases in cold climate periods, mainly linked to changes in the aerosol source strength both in Greenland (Fig. 14) and Antarctica (Fig. 15) [Fischer, et al., 2007b; Mayewski, et al., 1994; Röthlisberger, et al., 2002; Wolff, et al., 2006]. Only sulfate, which in Antarctica is controlled by marine biogenic emissions of dimethylsulfide emitting plankton species does not show a significant change over time in the EPICA Dome C ice core (Fig. 15) [Wolff, et al., 2006].
Fig. 13: Long-term trends in sulfate and nitrate concentration in Greenland ice reflecting the increase in Eurasia and North American emissions in the last century. Differences in the amplitude between different site reflect the different accumulation rates [Fischer, et al., 1998].

Fig. 14: Sea salt and mineral dust concentrations in the GISP2 ice core [Mayewski, et al., 1994]
Also the pronounced climate changes during Dansgaard/Oeschger events are connected to strong changes in sea salt and mineral dust in Greenland. Size distribution measurements indicate that this is mainly due to changes in the mineral dust source strength in the Asian desert regions [Ruth, et al., 2003].
4. Gases enclosed in ice

During the slow firmification process in the top 50-100 m of the ice sheet, the density increases with depth and the open pore space between snow grains is reduced (Fig. 16). Starting at a density of app. 810 kg/m³ first bubbles are occluded from the pore space. Because the air in the pore space communicates with the atmosphere overlying the ice sheet, those bubbles archive past atmospheric composition. Additionally, slightly aged air can be pumped out of the firm column itself. Bubble enclosure is completed when a density of app. 840 kg/m³ is reached. Due to the increasing pressure the air bubbles become gradually smaller with depth. At a depth of around 700-1000 m air bubbles disappear completely and the gases are included in the ice lattice as gas hydrates.

The density changes in the firm column and the slow bubble enclosure process have a series of consequences for the air archived in the bubbles:

- The ice at the depth of the bubble enclosure is older than the enclosed air. This ice age/air age difference $\Delta$age has to be taken into account when comparing e.g. greenhouse gas concentration and temperature records from the same ice core.

- The bubbles at a certain depth are not occluded at the same time. This implies that in a given sample the age of the air in individual bubbles is different. Effectively this leads to a low pass filtering of atmospheric records in ice cores.

- The air at the depth of bubble enclosure communicates with the atmosphere via the open pore space. Over large parts of the firm column the gas transport is diffusive. This leads to gravitational fractionation as well as thermal diffusion effects, which are important when considering isotope records in atmospheric gases. The induced change on greenhouse concentrations, however, is rather small.

The gravitational enrichment of heavier gases/isotopes at a certain depth $z$ below the snow surface can be deduced from the barometric formula.
Thermal diffusion occurs when a strong temperature gradient is sustained along the firn column (e.g. at the onset of Dansgaard/Oeschger events where heat conduction along the firn column is slow compared to gas transport through the open pore space). In that case the heavier isotopes are enriched at the colder side of the gradient. The fractionation of two different isotopes is given by

$$\Delta \delta + 1 = \frac{R(z)}{R_b} = e^{-\frac{\Delta M g_z}{RT}}$$

where $R_t$ and $R_b$ are the isotope ratios at the top and the bottom of the firn column and $T$ the respective temperatures. The thermal diffusion factor $\gamma$ is dependent on the gas species (e.g. $\gamma \approx 0.0033$ for $^{14}\text{N}_2$ and $^{14}\text{N}^{15}\text{N}$) and has to be determined experimentally. The thermal diffusion effect is used to determine the temperature change during rapid climate variations (Fig. 17).

Direct atmospheric records of greenhouse gases are restricted to the last few decades. In fact, most of what we know about the concentration changes of those gases over the last centuries is derived from ice core measurements. Fig. 18 shows the 100 ppmv increase of CO$_2$ over the last 200 years due to anthropogenic fossil fuel burning emissions as clearly supported by the accompanying change in $\delta^{13}\text{C}$. Note, the very good agreement of ice core and direct atmospheric measurements where they overlap, providing experimental proof of the reliability of ice core gas records. In the case of methane the anthropogenic increase is even more severe ranging from about 700 ppbv during preindustrial times to 1800 ppbv today.

During glacial times greenhouse gases show generally much lower concentrations of 180 -190 ppmv and 350 ppbv for CO$_2$ and CH$_4$, respectively (Fig. 19). In the case of CO$_2$ the reason for this is mainly a higher carbon storage in the deep ocean due to reduced Southern Ocean
mixing, changes in the carbonate equilibrium in the ocean and in the marine biogenic fixation of CO₂ [Köhler and Fischer, 2006] due to dust induced iron fertilization. In the case of CH₄, lower concentrations are mainly due to a decline in (boreal) wetland methane emissions as reflected in the decline of the interhemispheric methane gradient [Chappellaz, et al., 1997] and isotopically heavier atmospheric CH₄ for cold climate periods [Fischer, et al., 2007a].

Note,
that the interglacials prior to 450,000 a BP show reduced CO₂ and to some extent also CH₄ concentrations. At the same time temperatures were relatively cool, showing that the temperature dependency of both greenhouse gases remained essentially the same [Siegenthaler, et al., 2005; Spahni, et al., 2005] over the last 650,000 years.

Investigating the bipolar coupling of glacial/interglacial climate changes requires establishing a joint time scale for Greenland and Antarctic ice cores. Due to the fact that the interhemispheric mixing time is short relative to the atmospheric lifetime of methane and due to the fact that methane shows rapid variations in parallel to Dansgaard/Oeschger events (Fig. 20), Greenland and Antarctic methane records can be unambiguously matched [Blunier and Brook, 2001; EPICA community members, 2006] providing a common gas time scale. Note, that CO₂ shows only subdued changes during Dansgaard/Oeschger events, implying that those northern hemisphere climate changes have no strong direct effect on the global carbon cycle.

Calculating the climate dependent ice age/gas age difference using firnification models, also a common ice age scale can be established. This has been achieved over the last 55,000 years for the EPICA Dronning Maud Land and the NGRIP ice cores, showing for the first time that each Dansgaard/Oeschger event has a counterpart (Antarctic Isotope Maximum (AIM)) in the Southern Ocean with Antarctica warming when Greenland is in a cold stadial state and Antarctica cooling when Greenland is in a warm interstadial (Fig. 20). The current theory is that this “bipolar seesaw” is transmitted by changes in the Atlantic Meridional Overturning Circulation. This circulation is reduced in stadials exporting less heat from the South to the North Atlantic. Accordingly, the North Atlantic is cold while heat accumulates in the large Southern Ocean heat reservoir slowly warming this ocean basin. The amplitude of the

Fig. 20: Comparison of Antarctic and Greenland ice core temperature records after CH₄ synchronisation [EPICA community members, 2006]
Southern Ocean warming is then dependent on the length of the stadial in Greenland [EPICA community members, 2006]. During interstadials the opposite effect occurs, The CO2 concentrations show an accompanying change with the largest AIM, supporting the close link between Southern Ocean hydrography and biology and atmospheric CO2.

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