# $^{14}$C and $^{10}$Be around 1650 cal BC:

## are there contradictions between tree ring and ice core time scales?

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There is a debate about the date of the Minoan eruption of Santorini as reconstructed from a branch of an olive tree that was buried alive in trephra on Santorini[2] and the dating of an ice core horizon attributed to this eruption.[3] The olive branch was $^{14}$C-wiggle matched to the $^{14}$C calibration curve and yielded an age range of 1627-1600 BC[4] while the counting of annual layers in the Greenland ice cores produced an age of 1642 ($\pm$5) BC.[5] Here I will study the relative timing of the two time scales by comparing the cosmic ray signal as recorded by $^{14}$C in tree rings and $^{10}$Be in ice cores. The result reveals an intriguing age difference that is similar to the dating difference mentioned above. The origin of the difference is unclear. This analysis supports both the dating of the Santorini eruption with the olive branch and the identification of this eruption in the ice cores, but it suggests unrecognised uncertainties in the tree ring $^{14}$C data, the ice core chronology or both.

## Introduction

There are very different methods to date the Minoan eruption of Santorini and two of the methods applied by natural scientists seem to lead to conflicting results. One is based on the identification of the fallout of the Santorini eruption in Greenland ice cores and the other method is based on $^{14}$C dating of plant remains on Santorini. In the following, I will concentrate on the comparison of the tree ring chronology and the new Greenland ice core time scale. These two time scales underlie the dating of the olive tree[6] and the age determination of the ice core layer that has been attributed to this eruption.[7] The comparison can be done via cosmogenic radionuclides in tree rings and ice cores. Cosmogenic radionuclides are particles that are produced in the Earth's atmosphere by the interaction of galactic cosmic rays with atoms of the atmosphere.[8] Variations in the galactic cosmic ray flux produce a global signal in cosmogenic radionuclide records that can be used to compare different time scales. In particular, solar activity variations generate numerous time markers since they modulate galactic cosmic rays on decadal to centennial time scales (11, 88 and 207 yr solar cycles)[9] which leave a clear imprint in cosmogenic radionuclide records. In the following I will compare the $^{10}$Be record from the GRIP ice core[10] and the $^{14}$C record that underlies the $^{14}$C calibration method.[11] Both radionuclides vary similarly with changes in the cosmic ray intensity but after their production they behave absolutely differently. $^{14}$C oxidizes to $CO_2$ and enters the carbon cycle,[12] while $^{10}$Be is removed from the atmosphere within 1-2 years mainly by wet deposition.[13] This different geo-chemical behaviour has

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[2] Friedrich *et al.* 2006.

[3] Vinther *et al.* 2006.

[4] 2$\sigma$ error, Friedrich *et al.* 2006.

[5] 2$\sigma$ error, Vinther *et al.* 2006.

[6] Friedrich *et al.* 2006.

[7] Vinther *et al.* 2006.

[8] Lal & Peters 1967.

[9] *E.g.* Damon & Sonett 1991.

[10] Muscheler *et al.* 2004.

[11] Reimer *et al.* 2004.

[12] Lal & Peters 1967.

[13] McHargue & Damon 1991.

to be accounted for to get an accurate comparison between $^{10}Be$ and $^{14}C$ records.

In the following, I will repeat the main arguments that support the identification of the Santorini eruption in ice cores from Greenland. This will be followed by the presentation and comparison of the radionuclide data in tree rings and ice cores. Possible reasons for the $^{10}Be$-$^{14}C$ differences will be given in the subsequent discussion and possibilities to reconcile the ice core and tree ring time scales will be discussed.

## Identification of volcanic eruptions in ice cores

Strong volcanic eruptions eject particles into the stratosphere where they can be transported around the globe. Therefore, strong eruptions can leave their imprint in polar ice cores. Especially volcanic acids can be easily detected by electrical conductivity measurements (ECM) in ice cores.[14] Identifying such volcanic signals in ice cores is very useful for the synchronization of different ice core time scales.[15] However, ECM measurements cannot give unequivocal information about the source of the signal. Acidity and ECM peaks can be caused by melt layers[16] and nearby smaller volcanic eruptions can produce stronger signals than a strong volcanic explosion further away from the ice core site. Chemical analysis of tephra can add important information for the identification of the volcanic eruption responsible for the signal. Depending on the chemical signature it is possible to reject or support certain eruptions as the source for the tephra. In addition, the relative arrival times of tephra and acidity signal can provide information about the location of the volcanic eruption leaving its imprint in Greenland ice cores.[17]

Combining data from several ice cores can add crucial information for the identification and dating of a volcanic eruption. For example, high northern latitude eruptions could produce ECM spikes in ice cores from Central and Northern Greenland but they could be invisible in ice cores from Southern Greenland.[18] Strong volcanic eruptions in mid northern latitudes or equatorial regions are expected to produce a signal that is visible all over Greenland. In fact, the proposed Santorini ECM signal is visible in three major cores in Greenland that are from Southern (DYE3), Central (GRIP) and Northern Greenland (NGRIP).[19] Each of the cores was dated individually. Therefore, this event can be identified in all three cores and related to one another. However, for the construction of the GICC05 ice core time scale only the most reliable annual signals are used for absolute dating and the ECM signals are used to transfer the time scale to all of the cores.[20]

Based on the DYE3 ice core data[21] Hammer concluded that the date of the Santorini eruption was most likely 1645 BC. They assigned a strong acidity signal in the DYE3 ice core to the Santorini eruption. Chemical analysis revealed a high level of sulphuric acid which confirmed the volcanic origin of the ECM spike. The age was determined by annual layer counting in the ice core. The dating uncertainties were assumed to be ±7 years with an estimated upper limit of ±20 yrs.[22]

Using data from two additional ice cores in Greenland Vinther *et al.* (2006) re-dated this layer to 1641 BC with an assumed maximum counting error of 5 years. This implies an age of 1642 BC of the Santorini eruption considering the delay between volcanic eruption and deposition at the ice core site.

The connection of the 1641 BC acidity and tephra layer to the Santorini eruption has been debated in the past. The latest of these criticisms was advanced by Denton and Pearce (2008). Based on the geochemical analysis of the tephra they argue that the Aniakchak volcano in Northern Alaska was the most likely source for this acidity and tephra layer. In the following I will repeat Vinther *et al.*'s (2008) main arguments that the 1641 BC layer is indeed connected to the Santorini eruption:

[14] Hammer *et al.* 1980; Hammer *et al.* 1987.
[15] Vinther *et al.* 2006.
[16] Hammer *et al.* 1987.
[17] Vinther *et al.* 2006.
[18] Clausen *et al.* 1997.
[19] Vinther *et al.* 2008.
[20] Vinther *et al.* 2006.
[21] Hammer *et al.* 1987.
[22] Hammer *et al.* 1987.

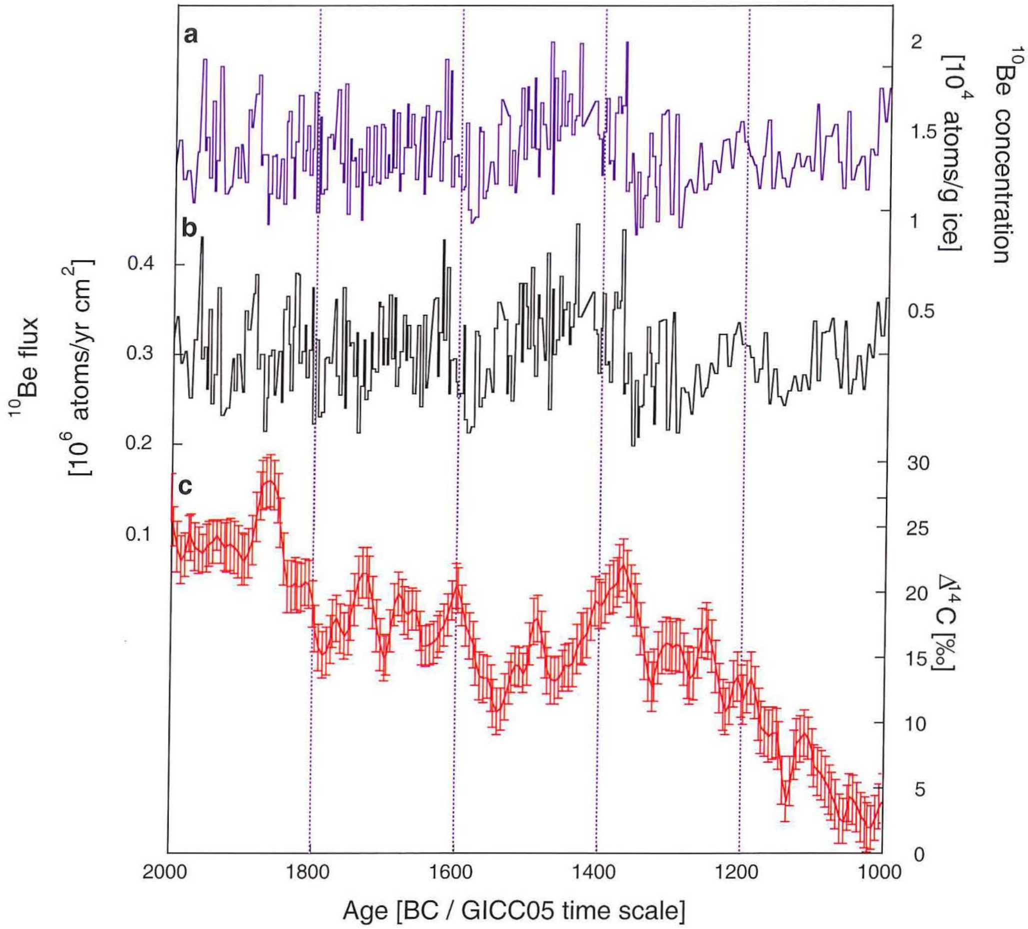



Fig. 1. $^{10}$Be and $^{14}$C for the period from 2000 to 1000 BC. Panel a shows the $^{10}$Be concentration measured in the GRIP ice core (Muscheler *et al.* 2004). The $^{10}$Be flux is shown in panel b. The ice core data is based on the newest official ice core time scale for the Greenland ice cores (Vinther *et al.*, 2006). Panel c shows the atmospheric $^{14}$C concentration in the $\Delta^{14}$C notation as derived from tree ring measurements (Reimer *et al.* 2004). The average time resolution of the GRIP 10Be data is approximately 4.5 years and the resolution of the $\Delta^{14}$C record is 5 years. However, the IntCal04 data exhibits less fine structure compared to the previous IntCal98 calibration data set (Reimer *et al.* 2004; Stuiver *et al.* 1998).

- The 1641 BC acidity layer is the only major acidity signal seen in the ice cores from Southern, Central and Northern Greenland (DYE3, GRIP and NGRIP) with an age that is close to independent age determinations based on $^{14}$C dating.[23] As mentioned, high-latitudes volcanoes do not necessarily leave an imprint in Southern Greenland ice cores.
- Including the rare element analysis Vinther *et al.* (2008) do not agree that the 1641 BC tephra is significantly different from the Santorini tephra. In addition, they see no contradiction between the findings of Ca-rich tephra shards and the assignment to the Santorini eruption.
- The sequence of events in Greenland (tephra deposited several months before the acidity peak) supports a distant highly explosive volcano where the sulphate aerosols are transported through the stratosphere. An Alaskan volcano more likely produces a synchronous sulphate and tephra signal.[24]

The $^{10}$Be-$^{14}$C comparison adds helpful information to this discussion and it supports the arguments by Vinther *et al.* (2008) as I will show in the following.

[23] Friedrich *et al.* 2006.

[24] Vinther *et al.* 2006.

## Data & geochemical behaviour of cosmogenic radionuclides

Fig. 1 shows the $^{10}$Be concentration and the $^{10}$Be flux as measured in the GRIP ice core.[25] The data are plotted versus the latest Greenland ice core time scale GICC05.[26] $^{10}$Be concentration and $^{10}$Be flux are similar during periods with relatively stable accumulation rates. However, especially during climatically stable periods it is not clear if the $^{10}$Be concentration or the $^{10}$Be flux are better representatives of the $^{10}$Be production rate and, therefore, it is unclear which record is better suited for the following timing analysis. Therefore, the calculations will be done with both records in order to evaluate if the conclusions depend on the applied record. The accumulation rate required for the calculation of the $^{10}$Be flux is deduced from the GICC05 time scale. The $^{10}$Be data can be compared to the results of $^{14}$C measurements on tree rings that are shown in Fig. 1c.[27] The $^{14}$C data is plotted in the $\Delta^{14}$C notation which depicts the variations in the atmospheric $^{14}$C concentration and can be inferred from the relationship between $^{14}$C age and calendar age.[28] The dating of the $^{14}$C record is based on dendrochronology with multiple replication and cross-checks between different chronologies.[29]

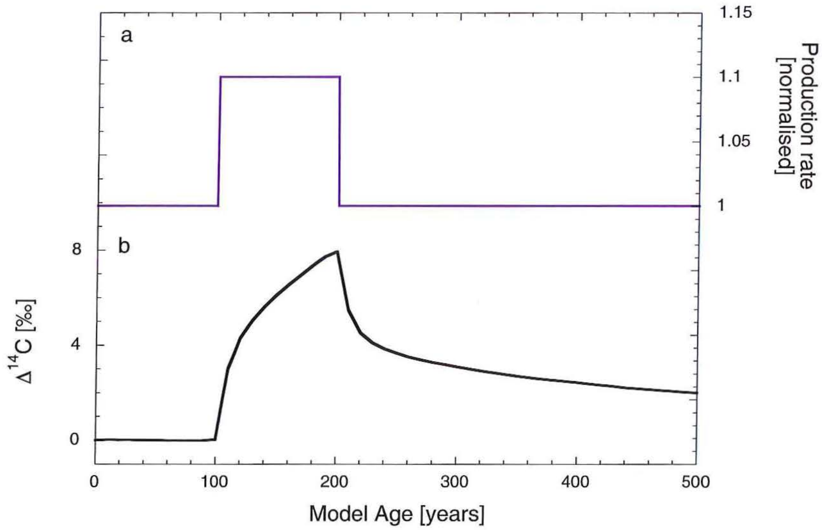
It is visible from the raw data that the $^{10}$Be record exhibits larger short-term variations compared to the $^{14}$C record. This is due to the different pathways of $^{10}$Be and $^{14}$C after their production. $^{10}$Be data show the short-term variations in the production rate due to the relatively short atmospheric residence time. For example, the solar 11-year cycle can be seen in annual $^{10}$Be records.[30] However, the $^{10}$Be deposition is also influenced by changes in weather and climate. "Weather noise" in the $^{10}$Be records contributes to the short-term scatter but it should not influence the longer-term variations. Persistent changes in climate can potentially affect the $^{10}$Be deposition for longer periods of time. By contrast, atmospheric $^{14}$C records do not show such high-resolution changes. $^{14}$C enters the carbon cycle and becomes part of a large reservoir of previously produced $^{14}$C. Annual changes in the $^{14}$C production rate are, therefore, hardly visible in the atmospheric $^{14}$C concentration. The long atmospheric residence time of $^{14}$C ensures that it is well mixed and regional differences in the $^{14}$C production, which is higher at the poles and lower at the equator, are not visible in $\Delta^{14}$C. Similarly, changes in the atmospheric circulation hardly influence the atmospheric $^{14}$C concentration. However, changes in the ocean circulation can have an impact on $\Delta^{14}$C. Increased exchange within the ocean could increase the transport of $^{14}$C-depleted carbon from the deep ocean to the upper ocean and thereby decrease $\Delta^{14}$C and vice versa for decreased ocean mixing. Such changes would likely be connected with major changes in climate. However, even major changes in ocean circulation are supposed to have only a limited influence on $\Delta^{14}$C.[31] Therefore, climate changes are a rather unlikely cause for $\Delta^{14}$C variations during the relatively stable Holocene climatic period.

Nevertheless, if $^{10}$Be and $^{14}$C records are compared quantitatively, the carbon cycle does have to be considered. Due to the large $^{14}$C reservoirs the atmospheric $^{14}$C changes are dampened and delayed compared to the changes in the $^{14}$C production rate. Fig. 2 illustrates the difference between $^{14}$C production rate and atmospheric $^{14}$C concentration changes. In particular, the delayed reaction of $\Delta^{14}$C must be taken into account if the timing between $^{10}$Be and $^{14}$C records is investigated. There are two approaches to include the influence of the carbon cycle. Assuming that the $^{10}$Be data represents the global radionuclide production rate, one can reconstruct the $^{14}$C production rate and calculate a $^{10}$Be-based atmospheric $^{14}$C concentration with a carbon cycle model. Alternatively, one can reconstruct the $^{14}$C production rate from the $\Delta^{14}$C data and compare these with the $^{10}$Be data.

$^{10}$Be measured in ice cores provides a more direct proxy record for changes in the incoming cosmic ray flux than $\Delta^{14}$C. $^{10}$Be has a mean atmospheric

[25] Muscheler *et al.* 2004.
[26] Vinther *et al.* 2006.
[27] Reimer *et al.* 2004.
[28] Stuiver & Polach 1977.
[29] Reimer *et al.* 2004 and references there.
[30] Beer *et al.* 1990.
[31] Marchal *et al.* 2001.

Fig. 2. A hypothetical change in the $^{14}C$ production rate (panel a) and the corresponding change in the atmospheric $^{14}C$ concentration (panel b). Due to the effects of the carbon cycle the $\Delta^{14}C$ changes are dampened and delayed with respect to the $^{14}C$ production rate changes.




residence time of the order of one year.[32] For ice cores from Central Greenland it has been shown that a dust-related $^{10}Be$ component is negligible during warm periods[33] and a recycled dust-borne $^{10}Be$ component can be neglected. Therefore, it is rather straightforward to correct for the delayed deposition by shifting the $^{10}Be$ data accordingly in time. However, potential weather and climate influences are hard to estimate and cannot be included in the following calculations. Nevertheless, such changes are sources of uncertainty for the $^{10}Be$ – $^{14}C$ comparison and will also be discussed.

## $^{10}Be$ – $^{14}C$ comparison

In the following I will compare the ice core $^{10}Be$ and the tree ring $^{14}C$ records after the known differences in the geochemical behaviour are corrected for. After the correction for the one-year delay in the $^{10}Be$ deposition, I assumed that the GRIP $^{10}Be$ concentration/flux reflects the globally averaged $^{10}Be$ production rate. With this assumption, the $^{14}C$ production rate can be reconstructed by using the results of theoretical production rate calculations.[34] The $^{10}Be$-based $^{14}C$ production rate was then used as input for a carbon cycle model[35] to calculate the atmospheric $^{14}C$ concentration. Common changes in the tree-ring $\Delta^{14}C$ record and the $^{10}Be$-based $\Delta^{14}C$ data can then be attributed to the similar production processes, and therefore be used to compare the different time scales. However, a perfect match between the two records cannot be expected because of the mentioned climatic influence on $^{10}Be$, and to a minor degree also on the $\Delta^{14}C$ data.

Fig. 3 shows the linearly-detrended $\Delta^{14}C$ record derived from the tree ring chronology[36] and the modelled $\Delta^{14}C$ based on $^{10}Be$ connected to the ice core time scale GICC05. It is obvious that measured and modelled $\Delta^{14}C$ records show some disagreements but there are several peaks that are most likely due to a common production origin. Especially the peaks around 1900, 1600 and 1400 BC seem suitable to study the timing between the tree ring chronology and the ice core time scale. The differences between the data (for example around 1450 and 1250 BC) indicate climate-related influences on the $^{10}Be$ and/or $^{14}C$ records. The $^{10}Be$ and $^{14}C$ data around the Santorini eruption do not exhibit dominant solar peaks such as, for example, during the early Holocene.[37] Therefore, this timing analysis is more uncertain for the period shown in Fig. 3 than during other periods. Nevertheless, Fig. 3 suggests that there is a time scale difference between the $^{10}Be$ and $^{14}C$ records. Even if there are differences between $\Delta^{14}C$ based on $^{10}Be$ flux and

[32] Raisbeck *et al.* 1981.
[33] Baumgartner *et al.* 1997.
[34] Masarik & Beer 1999.
[35] Siegenthaler 1983.
[36] Reimer *et al.* 2004.
[37] Muscheler *et al.* 2000.

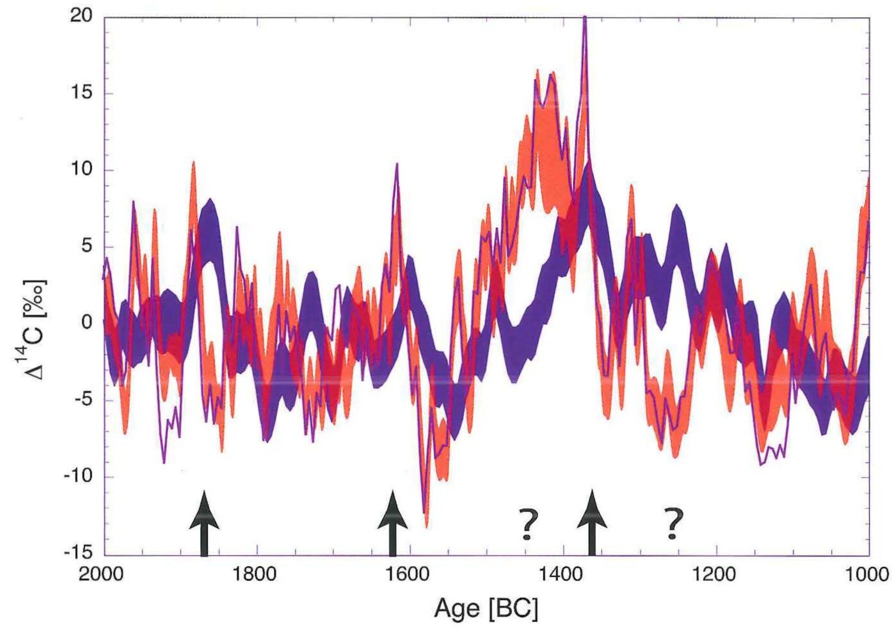



Fig. 3. Comparison of the tree ring and the $^{10}Be$-based $\Delta^{14}C$ records. The red band shows the $\Delta^{14}C$ record inferred from the $^{10}Be$ concentration measured in the GRIP ice core. The uncertainty range (1-$\sigma$ error) indicated by the band is based on the measurement uncertainties in the $^{10}Be$ data. The blue band shows the tree-ring $\Delta^{14}C$ data including its errors (1-$\sigma$ error). The purple line shows $\Delta^{14}C$ inferred from the $^{10}Be$ flux to Summit in Greenland. Although there are slight differences between $^{10}Be$ concentration and $^{10}Be$ flux, both records lead to a similar conclusion about the timing of tree ring and ice core records. All records are shown after removal of the linear trend from 2000 to 1000 BC. Arrows indicate periods/changes that seem suitable for a timing analysis. Unexplained differences (indicated by the question marks) complicate such an analysis.

$^{10}Be$ concentration (Fig. 3) these differences do not lead to significantly different conclusions about the timing between $^{10}Be$-based and tree-ring $\Delta^{14}C$ data. Fig. 4 illustrates this time shift even better. It shows the tree ring and the ice core data after the ice core time scale was shifted by 20 years, towards younger ages. This time shift of 20 years yields the highest correlation between $^{10}Be$-based and tree-ring $\Delta^{14}C$ records from 2000 to 1000 BC. Of course, one could increase the agreement between $^{10}Be$ and $^{14}C$ even more by individually adjusting all of the common $^{10}Be$ and $^{14}C$ peaks. For example, shifting the peak around 1750 BC by an additional 20 years would increase the local agreement considerably. However, such a change would imply large errors in the relative dating either in the ice core or in the tree ring time scale neither of which is very likely. In addition, such local adjustments seem to be problematic considering the overall differences between the two records.

# Discussion

It is interesting to note that this time shift of 20 years would reconcile the two methods to date the Santorini eruption as outlined above. Therefore, if either the tree ring chronology or the ice core chronology were to include errors in the order of 20 years, these results would confirm both (i) the identification of the Santorini eruption in the Greenland ice cores and (ii) the dating of the Minoan eruption of Santorini by means of the olive branch buried by the eruption. Of course, uncertainties in both time scales that add up to a difference of 20 years would lead to the same conclusion. However, both the ice core time scale and the $^{14}C$ dating method suggest smaller errors which pose questions about the result of the $^{10}Be$ – $^{14}C$ comparison. Possible solutions for the 20 year time shift and their plausibility are discussed in the following.

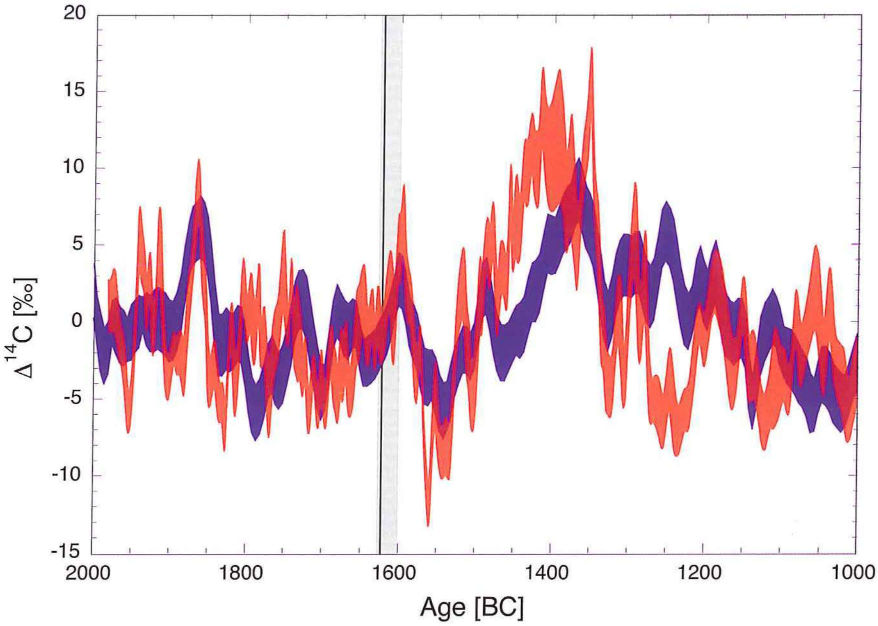



Fig. 4. Comparison of measured (blue band) and modelled $\Delta^{14}C$ (red band) after the ice core time scale was shifted by 20 years towards younger ages. Both records are linearly detrended. The modelled data is based on the $^{10}Be$ concentration measured in the GRIP ice core from Summit in Central Greenland. The age range of the $^{14}C$-dated olive tree and the tephra layer in the ice cores (shifted by 20 years) is indicated by the vertical grey band and the black line.

1) $^{10}Be$ transport uncertainties. As mentioned above $^{10}Be$ has a mean atmospheric residence time of the order of one year. It is very unlikely that this estimate is wrong by one order of magnitude.

2) Climatic impact on the $^{10}Be$ deposition and uncertainties in the $^{10}Be$ data. This is probably a likely cause for apparent shifts in the timing of the tree ring and the $^{10}Be$-based $\Delta^{14}C$ peaks. Such potential "problems" in the $^{10}Be$ data become obvious when the tree ring and $^{10}Be$-based peaks have different amplitudes or shapes. The period around the Minoan eruption of Santorini clearly exhibits such differences. Nevertheless, such differences are expected to lead to stochastic differences between $^{10}Be$ and $^{14}C$. A systematic shift in the time scales would not be expected with climate or data related uncertainties.

3) Carbon cycle uncertainties. The effect of the carbon cycle is accounted for by transferring the $^{10}Be$ data to a $^{10}Be$-based $\Delta^{14}C$ record. However, if the carbon cycle model does not represent the actual carbon cycle well during the period around the Minoan eruption of Santorini, one might obtain errors in the timing between the $^{10}Be$ and $^{14}C$ records. Fig. 5 displays calculations where different oceanic mixing rates are applied for the calculation of $\Delta^{14}C$. It shows that modelled $\Delta^{14}C$ peaks can depend on carbon cycle parameters. Therefore, such uncertainties could explain a systematic time difference between modelled and measured $^{14}C$ records. However, the inferred timing uncertainties are rather smaller than 10 years and the assumed carbon cycle differences for the two calculations shown in Fig. 5 are unrealistically large. In addition, there is no time difference between $^{10}Be$-based and measured $\Delta^{14}C$ for the period of the last thousand years. Therefore, a carbon cycle-related explanation for a 20-year time shift between $^{10}Be$ and $^{14}C$ records is rather unlikely.

4) Carbon cycle changes: A change in the carbon cycle can produce $^{14}C$ peaks that should have no corresponding peak in the $^{10}Be$ record. However, such changes cannot produce fast and strong changes in $\Delta^{14}C$. The most recent cold

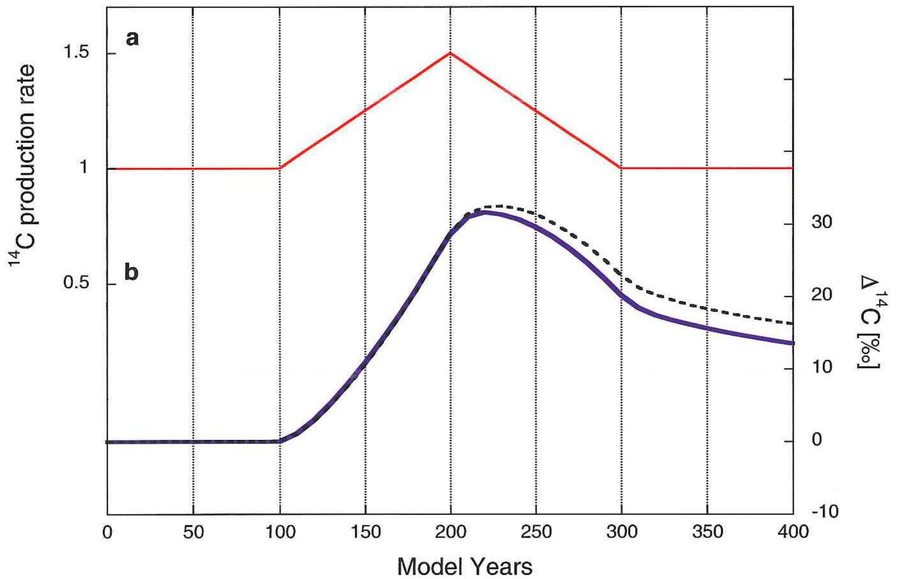



Fig. 5. An example for the sensitivity of the modelled $\Delta^{14}C$ to carbon cycle uncertainties. Panel a shows a hypothetical change in the $^{14}C$ production rate. Panel b shows the reaction of $\Delta^{14}C$ depending on the carbon cycle model. The blue curve shows the result according to the carbon cycle model by Siegenthaler (1983). The black dotted line shows the result with the same model but with a halved oceanic mixing rate.

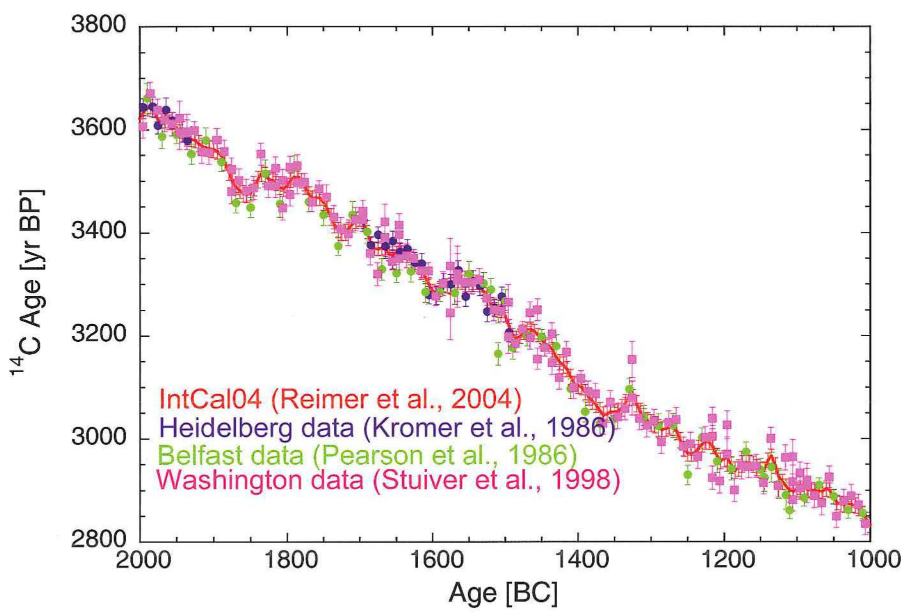



Fig. 6. $^{14}C$ calibration data from 2000 to 1000 year BC. The IntCal04 calibration record (Reimer *et al.*, 2004) is based on tree-ring data from Heidelberg (Kromer *et al.* 1986), Belfast (Pearson *et al.* 1986) and Washington (Stuiver *et al.* 1986) during the period around the Santorini eruption.

spell at the end of the last ice age is a good illustration. This dramatic cold period is most likely connected to a rearrangement of oceanic circulation. Nevertheless, model calculations suggest that the influence on $\Delta^{14}C$ was rather limited.[38] It suggests that a $\Delta^{14}C$ change of the order of 30‰ evolved during approximately one thousand years. Such a change could not explain the $\Delta^{14}C$ peaks that are important for the $^{10}Be$ – $^{14}C$ comparison shown in Fig. 3.

5) $\Delta^{14}C$ data problems: The IntCal04 calibration record is based on several $^{14}C$ records.[39] Differences between those records do exist (see Fig. 6). However, the differences cannot explain the timing difference as shown in Fig. 3. Regional offsets in $^{14}C$ might also exist.[40] This could be important for high-accuracy $^{14}C$ dating in certain regions.[41] However, regional $^{14}C$ differences

[38] Marchal *et al.* 2001.
[39] Reimer *et al.* 2004.
[40] Kromer *et al.* 2001.
[41] Kromer *et al.* 2001.

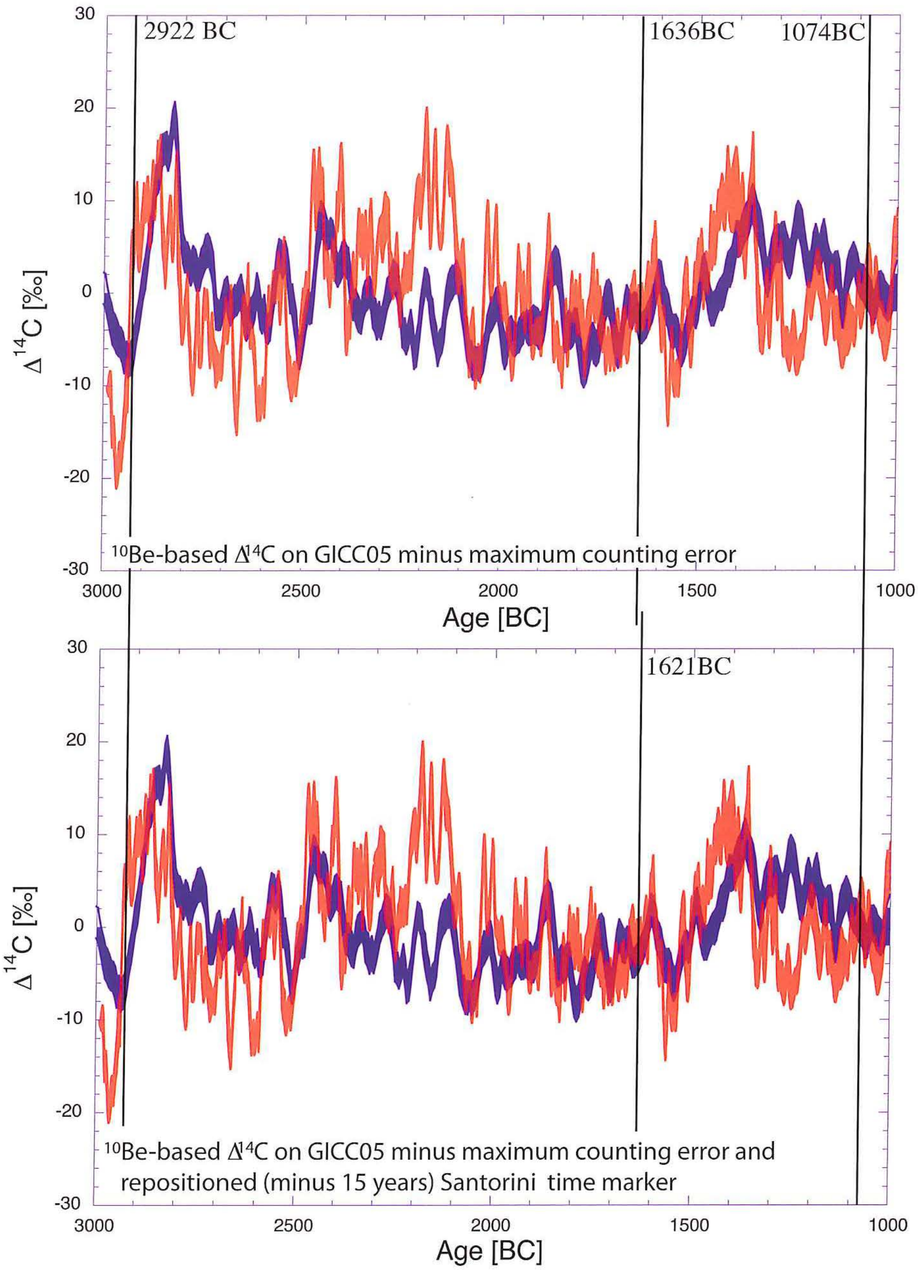



Fig. 7. Comparison of measured (blue band) and modelled $\Delta^{14}C$ (red band). The upper panel shows the comparison after the maximum counting error was subtracted from the GICC05 time scale. The lower panel shows this comparison after the Santorini time marker was repositioned at 1621 BC. The vertical lines indicate the locations of the ECM spikes that have been used to synchronise the Greenland ice cores.

cannot explain the systematic $^{10}$Be-$^{14}$C timing difference since the Intcal04 calibration curve is based on records from different regions in the world.[42]

In summary, in particular potential changes in the $^{10}$Be deposition and data uncertainties could explain part of the differences shown in Fig. 3. However, it seems unlikely that these uncertainties could explain the complete 20-years shift between the ice core and tree ring time scales. Carbon cycle uncertainties could explain a systematic shift but these are probably also smaller than the suggested 20 years.

Therefore, there does not seem to be a mechanism that could explain all of the $^{10}$Be-$^{14}$C timing difference without considering dating uncertainties. However, dating uncertainties of the $\Delta^{14}C$ record seem unlikely since the tree ring chronology is based on several millennia-long chronologies with internal replications of overlapping sections. In addition, cross-checks between different inde-

[42] Reimer *et al.* 2004.

pendent chronologies were made whenever possible.[43] Therefore, it is assumed that the $^{14}C$ calibration record based on tree-ring chronologies is absolutely dated back to 12,410 cal BP leaving virtually no room for dating uncertainties.[44] The ice core time scales contain more uncertainties. The maximum counting error around the Santorini eruption is estimated to be 5 years. The GICC05 ice core time scale was based on the ice cores that showed the most reliable annual signals and this dating was subsequently transferred to other ice cores via the ECM signal. The DYE3 ice core provides the best annual signals around the Santorini eruption.[45] Therefore, potential dating problems in the DYE3 ice core could be transferred to the complete Greenland ice core chronology. Fig. 7 shows the $^{10}Be$-$^{14}C$ comparison under the assumptions that (i) the errors of the GICC05 time scale represent an over-counting and must therefore be subtracted, and that (ii) the date of the Santorini time marker is too old by an additional 15 years. Altogether this would shift the ice core dating of the Santorini eruption to 1622 BC. It appears from Fig. 7 that the timing between $^{14}C$ and $^{10}Be$ is improved by these assumptions. It therefore seems likely that uncertain years in the ice core time scale do not represent real years. However, the shift of these 15 years cannot be explained within the errors of the ice core time scale. It is difficult to explain this additional shift since the layer attributed to the Santorini eruption was also independently dated in the GRIP and GISP2 ice cores yielding ages of 1636±7 BC in the GRIP ice core and 1670±21 BC and 1673±21 BC in the GISP2 ice core.[46] This makes it unlikely that a potential problem in the DYE3 ice core dating is responsible for the 20-year difference between $^{10}Be$ and $^{14}C$ records.

# Conclusion

The $^{10}Be$-$^{14}C$ comparison suggests a similar time scale difference around 1620 BC as shown by the individual age determinations of the Minoan eruption of Santorini in the ice cores and the olive tree. This result confirms the identification of the volcanic reference horizons in the ice cores and the dating of the olive tree. However, there is the time scale difference of approximately 20 years that is larger than the errors given for the dating of the olive tree and the errors of the ice core time scale. Considering both dating and data uncertainties, the 20-year difference does not throw doubt on the Radiocarbon dating of the Santorini eruption, nor does it cast doubt on the identification of the "Santorini layer" in the ice cores. However, a final conclusion about the origins of this suggested time scale difference cannot be given. Yet it seems likely that a combination of several uncertainties could add up to the 20-year difference between the $^{10}Be$ and $^{14}C$ records around the time of the Santorini eruption.

# Acknowledgment

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[43] Reimer *et al.* 2004 and references there.
[44] Friedrich *et al.* 2004.
[45] Vinther *et al.* 2006.
[46] Vinther *et al.* 2006 and references there.