



Measurements and analysis of stable isotopes from the 2012 South Dome ice core

Niels Ole Ørum

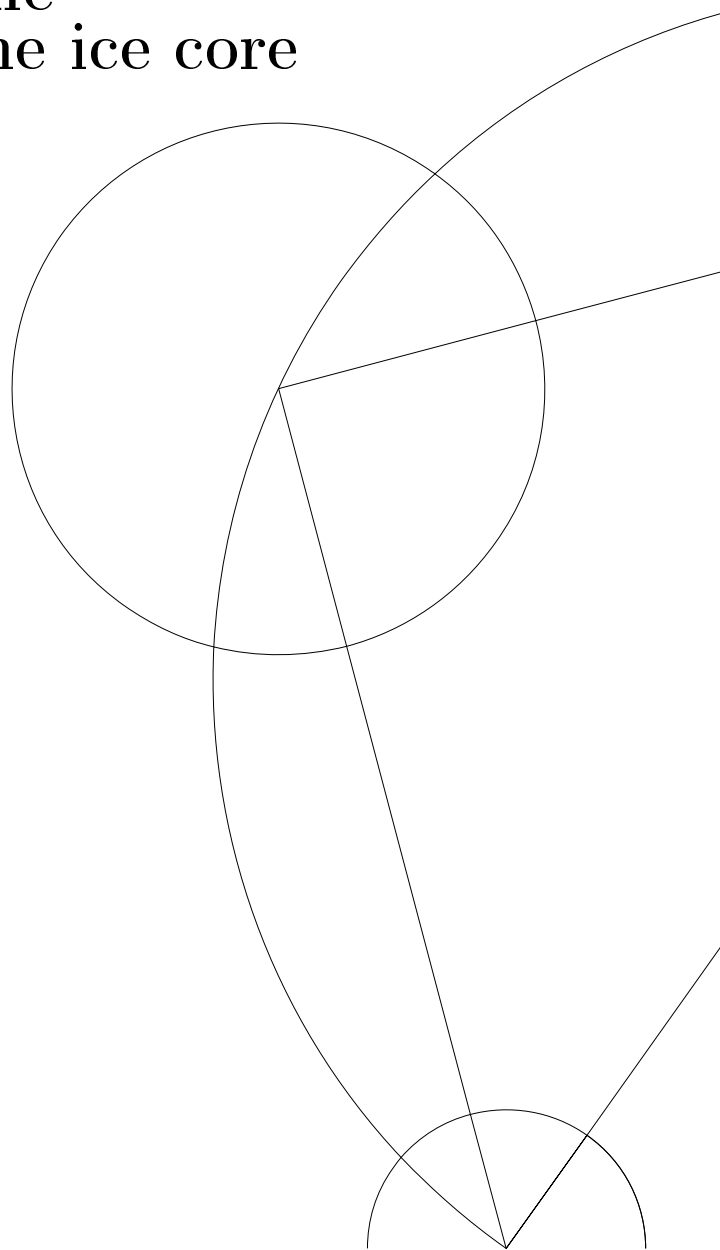
Bachelor's project in Physics

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Abstract

In order to make a full chronology of the 2012 South Dome shallow ice core, part of the core needed to be analysed for its isotopic composition. This has been done using laser spectroscopy with a sample size of 5 cm. In this way the core has been dated back to 1957. With knowledge of the depth/age relationship, average yearly $\delta^{18}\text{O}$ -values has been calculated. And using a Herron and Langway model, yearly accumulation rates has been estimated. Together with temperature records dating back to the year 1895 taken from the coast of southern Greenland, correlations between coastal temperatures, accumulation and $\delta^{18}\text{O}$, have been studied. Findings reveal a strong correlation between $\delta^{18}\text{O}$ and coastal temperatures. Correlations between accumulation and coastal temperature, as well as accumulation and $\delta^{18}\text{O}$ -values have, however, not given any statistically significant results. This is speculated to be related to the drill site's location on the ice divide.

Resumé

For at opstille fuldendt kronologi af iskernen boret på South Dome i Grønland i 2012 var det nødvendigt, af analysere dele af kernens isotopsammensætning. Dette er gjort med laserspektroskopi, med en prøvestørrelse på 5 cm. På denne måde er iskernen dateret tilbage til 1957. Med kendskab til dybde/alder sammenhængen, er de årlige $\delta^{18}\text{O}$ værdier udregnet, og Herron og Langway modellen er anvendt til at estimere den årlige akkumulation. Korrelationer mellem disse parametre, samt temperaturer fra kysten i Sydgrønland, målt tilbage til år 1895, er herefter blevet undersøgt. Der har vist sig at være en stærk sammenhæng mellem $\delta^{18}\text{O}$ og temperatur. Korrelationen mellem kysttemperaturer og akkumulation, samt akkumulation og $\delta^{18}\text{O}$ har ikke givet nogle statistisk signifikante resultater. Årsagen til dette, kan være relateret til iskernens placering, i forhold til isdeleren.

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1 Introduction

In 2012 a shallow ice core of ~ 45 m was drilled at the South Dome of the Greenland ice cap. The site is at $63^{\circ}32'N, 44^{\circ}34'W$, with an elevation of 2854 m see figure 7. (Lawson et al., 1982) At the same site a core was drilled in 1975, this core reached a depth of 80 meters.

In recent years the temperature in Greenland has been rising. To predict how this will affect the Greenland ice cap requires knowledge of the paleoclimate. A widely used proxy for temperature, going back hundreds of years, is the isotopic composition of the ice. This thesis will try to determine whether warmer years are measurable in the isotopic composition of the 2012 South Dome ice core. Much of the core has already been analysed, either by continuous flow analysis or laser spectroscopy of discrete samples. These methods are described in the following section. The experimental part of this thesis has been to cut, melt and analyse the core from 3 to 12 m. From the experimental data it will be explained how the entire core is dated.

With knowledge of the depth/age relationship, rates of yearly- precipitation and isotopic composition can be estimated. These are compared to annual mean coastal temperature measured at three different meteorological stations along the coast of southern Greenland.

How the snow on Greenland is densified into glacial ice is explained and modelled using a Herron and Langway densification model. Knowledge of densification is needed along with the depth/time relationship to estimate yearly snowfall.

Lastly the isotopic composition values of the 2012- and 1975 South Dome ice cores, are joined to form one long data series. This series is compared to temperature records dating back to the year 1869.

2 Theory

2.1 Stable isotopes

Isotopes are atoms with the same amount of protons, but different numbers of neutrons. As an example one can look at hydrogen, the most common isotope which is ^1H , the "1" noting the atomic mass. We also have the heavier isotope ^2H , named deuterium. The difference in atomic weight effects some physical properties. These properties will be discussed in the following section.

2.1.1 Delta notation

This thesis focuses on isotopes of oxygen bound in water.

The specific isotopes generally used in ice core studies are shown in table 1.

As shown in table 1 only a small fraction the elements are consisting of rare isotopes. When discussing the amount of a certain isotope in a substance, it is common to give the ratio defined as:

$$R = \frac{\text{abundance of rare isotope}}{\text{abundance of abundant isotope}} \quad (1)$$

ex:

$$R^{18}(\text{O}) = \frac{[^{18}\text{O}]}{[^{16}\text{O}]} \quad (2)$$

Oxygen:	¹⁶ O	¹⁷ O	¹⁸ O
Abundance	99.757%	0.038%	0.205%
Hydrogen	¹ H	² H	
Abundance	99.998%	0.012%	

Table 1: List of most common isotopes present in water.(Harris, 2010, table 21-1, p. 508)

However, the values here are usually quite small, and need to be referenced in relation to a standard. To do this, isotopic ratios are referenced in their "delta values"(δ) defined as:

$$\delta = \frac{R}{R_s} - 1 \quad (3)$$

where R denotes the isotopic ratio of the sample in question, and R_s denotes the isotopic ratio of the standard. This means we get values around zero. The ratios variate very little, so δ -values are usually given in ‰(10^{-3}). (Mook, 2000)

In order for the δ -notation to make sense across several groups of researchers, the same standard (R_s from equation 3) must be used. There is general agreement that VSMOW, which is an abbreviation of "Vienna Standard Mean Ocean Water" should be used.

Since VSMOW is expensive too precious to use in every day experiments, it is common for a lab to have other standards available. These local standards are water samples measured carefully with respect to VSMOW. When measurements are made, these are compared to a sample of a local standard but ultimately calibrated to VSMOW.

2.1.2 Isotope effects

There are some differences in the physical properties of different isotopes of the same atom. The reason for this is due to the difference in mass.

The kinetic energy of a molecule is:

$$E_{kin} = k \cdot T = \frac{1}{2} \cdot m \cdot v^2 \quad (4)$$

Where k is Boltzmann's constant, T is the absolute temperature, m is molecular mass and v is average molecular velocity.

Given that some amount of substance has the same temperature, it is apparent from equation 4 that heavier molecules(larger m) must have a lower average velocity(v) in order to posses the same amount of kinetic energy.

A similar principle applies to the binding energy of the molecules. Considering a liquid consisting of both heavy and light isotopes, all at the same temperature, the lighter isotopes will vibrate faster, making them more likely to evaporate. The opposite is also the case, as heavier isotopes are more likely to condensate.

2.1.3 Fractionation

The effects mentioned in section 2.1.2 causes a phenomenon, known as fractionation. Consider a parcel of moist air at low latitudes. The parcel contains n_o moles of $^1H_2^{16}O$ and n_j moles of $^1H_2^{18}O$. The isotopic ratio of the vapour is $R_v^{18}(O) = \frac{n_j}{n_o}$, making $\delta_v^{18}O = \frac{R_v}{R_s} - 1$.

As the parcel moves north it starts to cool, and the water in the parcel starts to condensate. As condensate forms the isotopic ratio of the water left in the cloud changes by a small amount:

$$R_p^{18}O = \alpha \cdot R_v^{18}O \quad (5)$$

Where R_p is the isotopic ratio of precipitation α is a fractionation factor. Expressed in δ this gives:

$$\delta_p^{18}O + 1 = (\delta_v^{18}O + 1) \cdot \alpha \quad (6)$$

The condensate is assumed to immediately fall as precipitation. The fractionation factor α is temperature dependant.

At some point the air parcel floats over an ice cap, where precipitation is deposited and later collected as part of an ice core. By measuring the isotopic composition of the precipitation, it is possible to estimate the temperature at the time when the precipitation left the parcel.

2.2 Dating

When dating an ice core, one marks where one year ends and another starts. There are several ways of determining this, but when snow accumulation is high enough variations in $\delta^{18}O$ is a good choice, due to its close correlation with seasonal temperature. (Rasmussen et al., 2006)

There are many other ways of dating a core apart from isotopic composition of the ice. Sometimes changes in the size of the bubbles in the ice allows for layer counting by eye. Soluble molecules, such as sea salt and nitrate, can be picked up in chemical analysis. Although these show annual tendencies, they are, to a larger extend than isotopic ratio, event based, making them less dependable (Cuffey and Paterson, 2010).

Once annual layers have been estimated, there are global events that can help verify dating across several cores, thereby improving accuracy over all. Such global events can be forest fires that leads to ammonia and black carbon in the ice. Volcanic eruptions deposit acid in the ice, changing its conductivity, which can easily be measured. In newer ice, nuclear weapons tests have deposited radioactive particles in the ice, making it possible to identify specific years.

2.3 Densification

As snow accumulates on an ice cap or glacier, it has a density around 60 kg/m^3 immediately after falling. Snow, as opposed to ice is porous, and air can move freely in between the snow particles. As new snow continues to accumulate on top of the old snow, now called firn, the pressure of the overlaying weight causes densification. As the firn densifies the air in between the crystals can no longer

move freely but is confined to channels, that eventually close and leaves bubbles in the ice. The point where the pores close off is the point where firn turns is classified as ice. The ice can further densify until it reaches the density of old glacier ice at around 917 kg/m^3 (Cuffey and Paterson, 2010). This densification happens in three stages.

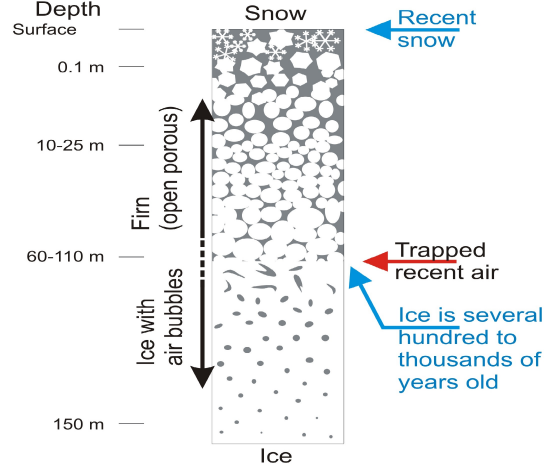


Figure 1: Illustration of the stages of densification. Taken from: http://www.iceandclimate.nbi.ku.dk/research/drill_analysing/cutting_and_analysing_ice_cores/analysing_gasses/firn_zone/

In the first stage, that happens from the top of the ice cap until a density of about 550 kg/m^3 , the dominant mechanism is considered to be packing and settling of ice crystals.

In the second stage densification happens slower and the mechanisms are not well known, this stage ends at "pore close off". Pore close off occurs at a density of around $820\text{-}840 \text{ kg/m}^3$.

Herron and Langway (1980) provide a model that fits densification with depth, considering the factors above.

2.3.1 Herron and Langway

In the following section regarding the Herron and Langway model, it is important to note that all densities are in units of $\frac{\text{Mg}}{\text{m}^3}$.

The model is based on the relationship:

$$\ln \left[\frac{\rho}{\rho_i - \rho} \right] = c_0 \cdot h + c_1 \quad (7)$$

where ρ is the density, ρ_i is the density of glacial ice, h is the depth below the top of the ice sheet and c_0 and c_1 are constants to be estimated. Taking the exponent of each side of the equation yields:

$$\frac{\rho}{\rho_i - \rho} = Z \quad (8)$$

Where $Z = \exp[c_0 \cdot h + c_1]$. Rearranging (8) we get:

$$\rho = \frac{\rho_i \cdot Z}{1 + Z} \quad (9)$$

The term Z takes two forms depending on which stage of densification is being calculated.

For the first stage:

$$Z = \exp \left(h \cdot \rho_i \cdot 11 \cdot \exp \left[-\frac{10160}{R \cdot T} \right] + \ln \left[\frac{\rho_0}{\rho_i - \rho_0} \right] \right) \quad (10)$$

Here R is the gas constant and T is the temperature.

For the second stage of densification:

$$Z = \exp \left(575 \cdot \exp \left[-\frac{21400}{R \cdot T} \right] \cdot \frac{(h - h_{0.55}) \cdot \rho_i}{A^{0.5}} - \ln \left[\frac{0.55}{\rho_i - 0.55} \right] \right) \quad (11)$$

Here A is the water equivalent accumulation, $h_{0.55}$ is the depth of transition from stage one to stage two.

2.3.2 Melt layers

When melting occurs on the top of the ice sheet, water percolates down through the porous firn, and at some point refreezes into solid ice (Cuffey and Paterson, 2010). This behaviour complicates estimates of both density and the depth/time relationship. When the water refreezes it does so as solid ice with a density of around 917 kg/m^3 , far higher than the typical firn.

From figure 2 it is apparent that melt layers can easily be identified in firn



Figure 2: Pieces of an ice core showing melt layers.

cores.

2.4 CRDS (Cavity Ring-Down Spectroscopy)

In recent years CRDS has become a widely used method for determination of isotopic ratios.

The basis for CRDS is light absorption of the molecules being tested. In section 2.1.2 it is mentioned that water molecules with different isotopic compositions have slightly different physical properties, this affects the wavelength at which

the molecule can absorb light.

To measure the absorption of a sample, a small chamber is built of highly

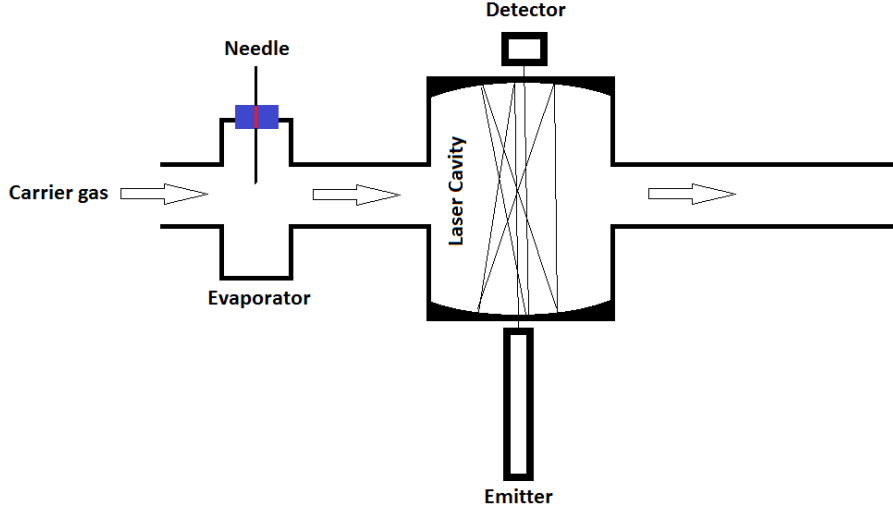


Figure 3: Drawing showing the concept of the CRDS apparatus.

reflective mirrors. A laser sends a beam into the chamber through a small hole, and the light bounces between the mirrors until it hits another small hole where the light intensity is then measured. Due to the reflectivity of the mirrors the travelled distance between the emitter and receiver is very large (~ 10 km), improving precision.

In this analysis the amount of sample inside the chamber needs to be kept constant, and the sample needs to be vaporised. To do this, a small amount of the liquid sample, $5 \mu\text{l}$, is sucked into a syringe, the needle is then pierced through a rubber seal and into an evaporator. The seal is a barrier to keep atmospheric air outside of the instrument. On the other side of the seal is a small hot plate attached to a spring. The plate needs to be hot ($\sim 140^\circ\text{C}$) in order to flash evaporate the sample as it is being injected into the evaporator. If the evaporation is not near instantaneous, fractionation will occur. The function of the spring is allow for the needle to press slightly down on the hot plate to insure contact, without bending.

A steady stream of carrier gas is lead from the evaporator, into the laser cavity. When the sample is vaporised it flows with the carrier gas.

Inside the laser cavity, where the sample now flows through at a known temperature and pressure, the laser is allowed to build up to a specific intensity and is then turned off. As the laser bounces back and forth through the sample, the molecules that are able to absorb the laser do so, weakening the intensity.

The difference in intensity of the light leaving the emitter with intensity I_0 , and the light entering the detector I has the following dependence:

$$I(I_0) = I_0 \cdot e^{-\alpha \cdot z} \quad (12)$$

Where α is the attenuation coefficient of the sample and z is the path length. From equation 12 one can calculate the amount of absorbing sample inside the chamber. This is done with different laser wavelengths, all in the vicinity of 1392 nm, each targeting a distinct isotopic composition. From the different ring-down times, the ratios of the different isotopes are determined. (Brand et al., 2009)

2.5 CFA(Continuous Flow Analysis)

The traditional way of analysing the isotopic ratio of ice cores includes cutting the ice core into pieces, where size determines the spacial resolution.

Many other features of the ice are subject to analysis. Some of these measurements are sensitive to the surrounding environment, and great care needs to be taken in order not to compromise the sample. This can result in a lot of tedious work and discardment of valuable ice.

In order to enhance spacial resolution and allow for more effective use of time and ice, a relatively new method, called CFA, has been put to use. This method uses one long piece of ice, as opposed to several smaller pieces. The ice sample is placed on a melt head and weighted down in order to enhance stability of the melting. The melting head on which the sample is placed is heated up, and the ice starts to melt from one end. The melting head is devised in such a way, that the central $\sim 90\%$ of the sample is guided onwards into different measuring apparatuses. The outer 10% of sample is used in less sensitive measurements. In this way, the centre of the ice core is never exposed to the outside environment, and only the ends must to be cleaned before sampling. Measurements of isotopic ratios with the technique explained in section 2.4 can be included in the analysis program of CFA. In this way measurements are made with a spacial resolution of a few millimetres as opposed to a few centimetres. (Gkinis et al., 2011)

There are however, problems associated with CFA. When analysis of a relatively solid ice core is made, the rate of melt is relatively stable. However, when dealing with the top 10/15 m of a core problems arise. Firn is much less dense than solid ice, but scattered throughout are melt layers. These layers have a density much closer to that of ice, and require a lot more energy to melt. Figure 2 in section 2.3.2 shows an example of such layers, and it is somewhat apparent that the layer is not even throughout the core. This can cause problems because the firn core might not melt evenly, and when that happens, the centre is exposed to outside air that might be sucked into the CFA apparatus in the form of bubbles, which can clog up the system.

2.6 Correlation

Suppose we have two sets of measurements X and Y , both consisting of N numbers: $\{x_1, x_2, \dots, x_N\}$ and $\{y_1, y_2, \dots, y_N\}$ respectively, with mean values, $E(X) = \mu_X$ and $E(Y) = \mu_Y$. If we want to get a measure of whether these two sets of data are independent of each other the covariance can be calculated:

$$Cov(X, Y) = E[(X - \mu_X) \cdot (Y - \mu_Y)] \quad (13)$$

If X and Y are independent then:

$$E[(X - \mu_X) \cdot (Y - \mu_Y)] = E(X - \mu_X) \cdot E(Y - \mu_Y) = 0 \quad (14)$$

If X and Y are not independent the covariance may take positive values if the data sets rise and fall at the same time, and negative values if a rise in X means a fall in Y .

The covariance is difficult to interpret as it is not normalised to the samples in question. For this reason it is common to calculate the correlation:

$$Corr = \frac{Cov(X, Y)}{\sigma_X \cdot \sigma_Y} \quad (15)$$

Where σ_X and σ_Y are the standard deviations of X and Y . (Chatfield, 2013)

Two sets of data can however give a high correlation by chance. To estimate whether a correlation between two data sets X and Y are random, one can calculate the chance that two random sets of numbers are correlated in the same way. This gives a measure of the statistical significance of a correlation.

3 Experiment

In this section, the experimental procedure used to obtain data for this thesis is explained.

3.1 Handling the ice core

When a core in question was drilled, the ice was cut into even 1 m pieces that were put into plastic bags and labelled.

When measuring the ice, it first needed to be cut into appropriately sized samples. The mean accumulation in south Greenland is high, at ~ 0.55 m/year ice equivalent (Herron and Langway, 1980), so a sample size of 5 cm was deemed sufficient.

Cutting ice into pieces was easy on a band saw. However, a saw removes a little bit of ice with each cut, this results in the last sample of each bag being a bit smaller than 5 cm.

After cutting, the samples were put into old tobacco cans for melting. Tobacco cans are made in order to keep their content moist, and as such have an efficient seal. The tobacco cans used for this experiment were very old and the seals were not completely tight. This enabled the sample to interact with the surrounding air. The samples were not in the tobacco cans for very long, and because of this, the error deriving from leaky tobacco cans is considered negligible.

It is important that the samples in the tobacco cans are completely melted before the content is moved into another container, otherwise fractionation would occur. After melting, the samples were shaken lightly to make sure that any condensate or aerosol water in the cans was mixed back in the sample. The samples were then poured into small plastic flasks. These flasks are slightly porous and if a sample is left at room temperature, atmospheric air would interact with the sample, thus contaminating it. This problem was addressed by freezing the sample until the analysis was carried out. When a sample is frozen the amount of sublimation is negligible.

For the samples to go into the autosampler they needed to be thawed and transferred into vials. These vials have special lids that the autosampler can pierce through in order to draw a sample.



(a) Band saw and core ready to be cut. (b) Tobacco can and flask used to melt and store samples.

Figure 4: Pictures taken during preparation of the samples.

3.2 Measuring isotopic ratio

The instrument used to measure the isotopic ratio of the South Dome ice core uses the method described in section 2.4. As described there is a seal the needle of the autosampler needs to pierce in order to inject the sample into the evaporator. This seal was replaced with a new one to avoid contamination of the samples.

When a new set of sample are being analysed, the first to be injected, are three standards. When the machine measures a δ -value, this value is offset from the actual value. This is why the three standards are injected. Two standards were chosen, one with a δ -value higher and one lower, of the expected measured sample values. These together form a straight line is used to estimate the offset of the machines internal calibration. The third standard is injected in order to estimate the validity of the calibration.

The three standards used has δ -values that are far from each other, this is a problem because a tiny amount of residue left inside the needle from a previous run can alter the values of the subsequent runs by a bit. To minimize this effect, when the sampler goes to extract a sample, it rinses the syringe in to the next sample. The amount of sample in each vial is 2ml, for each injection $5\mu\text{l}$ is needed. The tiny amount of residue from the needle that is injected into the sample in this was is disregarded due to its infinitesimal amount. Although the syringe is rinsed for every sample, memory from previous injections is detectable, to further minimize this, each sample is injected multiple times and the first few injections are discarded.

At the end of each complete set of samples, the third standard is again injected to make sure the calibration line still holds up. In the way $\delta^{18}\text{O}$ -values can be measured down to an uncertainty of $\pm 0.1\%$.

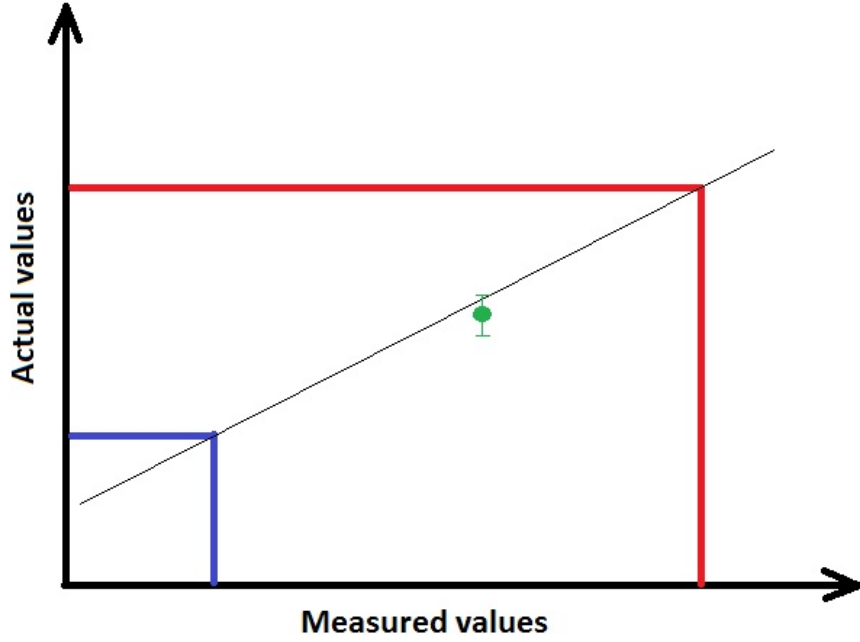


Figure 5: Drawing of how calibration works. Blue and red are calibration standards used to draw the calibration line shown in black. The green dot is the third standard, used to validate the calibration line and calculate uncertainties. All measured samples are calibrated on to the black line giving the actual δ -value.

4 Data

4.1 Dating the core

Dating the 2012 South Dome ice core was done, primarily, based on the $\delta^{18}\text{O}$ measurements.

The marker points were set by eye, and later compared and tuned to fit recognizable features in the 1975 ice core. The 1975 ice core had already been dated based on many factors including a certain yearly layer easily recognisable due to radioactive residue left behind by a nuclear weapons tests.

Figure 6 shows the complete $\delta^{18}\text{O}$ series for the 2012 South Dome core. The series is made from different data sets, both discrete and continuous. All the sets were interpolated linearly to millimetre precision, and where the sets overlapped, the mean was calculated. However, where the discrete data overlapped the CFA data, date markers were set according to the discrete data. This is because CFA had problems melting the firn core in a stable manner, making depth marking of the CFA data difficult.

From figure 6 it is apparent that part of the data is missing just before the 20 m mark. That part of the data is from CFA where there was a problem with the melting of the core. The missing data is considered to be confined to one year of accumulation.

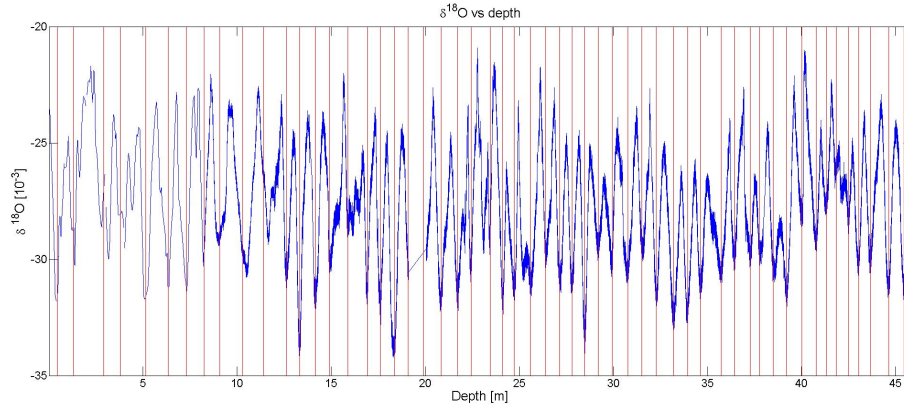


Figure 6: Complete $\delta^{18}\text{O}$ record for the 2012 South Dome core in blue. Red vertical lines show estimated yearly transitions

From figure 6 one can, by counting, see that the core goes back 54 entire years, back to the year 1958. However, the year 1992 does not have a full $\delta^{18}\text{O}$ -series.

4.2 Meteorological Data

Temperature records from three different weather stations on the coast of Greenland have been gathered from Cappelen (2014) in order to compare data gathered from the ice core to temperature data and see if correlations are apparent. The specific stations are shown on figure 7



Figure 7: Location of the three weather stations and the South Dome drill site.

4.2.1 Comparing yearly mean $\delta^{18}\text{O}$ to yearly mean temperature

As explained in section 2.1.3, $\delta^{18}\text{O}$ can be used as a so called proxy, to calculate the temperature at which the snow has condensed in a cloud. The data in figure 6 shows that there is a clear correlation between seasonal temperature and $\delta^{18}\text{O}$ values, enabling the identification of years. Comparison of the yearly mean $\delta^{18}\text{O}$ to the yearly mean temperature gives an idea of whether this proxy also says something about the general temperature at the coast.

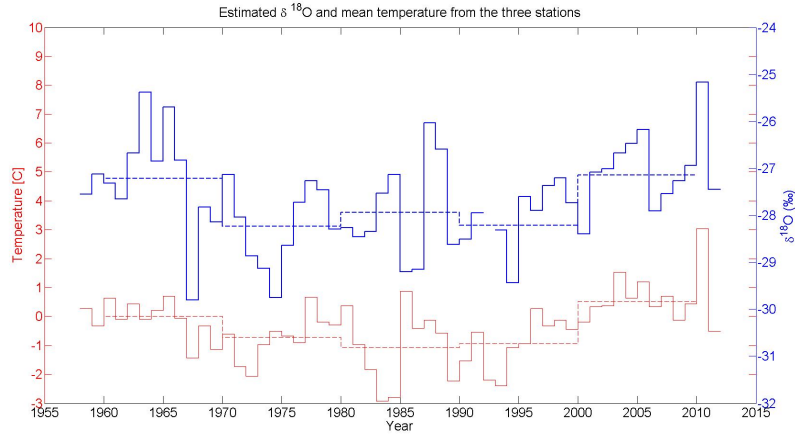


Figure 8: Plot showing yearly mean temperature from the three weather stations along side measured yearly mean $\delta^{18}\text{O}$ measured from the South Dome 2012 core. Dashed lines are decadal averages.

It is clear from figure 8 that coastal temperatures and $\delta^{18}\text{O}$ -values indeed co-vary. To quantify this, the correlation coefficient has been calculated.

<i>Core 2012 $\delta^{18}\text{O}$-values</i>		
	Correlation	Probability of chance correlation [%]
<i>Nuuk (1957-2011)</i>	0.45	0.07
<i>Narsarsuaq (1957-2011)</i>	0.52	0.01
<i>Tasiilaq (1957-2011)</i>	0.40	0.34
<i>Mean of 3 stations (1957-2011)</i>	0.42	0.18
<i>Mean of 3 stations (10 year mean) (1960-2009)</i>	0.90	3.67

Table 2: Table showing the correlation between measured $\delta^{18}\text{O}$ -values in the ice core and temperature measured at the stations shown in figure 7.

As expected from figure 8, table 2 confirms that there is a definite correlation, the significance of which to be discussed.

4.3 Ice core density

The ice core was drilled in one meter pieces with an even diameter. This made it easy to determine the mean density of each piece. A core made from solid ice is quite sturdy, but a shallow ice core consists of firn, which is more fragile. Little pieces are prone to break off, this along with natural fluctuations and melt layers causes noise in the data.

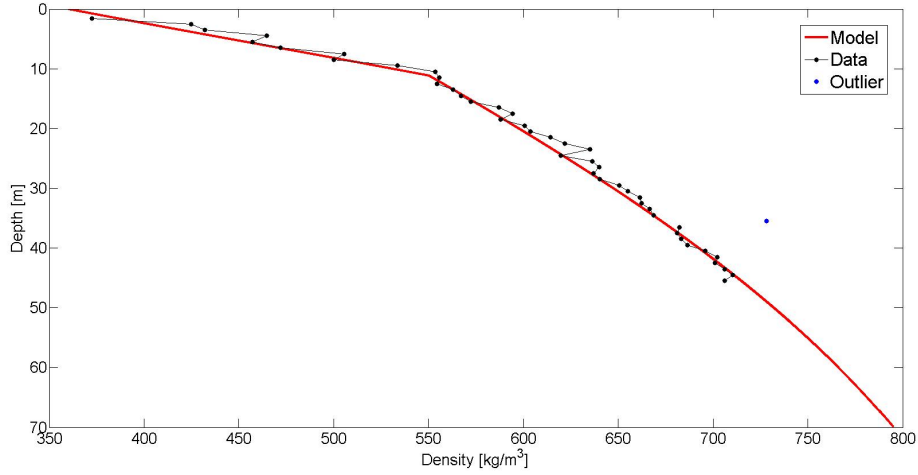


Figure 9: Plot showing the density data collected from the 2012 South Dome core, along with a least square fit based off Herron and Langway (1980).

Figure 9 shows the density data in black marks. The red line is a model based on the Herron and Langway explained in section 2.3.1. For this fit the blue mark is deemed an outlier and discarded. By knowing the density at a given depth and the depth markers for each year, the yearly accumulation can be estimated.

4.3.1 Accumulation

The yearly layers estimated from the $\delta^{18}\text{O}$ are in millimetre precision. Using the Herron and Langway (1980) method a density was estimated for each millimetre of the core. The ice equivalent length was then calculated by numerical integration; by dividing the density for each millimetre with the density of glacial ice and adding together all the points between two markers. Analytically this can be written as follows:

$$A = \frac{1}{\rho_i} \int_{z(a)}^{z(b)} \rho(z) \cdot dz \quad (16)$$

Here A is the accumulation in ice equivalent, ρ_i is the density of glacial ice, $z(a)$ and $z(b)$ are the depths that mark the beginning of year a and b respectively and $\rho(z)$ is the density at given depth z .

4.3.2 Comparing annual accumulation to yearly mean temperature

The amount of accumulation on Greenland has obvious implications on the mass balance of the ice sheet. For this reason a comparison of relationship between yearly mean temperature and yearly accumulation is interesting.

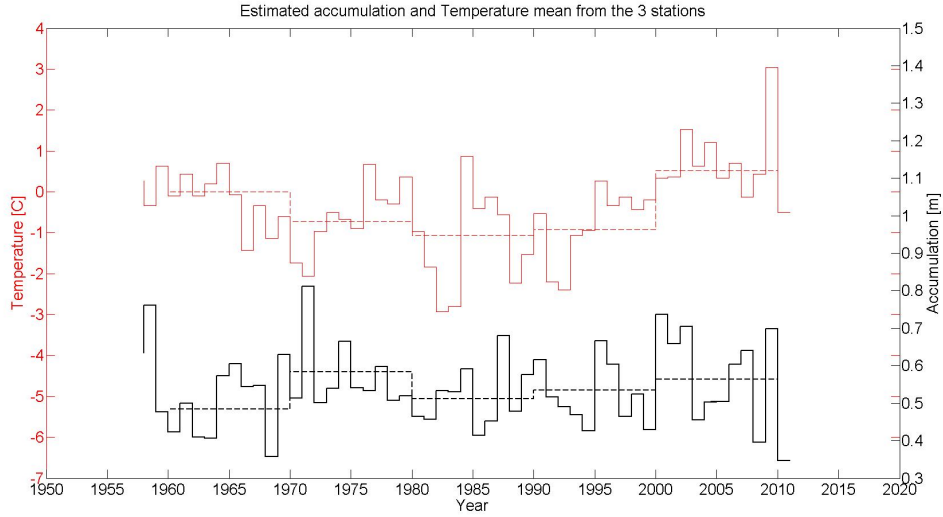


Figure 10: Plots comparing annual mean temperature measured, averaged over the three stations shown in figure 7, to estimate annual accumulation in ice equivalent units at the South Dome. Dashed lines are decadal averages.

<i>Estimated accumulation</i>		
	Correlation	Probability of chance correlation [%]
<i>Nuuk (1957-2011)</i>	0.15	28.70
<i>Narsarsuaq (1957-2011)</i>	0.13	36.76
<i>Tasiilaq (1957-2011)</i>	0.22	11.66
<i>Mean of the 3 stations (1957-2011)</i>	0.17	21.30
<i>Mean of the 3 stations (10 year mean) (1960-2009)</i>	0.0436	94.45

Table 3: Table showing the correlations between estimated accumulation at the South Dome and annual mean temperature measured at the stations shown in figure 7.

It is apparent from table 3 that there is a positive correlation, although it is rather weak, this is probably due, at least in part, to the large amount of natural noise in the yearly accumulation, caused by sastrugi (ripples in the ice).

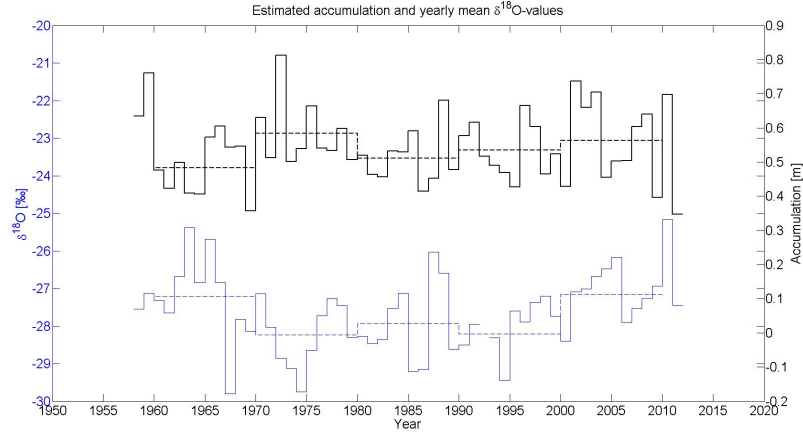


Figure 11: Plots comparing annual mean $\delta^{18}\text{O}$ -values to estimated values of yearly accumulation. Dashed lines are decadal averages.

4.3.3 Comparing yearly mean $\delta^{18}\text{O}$ -values to yearly accumulation

From figure 11 it seems there is some sort of correlation, however, from table 4 it is apparent that this is not the case. It seems that the newer part of the ice core shows more covariability, as opposed to the older part of the ice core. This was tested, but the result did not achieve sufficient statistical significance of 5% probability of chance correlation, which is considered the maximum in this thesis.

<i>Estimated accumulation</i>		
	Correlation	Probability of chance correlation [%]
<i>Mean $\delta^{18}\text{O}$-values (1957-2011)</i>	0.09	53.77
<i>Mean $\delta^{18}\text{O}$-values (10 year mean) (1960-2009)</i>	-0.35	56.22

Table 4: Table showing the correlations between estimated accumulation at the South Dome and annual mean $\delta^{18}\text{O}$. Note that the year 1992 is left out, due to inadequate data(see figure 6).

5 Discussion

Measurements of the $\delta^{18}\text{O}$ -values from 3 to 12 meters of the South Dome 2012 ice core have been done in a sufficient way. This along with measurements of the rest of the core, allows the estimate of a time/depth relationship back to the year 1957 at ~ 45.5 m. There was a problem in the CFA analysis of the ice core that resulted in the loss of $\delta^{18}\text{O}$ -values in what is estimated to be the year 1992. Several more different parameters could have been considered when dating the core, and are available for the part of the core analysed with CFA. However,

dating is still considered to be sufficiently accurate for further measurements.

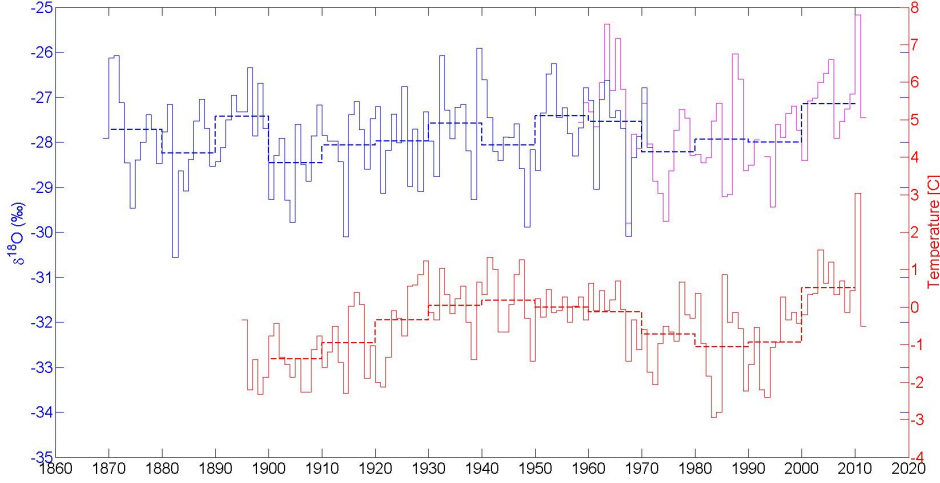


Figure 12: Plots comparing annual mean temperature, averaged over the three stations shown in figure 7, to annual $\delta^{18}\text{O}$ -values at the South Dome. Data in blue is taken from the 2012 ice core. Data in magenta is from the 1975 ice core. And the 10-year means are made with compiled data from the two cores, the mean $\delta^{18}\text{O}$ -values calculated in the overlap. Dashed lines are decadal averages. The 1975 data was taken from Lawson et al. (1982).

In figure 12, the part of the cores that overlap should, ideally, have approximately the same values. Although they have a high correlation coefficient of 0.80, they do vary by more than 1 ‰. This is a large variation considering a measurement accuracy down to $\sim \pm 0.01$ ‰. The difference between the two cores is assumed to be a natural uncertainty, caused by processes such as, unevenness in drifting or sastrugi, summer melting and penetration of the melt into the lower layers, and vapour diffusion, which redistributes the isotopic gradients occurring naturally. (White et al., 2009).

In order to minimise this uncertainty, more cores would need to be drilled at around the same location so the values can be averaged.

The values displayed in tables 3 and 4 are of a too low significance to be analysed further. Buchardt et al. (2012) found a correlation between accumulation and $\delta^{18}\text{O}$, using $\delta^{18}\text{O}$ in stead of temperature. By analysis of 52 different ice cores scattered all over Greenland, it was shown that where the core was drilled, in respect to the ice divide, influenced the correlation. Their findings might explain why there is little or no correlation at the South Dome, since this is in the middle of the ice divide.

Lawson et al. (1982) suggests analysis of isotopes and precipitation by dividing the year in a summer and winter season. This might be interesting to further investigate.

The meteorological data gathered from the coastal stations go back to 1895. The bottom of the 1975 South Dome ice core is estimated to be from 1869. Gathering temperature data from other sites around Greenland could provide

	Correlation	Probability of chance correlation [%]
<i>Mean of 3 stations (1895-1971)</i>	0.34	0.25
<i>Mean of 3 stations (1895-2011)</i>	0.42	< 0.01
<i>Mean of 3 stations (10 year mean) (1900-2009)</i>	0.83	0.08

Table 5: Table showing the correlation between measured $\delta^{18}\text{O}$ -values, and temperature measured at the stations shown in figure 7. (1895-1971) is calculated using data from the 1975 core and (1895-2011) using data from the two cores compiled, taking the mean in the overlap. The 10 year mean is calculated by compiling data from the 1975- and 2012 ice cores.

basis for further analysis.

6 Conclusion

During this project, the South Dome 2012 shallow ice core has been analysed, partly by discrete CRDS, and partly by CFA, thus estimating a time/depth correlation. Further the annual accumulation has been estimated and compared to values such as, coastal temperature and mean yearly isotopic ratios. However, the correlation coefficients, and the significance of these are too low to make any conclusions. Getting a better idea of the temperature/accumulation correlation, if any exist, would require more data in the form of more ice cores.

The $\delta^{18}\text{O}$ -values of the South Dome 2012 shallow ice core show a definite correlation to the South Greenland coastal temperature. The recent rise in temperature is clearly apparent in the $\delta^{18}\text{O}$ values.

Temperature records, used in this thesis, do not date back as long as the oldest part of the 1975 ice core. Gathering temperature data from other stations might provide records dating back further, potentially making better correlations.

Estimates of accumulation across the joined 1975 and 2012 series might make significant correlation possible. Dividing years into a summer and winter season might also reveal season specific variations in accumulation rates and $\delta^{18}\text{O}$ -values.

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