



## In search of stratigraphic subdivision of the period 8–0 ka in Greenland ice cores

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**Abstract:** The  $\delta^{18}\text{O}$  data for the last 8000 years in the Greenland NGRIP1, GRIP, DYE-3 and GISP2 ice cores have been analyzed stratigraphically in search of potentially meaningful boundaries and units. Pattern matching of the profiles is supported by using graphical display enhancements, calculating spectral trend curves and generating a compound profile. Techniques routinely used in subsurface geology have been applied in correlating the profiles. Four major stratigraphic units are identified (8.1–4.9, 4.9–3.3, 3.3–1.9 and 1.9–0.1 ka b2k), resulting in an improved understanding of the climate change after the Holocene Climate Optimum. Correlatable higher-order boundaries are identified within these units. The layers between the boundaries show  $\delta^{18}\text{O}$  patterns which generally are similar in character, the differences being ascribed to lateral variations in the factors that control the isotope content of the ice. The layering forms a series of short-lived low-amplitude aperiodic oscillations on a centennial time scale. The suggestion is that these higher-order boundaries and  $\delta^{18}\text{O}$  oscillations have climatic significance. Equivalent units are tentatively identified in ice-core data from the Agassiz and Renland ice caps. Comparison with other climate proxies or stratigraphies from the Northern Hemisphere is expected to render support for the here proposed scheme. It will then serve to guide and constrain the analysis of the dynamics of the climatic fluctuations for the study period.

Key words: Arctic, Holocene, ice cores, stratigraphy, pattern matching.



## Introduction

This paper reports newly observed patterns in  $\delta^{18}\text{O}$  profiles in Greenland ice cores from the Holocene, specifically from the interval between 8.140 and 0.120 ka b2k. Ice cores from Greenland have yielded much information for the glacial period over recent decades (*e.g.* Lowe *et al.* 2008). However, results from the Holocene part of the ice-core record have appeared to conflict with data from elsewhere, in apparently lacking evidence of the Holocene Climate Optimum (HCO) approximately 9 to 6 ka b2k (Vinther *et al.* 2009). This suggested that Holocene climate was too spatially variable for the recognition of regional trends. The study by Vinther *et al.* (2009) has shown how the HCO was masked in the ice-core data by the effects of changes in ice-surface elevation through time. Correcting the  $\delta^{18}\text{O}$  data for the effects of elevation change has confirmed that millennial-scale climate change through the Holocene (including the HCO) was regionally consistent. This has opened the possibility of exploring the  $\delta^{18}\text{O}$  data for meaningful information at shorter time scales within the Holocene.

In this paper we analyze the  $\delta^{18}\text{O}$  profiles of the NGRIP1, GRIP, GISP2 and DYE-3 ice cores from the main Greenland Ice Sheet (GIS) for the period between 8.140 and 0.120 ka b2k. We also use the  $[\text{Ca}^{2+}]$  data series of GISP2 as an input. In addition,  $\delta^{18}\text{O}$  ice-core profiles from the Renland (East Greenland) and Agassiz (Ellesmere Island, Canada) ice caps (Fig. 1) are investigated.

We have chosen our study period to post-date the well-known and relatively high amplitude events of the early Holocene: the 11.5 ka Preboreal oscillation, the 9.3 ka cold event and the 8.2 ka cold event (Rasmussen *et al.* 2007). The 8.2 ka cold event in particular has been the subject of many studies as a potential analogue for present-day climate change (Alley and Ágústadóttir 2005; Rohling and Pälike 2005; Kobashi *et al.* 2007).

Oxygen isotope ratios ( $\delta^{18}\text{O}$ ) are widely used for their close dependence on the state of the ocean-climate-icecap system. Although once thought to be directly tied to palaeotemperature (Emiliani 1966), it is now generally accepted “that the factor that translates an isotopic change to a temperature change is variable” (Wolff *et al.* 2010; see also Jouzel *et al.* 1997). This leaves unimpeded that an increase of the  $\delta^{18}\text{O}$  is generally considered to correspond with a warming climate and a decrease with a cooling climate, that is to say over the ice-core interval with highly variable

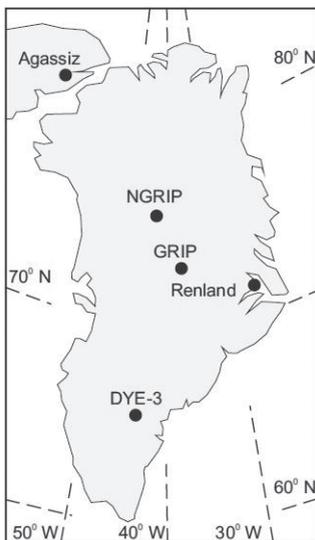


Fig. 1. Location map of the drill sites for the ice cores analyzed in this paper.

$\delta^{18}\text{O}$  records, *i.e.* the period before 8 ka. The work by Vinther *et al.* (2009) has confirmed the reliability of  $\delta^{18}\text{O}$  records as a proxy for temperature fluctuations on a millennial scale for the period following the 8.2 ka cold event. Building on the work of Severinghaus *et al.* (1998), Kobashi *et al.* (2008) have developed a method to derive Greenland surface-snow temperatures from isotopic data of air trapped in the GISP2 ice core, thereby overcoming shortcomings of  $\delta^{18}\text{O}$ -based methods for temperature quantifications.  $[\text{Ca}^{2+}]$  is a measure of atmospheric dust flux and is also accepted as a proxy for the state of climate; refer to De Jong *et al.* (2009) and Rasmussen *et al.* (2014) for examples.

The objective of our paper is to identify boundaries and units in the  $\delta^{18}\text{O}$  data series which have been previously considered to show little variability. Our focus is essentially stratigraphy, *i.e.* the regionally relevant stacking pattern of ice layers resulting from all the (climatic) processes, periodic through non-periodic, that contribute to the  $\delta^{18}\text{O}$  profiles. We consider stratigraphic division not only a tool, enabling identification of boundaries and layers, but also a result: the boundaries and layers are shown in their mutual relationship, including lateral and vertical variations, thus generating new information and insight. The emphasis of our work is on regionally relevant patterns and events of change rather than on alternations of cold and warm periods defined in terms of numerical temperature change. The identification of driving mechanisms with well-defined recurrence times, *e.g.* through a spectral analysis of the time series – the subject of many climate studies – is beyond the scope of our work. Underpinning our approach is the idea that, through a stratigraphic analysis of the records, a framework of boundaries and units may be erected, which in turn could offer guidance and constraints to further work on understanding the climatic information, more specifically the climatic parameters and controlling mechanisms, contained in the GIS isotopic data series. It may also serve as a reference for climate-proxy studies in other areas.

## Data

The data for our study comprises the 20-year average values for  $\delta^{18}\text{O}$  and  $[\text{Ca}^{2+}]$  as made available by Seierstad *et al.* (2014) and by the Centre for Ice and Climate of the University of Copenhagen ([www.icecores.dk](http://www.icecores.dk); see also Vinther *et al.* 2006).  $\delta^{18}\text{O}$  data are available for all ice cores,  $[\text{Ca}^{2+}]$  for GISP2 only. The data for the Renland ice core and for the combined A84/A87 ice cores, which were cut within 30 m of each other, are corrected for Holocene uplift (Vinther *et al.* 2009). Data corrected for site-elevation changes are not available for the GIS ice cores. The data for changes in site-elevation of the DYE-3, GRIP and NGRIP1 ice cores (Fig. 2; see also fig. 2a in Vinther *et al.* 2009) suggest relatively mild thinning of GIS over the last 8000 years. The trends are smooth

and similar in nature; any correction will be small with little impact on the results of the analyses presented in this paper. Note also that we did not apply any data processing such as removing low- or high-frequency variability prior to interpretation (Vaughan *et al.* 2014; see also Bartlein 2007). The GISP2  $\delta^{18}\text{O}$  and  $[\text{Ca}^{2+}]$  data series start at 120 years bk2. Therefore, we have not included the period from 120 years bk2 to today in our analysis.

## Sources of error and uncertainty

As we are targeting relatively high-resolution features of the data, questions of data quality and uncertainties in the time scales for the ice cores are important. We consider here possible sources of error in the ice-stratigraphic record itself, in the coring process, in the dating of the individual cores, and in the matching of the time scale between the various cores.

The GIS ice cores are not considered to suffer from imperfections in stratigraphy over the time period spanned by this study (Rasmussen *et al.* 2006; Vinther *et al.* 2006). The Renland stratigraphy is considered undisturbed and continuous for the Holocene (Vinther *et al.* 2008). The stratigraphic record of the Agassiz ice cores is affected by various amounts of winter scouring of snow by katabatic winds (Fisher *et al.* 1983, 1995), and varying degrees of surface melting, the latter notably between 9 ka and 11 ka (Vinther *et al.* 2008). Net ablation in some periods of the early Holocene cannot be ruled out for the Agassiz ice cap (Fisher *et al.* 1983, 1995).

Snow and firn diffusion, ice-flow related thinning and surface melt adversely affect the vertical resolution of the ice-core stratigraphy and cause damping of temporal variations in  $\delta^{18}\text{O}$ . Deconvolution was applied to the  $\delta^{18}\text{O}$  data from GIS ice cores by Vinther *et al.* (2006) to improve annual layer differentiation, accepting the risk of introducing occasional oscillations of non-annual origin. The 20-year averaging of the  $\delta^{18}\text{O}$  data helps to reduce the adverse effects of damping, with the possible disadvantage that averaging may lead to smoothing of curves and trends, and some loss of vertical resolution.

Regarding the coring process, serious problems pertaining to coring, core quality and core treatment with adverse affects on the depth-time conversion have not been reported and we assume this is not a source of any significant error.

Concerning the depth-to-age conversion of the ice cores, considerable effort has gone into dating the individual cores in recent years. Using the techniques of matching volcanic reference horizons (as represented in electrical conductivity measurements, ECM) and annual layer counting, the GICC05 synchronized stratigraphical time scale for the Holocene was constructed in the DYE-3, GRIP and NGRIP ice cores by Rasmussen *et al.* (2006) and Vinther *et al.* (2006). The time scale over the past 7,900 years is based on annual layer counting of stable isotope data (Vinther *et al.*

2006), while the early Holocene part is based mainly on annual layer counting in chemical impurity data (Rasmussen *et al.* 2006). Seierstad *et al.* (2014) presented an updated synchronization of the NGRIP, GRIP and GISP2 ice cores for the past 104 ka, based on the identification of a large number of stratigraphic markers in a range of chemistry records (chemo-stratigraphic matching) and a small number of tephra marker horizons. Using the new synchronization to evaluate the GICC05 timescale in the Holocene, they concluded that the offsets between the two synchronizations are minor and are too small, on a centennial to millennial timescale, to justify a revision of GICC05; see also below.

Possible uncertainties in the matching of synchronous events in the ice cores is a critical issue for our study. For the DYE-3, GRIP and NGRIP ice cores, Vinther *et al.* (2006) consider the risk of mismatching ECM volcanic reference horizons low, *i.e.* one year at most, reference horizons being available approximately every 50–100 yrs in the data. Within the matched sections, the uncertainty is estimated to be at most 4–5 yrs. The maximum error in the annual layer counting ranges from 0.25% to 2.0% for various intervals of the Holocene (Vinther *et al.* 2006: table 2). Observing a varying density of chemo-stratigraphic match points in NGRIP, GRIP and GISP2 in the Holocene section, Seierstad *et al.* (2014) report that the synchronization uncertainty may be up to 1.5 m or 15 years in the section between 2.7 and 7.2 ka b2k. They also state that the old and new synchronizations agree within one year back to 1.8 b2k. The comparison for the section from 1.8 to 10.3 ka b2k typically shows agreement within 1–2 years, with the exception of the intervals around 5.94, 8.26, 8.97 and 9.60 ka b2k, which show offsets of 3–4 years. A maximum offset of 9 years is reached at 4.26 ka b2k. These uncertainties are acceptable at the resolution of our study, *i.e.* the centennial to millennial scale.

The Holocene in the Agassiz ice cores (A77, A79, A84, A87) and in the Renland ice core has been matched to the NGRIP1 ice core and to the GRIP ice core, respectively, using volcanic markers identifiable in ECM in combination with interpolation of data between the match points by Vinther *et al.* (2008). Multiple sections of annual layer counting were available in A77, providing guidance and constraints to the cross-dating of the Agassiz cores.

The uncertainty for the A77 time scale is assessed by Vinther *et al.* (2008) to be less than 5% for the past 8 ka and 200–300 yrs for the 8–11.7 ka period. For the other Agassiz ice cores the uncertainty is estimated to be up to 400 yrs in the periods 2–4 ka, 5.5–7 ka and 9–11 ka (*i.e.* the periods with few common fix points); it is less than 5% for the past two millennia.

The matching uncertainty in the Renland core is estimated to be about 100 yrs or less, with the exception of the period 5–7.3 ka for which the uncertainty is approximately 200 yrs (Johnsen *et al.* 1992; Vinther *et al.* 2008) and the period A.D. 934 to A.D. 1988 for which the uncertainty is estimated to be a few years only.

## Method

Traditional stratigraphic studies tend to focus on defining type sections, sometimes supplemented by reference sections; for example, the older work by the INTIMATE group on the NGRIP ice core (Lowe *et al.* 2008), broadly speaking, falls into this category. Stratigraphic practice, however, is an ever-difficult reality when dealing with 1D datasets such as those from drilled or cored wells. Generating well-to-well correlation frameworks is routine in subsurface geology. Quantitative data such as natural gamma ray (GR) recorded by wireline-logging tools are analyzed and interpreted (*e.g.* Doornenbal and Stevenson 2010, and references therein). All wells being equally important, the local and/or regional stratigraphic variation is captured in this multi-well approach, leading to a better understanding of the vertical succession and lateral variation than obtained by a single well analysis. It avoids the focus on one well, and it honors the notion that stratigraphy is the outcome of a large number of interacting factors and that the nature and strength of the stratigraphic signal may vary spatially and temporally. The advantages of such an essentially quantitative approach in ice-core studies are demonstrated in a recent publication on the NGRIP, GRIP and GISP2 cores by Rasmussen *et al.* (2014). Somewhat similarly, characterizing spectral properties of ice-core data also tends to focus on individual profiles (Johnsen *et al.* 1997), with the risk of overvaluing the results of a one-trace analysis.

Generally speaking, conventional stratigraphy deals with boundaries and units of layered rocks: a stratigraphic unit is a body of rock that is defined by one or more distinctive features and has more or less distinct boundaries. We apply the same principles in the stratigraphic study of bodies of ice: units are defined which differ from overlying and underlying units on the basis of one or more criteria. We look for patterns of change events in the  $\delta^{18}\text{O}$  profiles that are correlatable between ice cores, and construct a  $\delta^{18}\text{O}$ -stratigraphic correlation framework. Patterns are primarily identified in and matched between ice cores by visual interpretation of the profiles, in an approach similar to that applied in subsurface wireline-log-based well-to-well correlation. Note that correlations are guided and constrained by the volcanic reference horizons (VRH) of Vinther *et al.* (2006). Work is facilitated by using simple graphical display enhancements such as core-specific horizontal scaling and coloring. In addition, we apply the quantitative technique of calculating maximum entropy-based (MEM-based) spectral trend curves of  $\delta^{18}\text{O}$  (generated in the software package CycloLog) as a means to evaluate the validity of the visually matched patterns (see Figs 2 through 5). A powerful feature of the spectral trend curve is that it shows trends and trend changes in the (input) data. These may be overlooked when dealing with low-variability data series. We refer to De Jong *et al.* (2009), Nio *et al.* (2006) and Qayyum and Smith (2014) for an explanation of the spectral trend curve (INPEFA), which is routinely used in correlating wells in the oil/gas industry. Relevant to mention here is the following ‘trait’ of the spectral trend curve.

Because it is always normalized (to lie between values 0 and 1), calculating the curve from part of a data set generally increases the apparent variance compared with the same segment of the data in the curve of the full data set. As a result, the patterns of the curve generated from part of the data set are more pronounced than those of the curve of the full data set (see Qayyum and Smith 2014). We will apply this trait of the trend curve in our analysis of the higher-order patterns of the data series.

The working hypothesis at the start of the interpretation was that climate changes are regional phenomena and that climate proxies such as isotopic profiles from different locations within a region can be expected to show similar, synchronous, patterns (Jouzel *et al.* 1997; Vinther *et al.* 2009). Local factors, however, may lead to spatial variation in the profiles, thereby masking the temporal variation. This problem is likely to increase with decreasing temporal variability in the data and increasing temporal resolution, as is the case for our study interval. The problem may be further compounded by uncertainties in the data and in the dating of the proxies. Conceivably, regionally relevant changes may not show in any of the profiles due to these effects. In our analysis, however, we have assumed that, if a distinct change occurs in one profile, this represents a potential regional climate change. Its regional significance, then, is assessed through correlation with the other profiles, supported by a simple arithmetic data-stacking technique.

The stratigraphic correlation approach is founded on the concept of reproducibility of results. If patterns observed in one ice core are repeated in the same type of data series of one or more other ice cores, they can be considered robust; the more frequent the repetition (in additional ice cores), the more robust the result. In our opinion, this applies not only to data series with high variability (high signal-to-noise ratio) but also to data series with low variability (low signal-to-noise ratio). From a pure statistical point of view this may not be very satisfactory or convincing: statistical significance needs to be demonstrated. It is, however, not a *conditio-sine-qua-non* in stratigraphy. A relatively indistinct boundary or event that is repeated and is considered stratigraphically meaningful, may fail on rigorous statistical testing. The stratigraphical approach in such case will argue that a weak signal may still be a valid signal, because it is repeated in other profiles; the test outcome raises a red flag, it signals that one has to be very cautious and look for corroborative evidence.

## GIS ice cores – results and interpretation

**Major units.** — Figure 2 shows the  $\delta^{18}\text{O}$  profiles and the corresponding spectral trend curves for the four GIS ice cores plotted against the GICC05 time scale. Within the study interval of 8–0 ka, the  $\delta^{18}\text{O}$  profiles of the four GIS cores are apparently rather featureless on a millennial scale. A closer look at the four

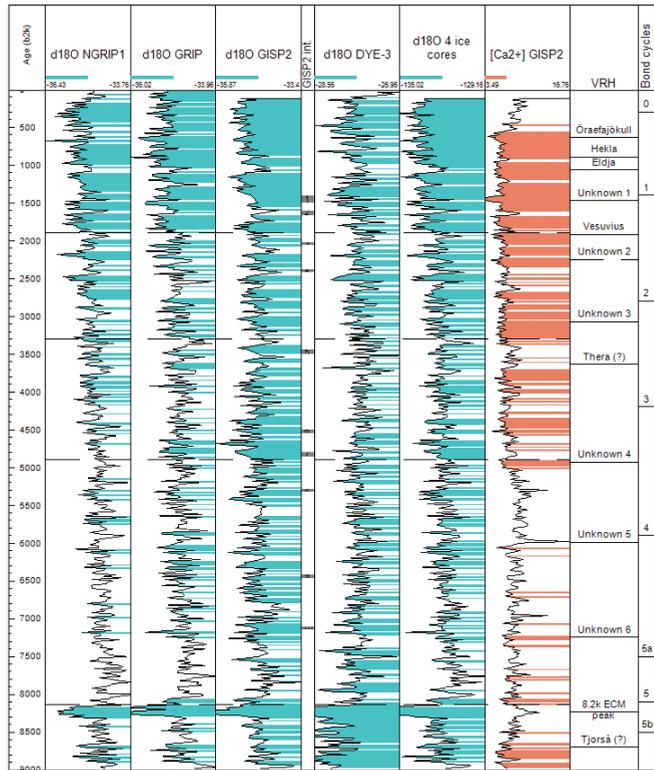
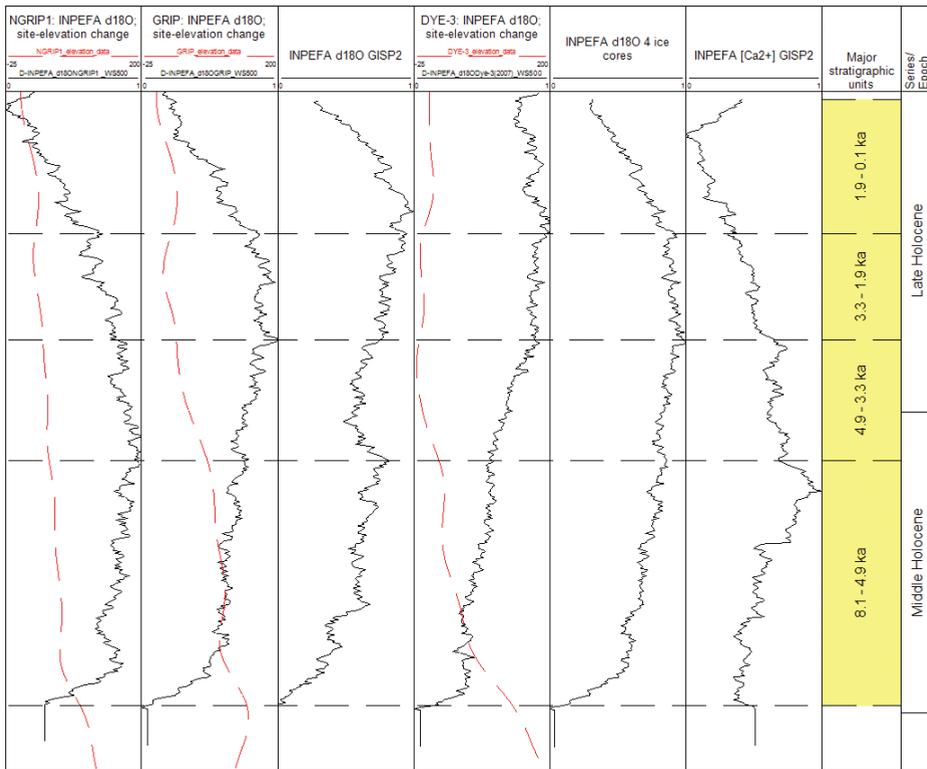


Fig. 2. Time plot, showing  $\delta^{18}\text{O}$  profile for the NGRIP1, GRIP, GISP2 and DYE-3 ice cores and the compound profile for the period 9–0 ka b2k.

Key to vertical tracks: 1 – GICC05 time scale (Rasmussen *et al.* 2006; Vinther *et al.* 2006); 2, 3, 4, 6, 7 –  $\delta^{18}\text{O}$  (per mille), horizontal scale defined by the minimum and maximum values in the interval 8–0 ka ( $\delta^{18}\text{O}$  NGRIP1: -36.43 to -33.76;  $\delta^{18}\text{O}$  GRIP: -36.02 to -33.96;  $\delta^{18}\text{O}$  GISP2: -35.87 to -33.4;  $\delta^{18}\text{O}$  DYE-3: -28.56 to -26.95;  $\delta^{18}\text{O}$  4 ice cores: -135.02 to -129.16); bi-coloring to highlight relatively high and low values (color boundary at mid-point of horizontal scale range); 5 – intervals with interpolated data in GISP2  $\delta^{18}\text{O}$  curve; 8 – GISP2  $[\text{Ca}^{2+}]$  (ppb), horizontal scale

ice cores, however, shows patterns with similarities and differences between the ice cores, as suggested by the distribution of the colors in tracks 2, 3, 4 and 6 of Fig. 2. For instance, the NGRIP1 profile shows a predominance of relatively high  $\delta^{18}\text{O}$  values (white color) in the interval from 8.1 ka to approximately 4.8–5.0 ka, followed by a predominance of relatively low values (blue color) through to 0 ka, notably after 1.9 ka. A similar change is at about 3.3–3.4 ka in the DYE-3 profile and in the GRIP profile, though slightly less pronounced in the latter. The change to relatively low values after 1.9 ka is very distinct in the GRIP profile. A corresponding change is present in the GISP2 profile, but less pronounced, probably due to the effect of data interpolation over two



defined by the minimum (3.49) and maximum (16.76) values in the interval 8–0 ka; bi-coloring to highlight relatively high and low values (color boundary at 1/4th of horizontal scale range); 9 – VRH: volcanic reference horizons after Vinther *et al.* (2006); 10 – Bond cycles (Bond *et al.* 2001); 11, 12, 13, 14, 15 – spectral trend curve (INPEFA) of  $\delta^{18}\text{O}$  data (solid black line; scale dimensionless from 0 to 1) and elevation-change curve (dashed red line; scale -25 to 200 m; from Vinther *et al.* (2009)); 16 – spectral trend curve of  $[\text{Ca}^{2+}]$  for GISP2 (scale dimensionless from 0 to 1); 17 – major stratigraphic units (this paper; unit boundaries shown as dashed horizontal lines); 18 – formal subdivision of the Holocene proposed by Walker *et al.* (2012). See text for further explanation.

intervals (track 5 in Fig. 2). These and other similarities and differences are visualized in the patterns of the  $\delta^{18}\text{O}$  spectral trend curves (tracks 11–14 in Fig. 2); long-term and (superimposed) short-term trends and trend changes in the data, which are present but less easy to identify in the corresponding untreated profiles (tracks 2, 3, 4 and 6 in Fig. 2), are revealed.

To facilitate the analysis of the individual profiles and their relevance in the regional context, a compound GIS  $\delta^{18}\text{O}$  profile was generated through stacking of the data of the four ice cores; a corresponding spectral trend curve was also generated (tracks 7 and 15 of Fig. 2). The change at 1.9 ka is now pronounced: relatively low  $\delta^{18}\text{O}$  values (blue color) predominate in the period 1.9–0 ka and

they are overall numerically lower than in the preceding period. As expected, the  $\delta^{18}\text{O}$  spectral trend curve of the stacked data (track 15 in Fig. 2) also shows a well-defined long-term pattern change at 1.9 ka: the curve turns to the left in an upward direction (up-to-the-left). Looking at the data of the individual ice cores, one may conclude that the central Greenland ice cores all contribute in a similar manner to the regional profile: a predominance of low values which are overall numerically lower than before. The contribution of DYE-3 to the regional trend change is limited.

The change at approximately 3.3–3.4 ka in the DYE-3 and GRIP profiles is reproduced in the stacked profile. The spectral trend curve shows a similar up-to-the-left long-term pattern change as at 1.9 ka, though less pronounced. A very subtle up-to-the-left long-term trend change can be observed in the GISP2 spectral curve at about 3.3 ka; a corresponding/synchronous change is visible in the NGRIP1 profile, but higher order features somewhat mask the long-term pattern change. It can be concluded that all individual profiles contribute in an overall similar manner to the long-term trend change observed in the compound curve at about 3.3 ka.

While the  $\delta^{18}\text{O}$  curve of NGRIP1 and the corresponding spectral trend curve show a long-term pattern change at about 4.9 ka, the compound  $\delta^{18}\text{O}$  curve and its spectral trend curve do not. A short-term trend change only occurs. This is not a surprise: trend changes occur in the GRIP2, GISP and DYE-3 profiles, as is highlighted in the spectral trend curves, but their nature is different from that in NGRIP1. This means that, although there is a regionally relevant break at about 4.9 ka, the long-term trend reversal change as seen at NGRIP1 is not corroborated by similar, long-term, changes in the other profiles, indicating spatial variations of  $\delta^{18}\text{O}$  over the period 4.9–3.3 ka. The interval between the 8.2 ka cold event and 4.9 ka shows in all profiles a predominance of relatively high  $\delta^{18}\text{O}$  values, indicative of limited spatial variation. Remarkably, the  $[\text{Ca}^{2+}]$  profile of GISP2 shows a subtle change at about 5 ka: the values of the period after 5 ka are overall lower than before. A small but lasting change in atmospheric circulation took place at approximately 5 ka. It is concluded that there is a subtle but meaningful  $\delta^{18}\text{O}$ -stratigraphic boundary at approximately 4.9 ka.

The stratigraphic analysis of the  $\delta^{18}\text{O}$  curves of the four GIS ice cores, with secondary input from the GISP2  $[\text{Ca}^{2+}]$  profile, results in a subdivision of the period 8–0 ka into four layers on a millennial time scale. The nature of the boundaries and units is summarized in Table 1. Elaborating on the work by Vinther *et al.* (2009), the climatological interpretation in very general terms may be as follows. The period 8.1–4.9 ka is a relatively warm period in the GIS entire area – the HCO. At approximately 4.9 ka a period starts with spatial variations in climate, cooler spells occurred in places – the end of the HCO from a regional point of view. From 3.3 ka onwards, cooler spells become important, with regional variations, and after 1.9 ka relative cool periods dominate throughout the GIS area.

Table 1

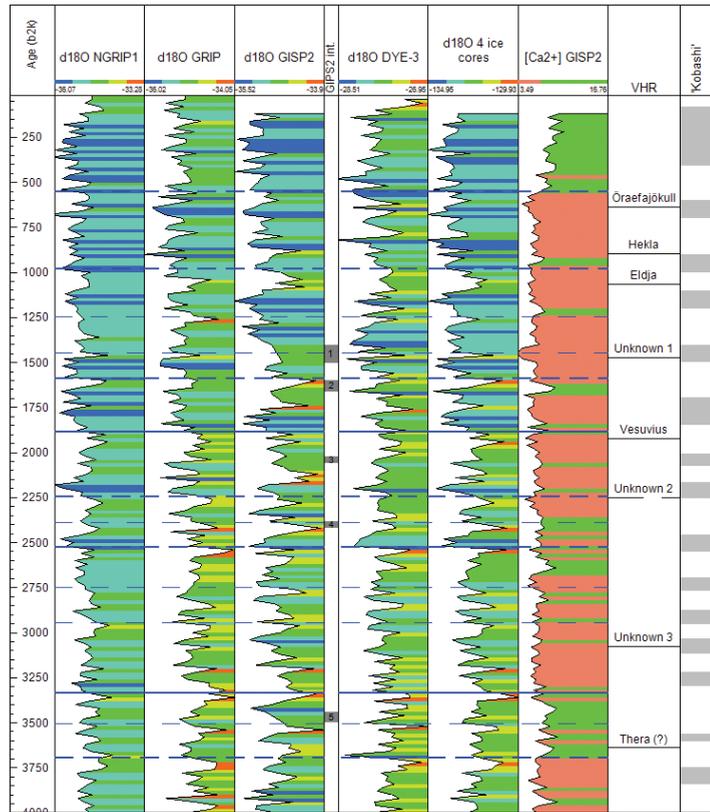
Millennial-scale stratigraphic boundaries (top) and major units (bottom) in GIS  $\delta^{18}\text{O}$  profiles. See text for explanation.

	NGRIP1	GRIP	GISP2		DYE-3
Boundary (ka b2k)	$\delta^{18}\text{O}$	$\delta^{18}\text{O}$	$\delta^{18}\text{O}$	[Ca <sup>2+</sup> ]	$\delta^{18}\text{O}$
±1.9	long-term up-to-left, distinct	long-term up-to-left, distinct	long-term up-to-left, distinct	short-term	short-term
±3.3	short-term	long-term up-to-left, distinct	long-term up-to-left, indistinct	±3.4: long-term, indistinct	long-term up-to-left, distinct
±4.9	long-term up-to-left, distinct	short-term	short-term	±5.0: long-term, distinct	long-term up-to-right, distinct
8.15	distinct	distinct	distinct	8.2: indistinct	distinct

Layer (b2k)	GIS $\delta^{18}\text{O}$ profiles	spatial variation	GISP2 [Ca <sup>2+</sup> ] profile
0.1–1.9	homogeneous	limited	low values
1.9–3.3	heterogeneous	some	
3.3–4.9	very heterogeneous	large	
4.9–8.15	homogeneous	limited	high values

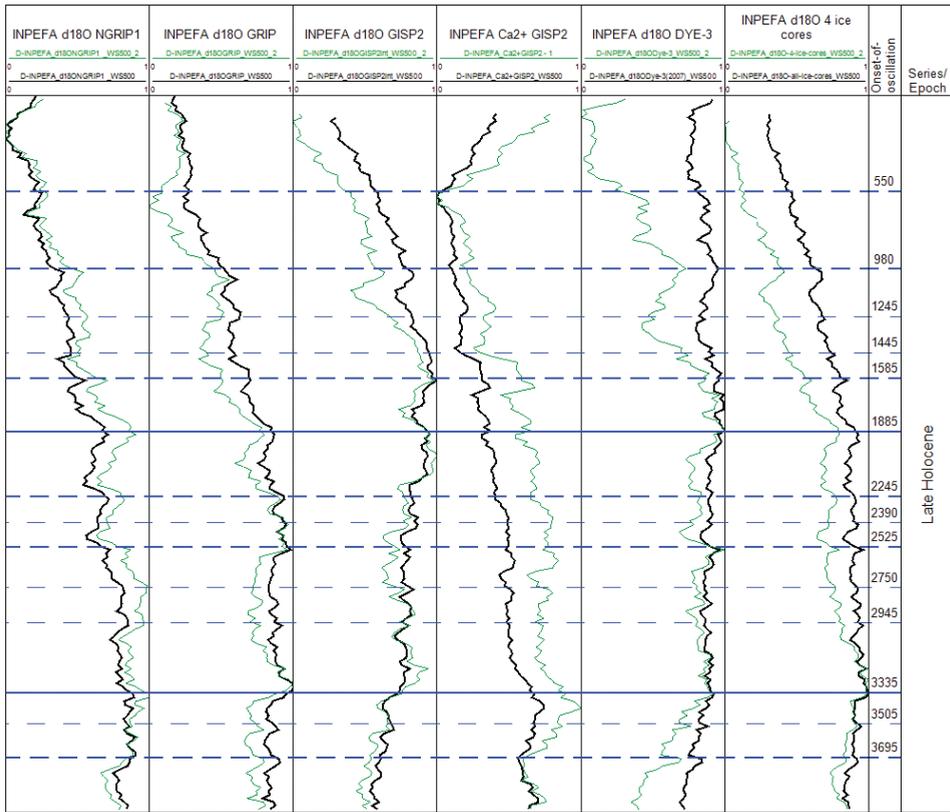
**Minor units.** — Now that we have analyzed the long-term, millennial-scale, patterns in the profiles, we will focus on the centennial-scale level, *i.e.* on the patterns superimposed on the long-term trends. Figure 3 (top) shows a correlation of the four GIS ice cores and the compound curve for the 4.0–0 ka interval. A similar correlation over the interval 8.5–4.0 ka is displayed in Fig. 3 (bottom). In addition to the spectral trend curves for the total study interval (thick lines in Fig. 3), we have calculated curves for the shorter intervals (thin lines in Fig. 3). Typically, the patterns in the data are more pronounced in the latter curves than in the former. Though not identical, the patterns of the  $\delta^{18}\text{O}$  profiles, and the corresponding spectral trend curves, show many similarities. In addition to the 1.9, 3.3 and 4.9 ka boundaries, a number of correlatable events of change can be identified and a stacking of units, bounded by these

Fig. 3 (top)



correlatable events, which reproduces between the ice cores. An oscillatory pattern of layers is present with relatively low  $\delta^{18}\text{O}$  values alternating with layers with relatively high values, the latter often thin and ‘spiky’. We have connected by horizontal lines in Fig. 3 distinct correlatable decreases in  $\delta^{18}\text{O}$  values (in an upward direction), corresponding with correlatable up-to-the left (*i.e.* in an upward direction) turning-points in the spectral trend curves. Each of these changes or breaks we call an ‘onset-of-oscillation’. Distinct increases of  $\delta^{18}\text{O}$  values, often punctuated and/or corresponding with up-to-the-right turning-points in the spectral trend curves, also occur. We have not indicated these ‘intra-oscillation-events’ in the figure, for reasons of legibility. Note that the occurrence of a correlatable intra-oscillation-event may facilitate the identification of the (overlying) correlatable onset-of-oscillation. The nature of the breaks (turning-points), however, is often subtle and the fit of the patterns not perfect in many cases, as illustrated by the following examples.

The onset-of-oscillation-event at 2.245 ka in NGRIP1 overlies a thin layer with relatively high  $\delta^{18}\text{O}$  values. A correlatable event in GRIP, though slightly less



pronounced, is readily identifiable in the  $\delta^{18}\text{O}$  curve. The corresponding turning-points in the spectral trend curves are distinct, even though the overall patterns differ between the ice cores. The event is also present in the GISP2 profile. Remarkable are the relatively high  $\delta^{18}\text{O}$  values at approximately 2.100–2.150 ka in the latter, which have no counterparts in the nearby GRIP profile and in NGRIP1; note also the effect of these high values on the pattern of the spectral trend curves. An onset-of-oscillation is present in DYE-3 at a slightly younger age than 2.245 ka. The patterns of  $\delta^{18}\text{O}$  profile and of the spectral curves around this onset-of-oscillation bear many similarities with those of GRIP and NGRIP1, strongly suggesting that these onsets-of-oscillation are the same. The stacked curve shows a subtle but distinct pattern change in the mid-term trends. We conclude that a high-order regionally relevant break occurs at 2.245 ka. Whether the differences in  $\delta^{18}\text{O}$  profiles around this age reflect spatial variations in climate (change) or are to be ascribed largely to factors of error or uncertainty remains unresolved.

Fig. 3 (bottom)

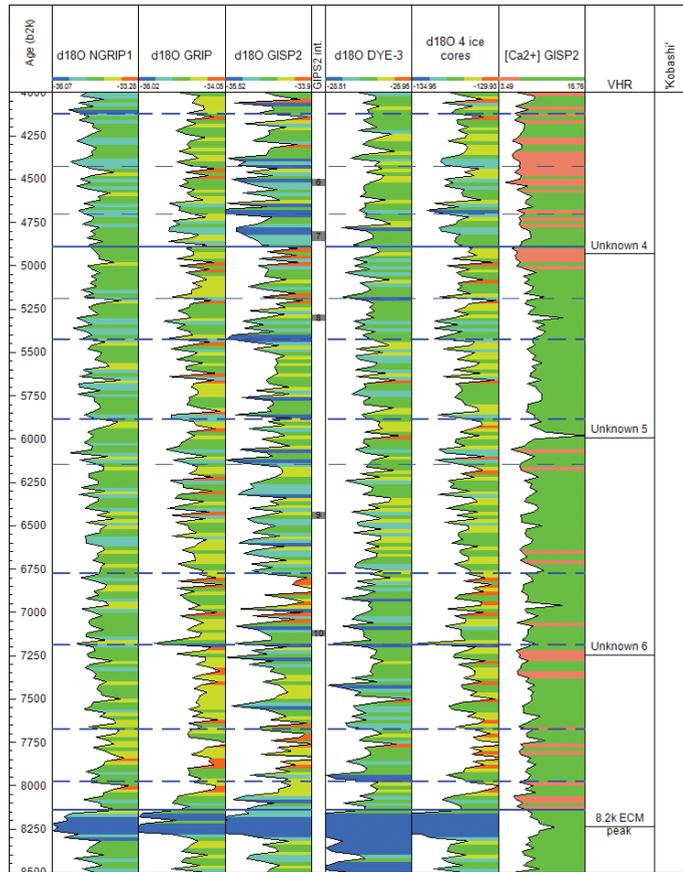
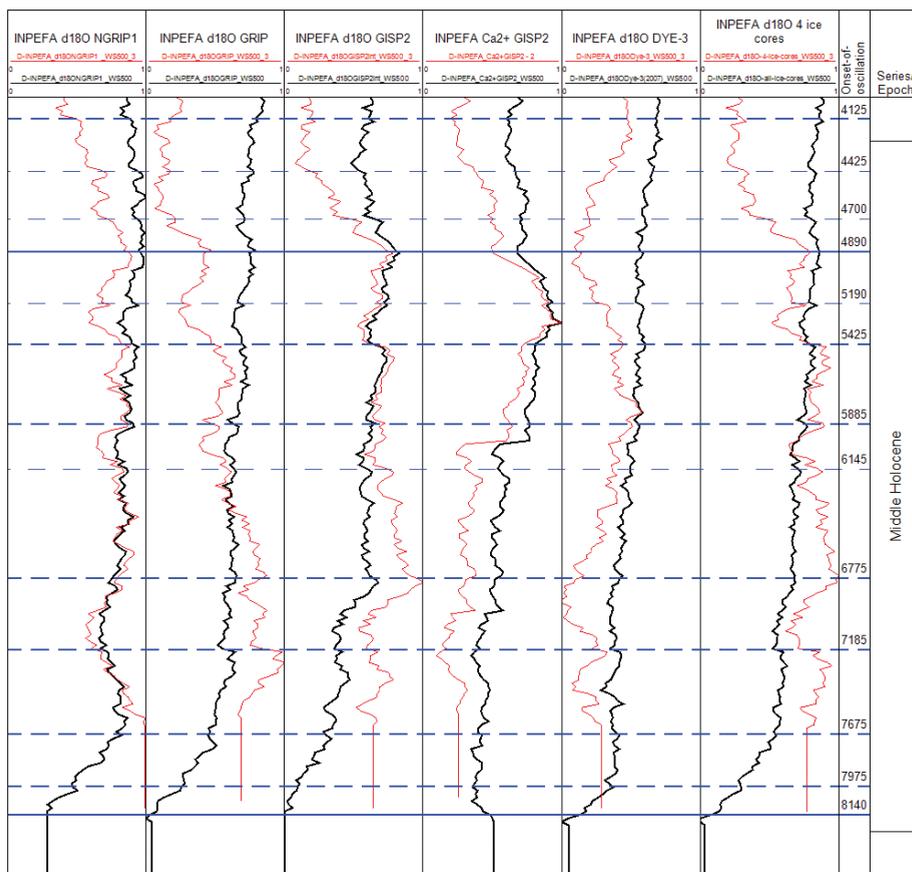


Fig. 3. Time plot, showing  $\delta^{18}\text{O}$  profile for the NGRIP1, GRIP, GISP2 and DYE-3 ice cores and the compound profile for the period 4.0–0 ka b2k; (bottom) same plot for the period 8.5–4.0 ka b2k. Key to vertical tracks: 1 – GICC05 time scale (Rasmussen *et al.* 2006; Vinther *et al.* 2006); 2, 3, 4, 6, 7 –  $\delta^{18}\text{O}$  (per mille), with coloring to highlight variability (scale ranges:  $\delta^{18}\text{O}$  NGRIP1: -36.07 to -33.28;  $\delta^{18}\text{O}$  GRIP: -36.02 to -34.05;  $\delta^{18}\text{O}$  GISP2: -35.52 to -33.9;  $\delta^{18}\text{O}$  DYE-3: -28.51 to -26.95;  $\delta^{18}\text{O}$  4 ice cores: -134.95 to -129.93); 5 – intervals with interpolated data in GISP2  $\delta^{18}\text{O}$  curve; 8 – GISP2  $[\text{Ca}^{2+}]$  (ppb; as in Fig. 2); 9 – VRH (as in Fig. 2); 10 – cool periods

The onset-of-oscillation-event at 2.525 ka is marked in the  $\delta^{18}\text{O}$  profiles of DYE-3 and NGRIP1 by a distinct decrease in values in an upward direction. The spectral trend curves show well-defined pattern changes, notably that of DYE-3. The corresponding change in GRIP is unmistakable in the spectral trend curves. Most obvious feature of the  $\delta^{18}\text{O}$  profile around this age in GRIP is the high value spike. A closer look shows a subtle overall increase in values below the spike and an overall decrease above it, *i.e.* the pattern revealed by the spectral trend curves. The  $\delta^{18}\text{O}$  profile of GISP2 around this age shows similarities with



after Kobashi *et al.* (2011: fig. 3); 11, 12, 13, 15, 16 – spectral trend curves of  $\delta^{18}\text{O}$  data (scale dimensionless from 0 to 1; thick curves same as in Fig. 2, thin green and red curves generated for top and bottom intervals, respectively); 14 – spectral trend curves of  $[\text{Ca}^{2+}]$  (thick curve for total study interval; thin green and red curves generated for top and bottom intervals, respectively); 17 – onsets-of-oscillation (this paper; solid lines – major unit boundaries, bold dashed lines – distinct boundaries, dashed lines – fairly distinct boundaries); 18 – formal subdivision of the Holocene proposed by Walker *et al.* (2012).

See text for further explanation.

the other profiles on a very detailed scale; the spectral trend curves show an overall trend change, not a pronounced turning-point. The onset-of-oscillation is well-developed in the stacked  $\delta^{18}\text{O}$  curve and the pattern change in the corresponding spectral trend curves is clear. Summarizing, we propose a high-order regionally relevant break at 2.525 ka. As for the break at 2.245 ka, the cause of the differences in the  $\delta^{18}\text{O}$  profiles around this age remains unclear.

An onset-of-oscillation occurs at 3.700 ka in the stacked  $\delta^{18}\text{O}$  curve, with a pattern change in the corresponding spectral trend curves. Looking at the input

curves, the breaks and patterns in the GRIP and DYE-3 ice cores are quite similar to those of the stacked data. Remarkably, the thin layer with high  $\delta^{18}\text{O}$  values which stands out in GRIP and DYE-3, has a complex character, with variations on the finest scale, in NGRIP1 and GISP2. In spite of this lateral variability, we propose the onset-of-oscillation at 3.700 ka to be regionally important. Once more, the cause of the lateral variability remains unclear.

Applying the correlation and interpretation ‘strategies’ outlined above we propose a minimum of 27 correlatable onsets-of-oscillation for the study interval (Fig. 3), including the 1.9 (1.885) ka, 3.3 (3.335) ka and 4.9 (4.890) ka boundaries. The interpretation is based on the integrated analysis of the four individual ice-core profiles and the stacked profile. To account for the fine-scale variability between the individual curves, the ages of the onsets-of-oscillation are determined in the compound curve. Note that there is a reasonable match between the  $\delta^{18}\text{O}$  onsets-of-oscillation and GISP2  $[\text{Ca}^{2+}]$  pattern changes.

## Renland and Agassiz ice cores

We have selected the combined Agassiz 84/87 profile from the available Agassiz ice-core data for our analysis, given that this is uplift-corrected (Vinther *et al.* 2009). The Renland data are also uplift-corrected (Vinther *et al.* 2009). Figures 4 and 5 illustrate our tentative interpretation of the Agassiz 84/87 and Renland ice cores, respectively, in terms of the onset-of-oscillation events defined in the GIS ice cores. The initial match, using the Vinther *et al.* (2009) chronology for Renland and Agassiz, leaves room for improvement. By making relatively small adjustments, with the possible errors and uncertainties discussed above in mind, a better fit with the onsets-of-oscillation in the GIS ice cores can be achieved. In making these adjustments, we are working on the assumption that we are correct in deducing a regional element in the pattern of climate change between 8 and 0 ka; that is to say that the pattern of change in one part of the larger Greenland area will resemble that of another.

The next step is to check if the major units proposed for the GIS ice cores are present in the Renland and Agassiz 84/87 profiles. The most pronounced long-term trend change in the Renland profile is at approximately 3.7 ka (see track 5 in Fig. 5), *i.e.* 400 years earlier than the major unit boundary of 3.3 ka in the GIS cores, which is similar in nature. The matching uncertainty is estimated to be about 100 yrs or less for this interval. This would mean that either the long-term cooling trend in the Renland area starts several centuries earlier than in the GIS area or the 3.3 ka trend change is masked by higher-order events. We prefer the latter explanation, assuming synchronicity of climate changes on a regional scale. Regarding the major unit boundaries of 4.9 and 1.9 ka in the

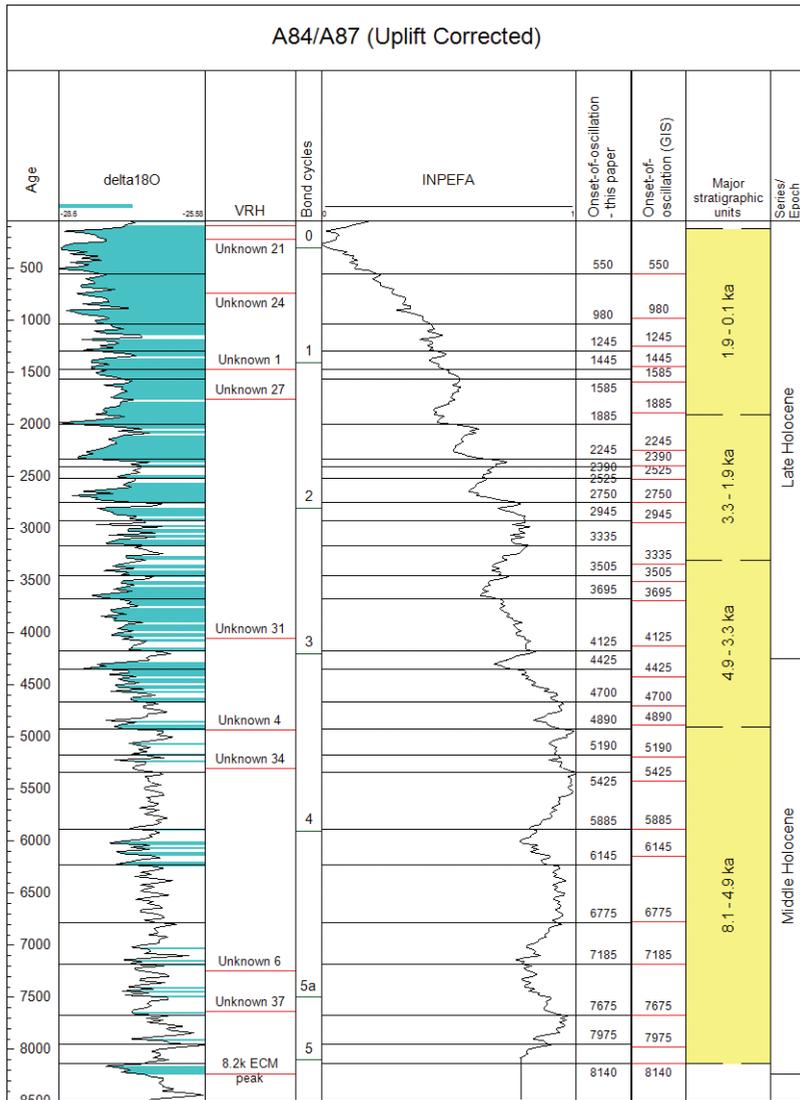


Fig. 4. Time plot, showing  $\delta^{18}\text{O}$  profile for the Agassiz 84/87 ice core for the period 8.5–0 ka b2k. Key to vertical tracks: 1 – GICC05 time scale (Rasmussen *et al.* 2006; Vinther *et al.* 2006); 2 –  $\delta^{18}\text{O}$  (per mille), horizontal scale defined by the minimum (-28.6) and maximum (-25.58) values in the interval 8–0 ka; bi-coloring to highlight relatively high and low values (color boundary at mid-point of horizontal scale range); 3 – VRH: volcanic reference horizons after Vinther *et al.* (2006); 4 – Bond cycles (Bond *et al.* 2001); 5 – spectral trend curve (INPEFA) of  $\delta^{18}\text{O}$  data (scale dimensionless from 0 to 1). Tracks 7 and 8 show the position (ages) of the onsets-of-oscillation and major stratigraphic units identified in the GIS ice cores (Figs 2 and 3), respectively. Track 6 shows these onsets-of-oscillation at the position (age) tentatively proposed in this paper. Track 9 – formal subdivision of the Holocene proposed by Walker *et al.* (2012). See text for further explanation.

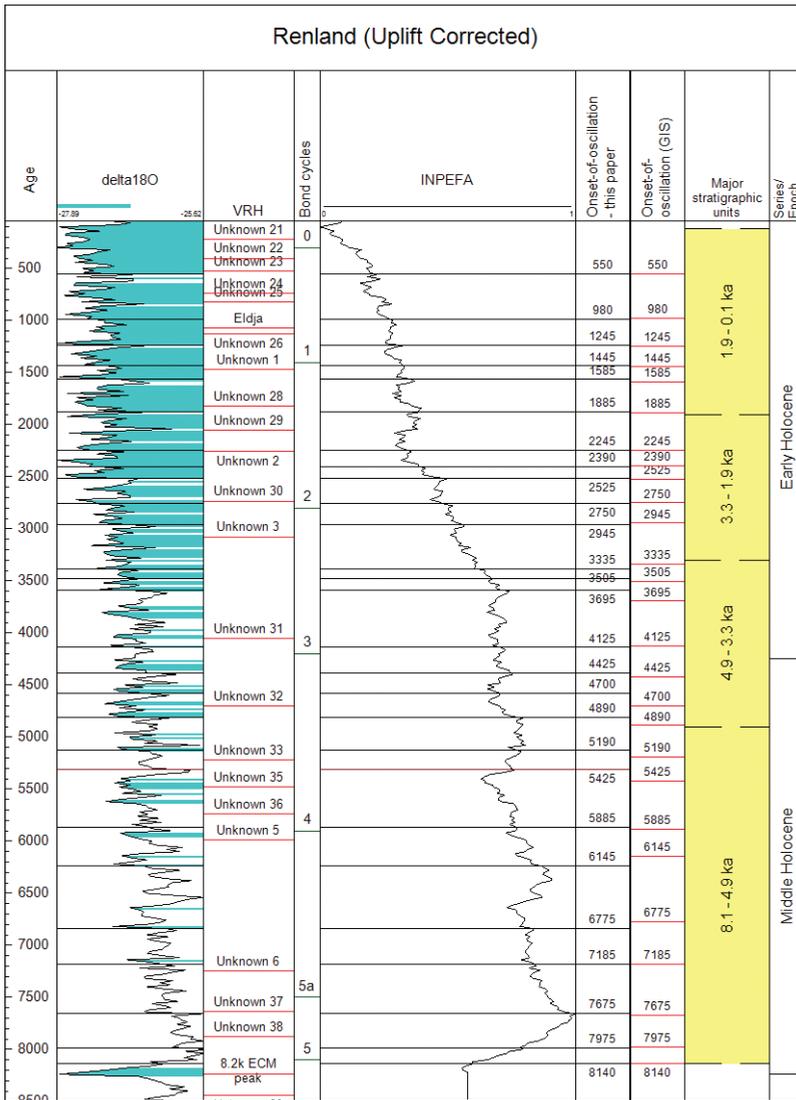


Fig. 5. Time plot for Renland ice core, as Fig. 4. Horizontal scale  $\delta^{18}\text{O}$  in track 2 from -27.89 to -25.62 (per mille).

GIS ice cores, the pattern changes in the Renland spectral trend curve at about these times are considered supportive evidence. The matching uncertainties for the Agassiz 84/87 ice cores are substantially larger than for the Renland ice core, making an interpretation more speculative. Remarkably, the Agassiz 84/87 data shows a well-developed long-term trend change at approximately 4.9 ka. The past two millennia are dominated by relatively low  $\delta^{18}\text{O}$  values. A pattern change occurs around 3.3 ka.

## Concluding remarks

The combined analysis of the datasets of the individual GIS ice cores and the stacked  $\delta^{18}\text{O}$  profile reveals a layering for the 8–0 ka period consisting of four major units (8.1–4.9 ka, 4.9–3.3 ka, 3.3–1.9 ka, 1.9–0.1 ka) and a large number of minor units with durations ranging from 135 to 630 years. The succession of major units helps us to better understand the general change towards a cooler climate after the HCO over Greenland, while the minor units are thought to represent short climatic oscillations superimposed on the main pattern.

We observe lateral and vertical variability in the profiles, which we believe to be real and to result from lateral and vertical variability in the factors that control the  $\delta^{18}\text{O}$  values of the ice, *i.e.* climate, rather than factors of error and uncertainty in the data. The unit boundaries are correlatable and, hence, near-synchronous, suggesting that climate changes in the area were near-synchronous. The driving mechanism or combination of driving mechanisms of these changes, however, are subject of ongoing debate.

$\delta^{15}\text{N}$  data of  $\text{N}_2$  are available for GISP2 on a gas-age time scale with a sample increment ranging from 2 to 92 years over the study interval (Seierstad *et al.* 2014). Comparison of  $\delta^{18}\text{O}$  and  $\delta^{15}\text{N}$  data series, hence, is not straightforward and we, therefore, did not include the latter in our analysis. Track 10 of Fig. 3 (top) shows the cold periods over the past 4000 years in the Greenland temperature curve generated by Kobashi *et al.* (2011: fig. 3) from argon ( $\delta^{40}\text{Ar}$ ) and nitrogen ( $\delta^{15}\text{N}$ ) isotopic ratios in the GISP2 ice core. A visual comparison with our units shows a poor match. The statistical correlation between the temperature and the  $\delta^{18}\text{O}$  of GRIP, GISP2 and NGRIP is low, indicating that  $\delta^{18}\text{O}$  is sensitive to more factors than temperature only, *e.g.* seasonal storm activity and snowfall obscuring temperature signals (Kobashi *et al.* 2011).

Bond *et al.* (1999, 2001) presented evidence for a “series of shifts in ocean surface hydrography during which drift ice and cooler surface waters in the Nordic and Labrador Seas were repeatedly advected southward and eastward, each time penetrating deep into the warmer strands of the subpolar circulation” during the Holocene, with a cyclicity of approximately  $1470 \pm 500$  years. The peaks of these ice-rafted debris (IRD) events or Bond cycles in the Holocene are at about 1.4, 2.8, 4.2, 5.9, 8.1, 9.4, 10.3 and 11.1 ka years ago, the youngest event corresponding with the Little Ice Age. The 1470 years periodicity has become a subject of debate and did not unequivocally pass rigorous statistical testing by Ditlevsen *et al.* (2007; see also Vaughan *et al.* 2011). The difference in time scales, decadal to centennial versus millennial, prohibits a meaningful comparison of our results with the IRD events (Fig. 2).

Based on the statistical evaluation of a large number of climatic datasets and proxies, including the GRIP, NGRIP and Agassiz ice cores, on a decadal to centennial time scale, Wanner *et al.* (2008, 2011) identified six cold events

in the Holocene: at about 8.2, 6.3, 4.7, 2.7, 1.55 and 0.55 ka BP. They stated that a clear cyclicity was not found, and that the spatiotemporal variability of temperature and humidity/precipitation during these events was high. The 4.7, 2.7, 1.55 and 0.55 cold relapses may correspond with events identified in the 8–0 ka period of this paper; we refer to Blaauw (2012) for a discussion of the risks involved in forcing matches between time series. The definition of cold and warm periods by Wanner *et al.* (2011), as time spans where the proxy values fall outside one half of a standard deviation of the mean value of the data series, is statistical, whereas we look at patterns and pattern changes in a semi-quantitative way. A relatively weak signal of climate change at a certain location can be meaningful in our approach, if it correlates with signals (weak or strong) elsewhere. This is considered the primary reason for the larger number of oscillations revealed by our method. Weak signals may well have been missed in the study by Wanner *et al.* (2011).

Walker *et al.* (2012) argue in a discussion paper that the informal threefold subdivision of the Holocene into early, middle (or mid) and late should be formalized to overcome the problems associated with the inconsistent usage of the terms. The 8.2 event is proposed by these authors as the boundary between the Early and Middle Holocene. The GSSP (Global Stratotype Section and Point) has been located in the NGRIP1 ice core at a depth of 1228.67 m, with an age of 8.236 ka b2k on the GICC05 timescale. Registered most strongly in localities around the North Atlantic Ocean, the 8.2 event is found in proxy records from all over the world. The boundary between the Middle and Late Holocene is placed by Walker *et al.* (2012) at the so-called 4.2 event, which is a mid/low-latitude aridification event. It is defined in the Mawmluh Cave speleothem stable isotopic record at an age of 4200 cal a BP (equivalent to 4.250 ka b2k). Our study interval comprises the period from 8.140 to 0.120 ka b2k and falls within the Middle and Late Holocene of the proposed scheme by Walker *et al.* (2012). Our minor unit starting with the onset-of-oscillation at 4.425 ka b2k may be the equivalent of the 4.2 event. It is a minor event in our stratigraphic scheme, which is in agreement with the report by Walker *et al.* (2012) that the 4.2 event is pronounced (only) in mid/low-latitude records.

On a general note, we would like to emphasize that there is no fundamental disagreement between the subdivision proposed by Walker *et al.* (2012) and our subdivision of the Greenland ice cores. Walker *et al.* (2012) propose a formal chronostratigraphic subdivision of the Holocene, using two well-defined events as the boundaries in a tripartite scheme. Our study analyzes trend and trend changes in ice-core data series for a time interval that falls within the Middle and Late Holocene of the scheme of Walker *et al.* (2012). The objectives are different, the results are complementary.

As commonly accepted for the older parts of the GIS ice cores, we suggest that the pattern revealed by our multi-profile stratigraphic analysis of the period

8–0 ka represents climate changes that affected a larger area than Greenland and its immediate surroundings. Evidence in support comes from the analyses of the Renland and Agassiz A84/87 ice cores. Comparing long-term high-resolution climate proxies from other areas, *e.g.* ocean-bottom sediment cores; dendrochronological records, with the subdivision proposed in this study will lead to further insight and understanding of Northern Hemisphere climate change in the recent past. In our opinion, the focus of such comparisons should be as much on signals and patterns of relative climate change as on the statistical analysis of climate proxy data.

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