Transferring radiometric dating of the last interglacial sea level high stand to marine and ice core records

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Abstract

In order to derive a radiometric age marker for the end of the penultimate glacial–interglacial transition, we compiled published U-series isotope measurements on corals from the period extending from stage 6 to the middle of the last interglacial, and computed the corresponding open-system ages using Thompson et al. model (Thompson, W.G., Spiegelman, M.W., Goldstein, S.L., Speed, R.C., An open-system model for U-series age determinations of fossil corals. Earth Planet. Sci. Lett. 210 (2003) 365–381). We obtain a global mean age of 126 calendar kyr BP (ka) ± 1.7kyr (2σ) for the beginning of the last interglacial sea level high stand. After showing that the phase relationships observed between changes in sea level, North Atlantic benthic and planktonic foraminifera oxygen isotopic records, and atmospheric methane over the last deglaciation were likely also valid over the penultimate deglaciation, we derive an age of 131.2ka ± 2kyr (2σ) for the abrupt increase in atmospheric CH4 and North Atlantic surface temperature marking the end of the penultimate glacial–interglacial transition. This age is consistent with U–Th dates of the penultimate glacial–interglacial transition recorded in speleothems from sites where speleothems isotopic records are synchronous with North Atlantic temperature records over the last deglaciation. Finally, we show that the phase obtained between the climatic response and northern hemisphere summer insolation is not constant from Termination II to Termination I, implying that northern hemisphere summer insolation alone cannot explain the timing of terminations.

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1. Introduction

Uncertainty on polar ice and marine cores dating remains very large prior to the time span covered by 14C dating techniques or by layer counting (i.e., approximately the last 15 to 40kyr). Prior to 40 calendar kyr BP (ka), marine cores are generally dated by correlation to SPECMAP reference records (Imbrie et al., 1984; Martinson et al., 1987; Bassinot et al., 1994) which are themselves dated by orbital tuning. This technique assumes constant phases between the obliquity and precession components of insolation and of planktonic or benthic δ18O throughout the record and produces chronologies with uncertainties of about 5kyr (i.e. roughly a quarter of a precessional cycle) along the entire record.

One advantage of central Antarctic ice cores over deep-sea cores is that it is possible to use ice flow and accumulation models to date ice core records (Parrenin et al., 2004). Since snow accumulation processes are...
relatively regular in Central Antarctica and ice cores records resolution is usually very high (typically 100 yr), the duration of climatic events (and thus climatic phasing) should be well estimated in ice cores chronologies. It is thus generally accepted that the most promising approach to obtain consistent chronologies for polar ice and marine records is to define a reference ice core chronology and transfer it to deep-sea cores, using markers present in both archives as tie points.

However, glaciological models alone do not yield accurate absolute ages because of poorly known parameters (e.g. glacial accumulation rate, basal melting and sliding). This is why Parrenin et al. (2001) developed an inverse method in order to derive model parameters that give optimal agreement with independent absolute age markers. This method provides ice core chronologies that are accurate with respect to both absolute ages and events durations. Published age scales for Dome C, Dome Fuji and Vostok were derived using no accurate ages markers prior to 41ka but wide orbital control windows instead. As a consequence, there are currently large discrepancies between the timing of the penultimate glacial–interglacial transition in these three East Antarctica ice cores age scales, although the shape of the records is very similar from one site to another. More specifically, the mid point of the penultimate deglaciation is dated at about 135.3ka in Dome F ice oxygen isotopic ($\delta^{18}$O) record (Watanabe et al., 2003), whereas it takes place around 133.9ka in Vostok ice deuterium ($\delta^D$) record (Parrenin et al., 2004), around 130.6ka in Dome C $\delta^D$ record on the EDC2 time scale (EPICA, 2004) and around 132.4ka in Dome C $\delta^D$ record on the latest EDC3 time scale (Parrenin et al., 2007). The discrepancies amount to roughly 5 ky, consistently with the wide orbital windows used for each age scale.

This large uncertainty on polar ice and marine cores dating around 130ka contributes to maintain debate on the mechanisms responsible for deglaciations: it is currently still unclear if changes in summer northern hemisphere insolation are the main forcing factor triggering deglaciations as proposed by (Milankovitch, 1941).

There is thus a need for an absolute reference age scale for marine and ice cores records. The purpose of the present paper is to provide a radiometric age marker for the beginning of the last interglacial in polar ice and North Atlantic deep-sea cores.

2. Method

Our approach is to examine whether the phase relationships observed between sea level, North Atlantic (benthic and planktonic) foraminifera $\delta^{18}$O records, and atmospheric CH$_4$ over the last deglaciation may also be valid over the penultimate deglaciation. After showing that it is likely the case, we derive a radiometric age for the abrupt increase in atmospheric CH$_4$ and surface temperature in the North Atlantic region marking the end of the penultimate glacial–interglacial transition, from corals U–Th ages of the beginning of stage 5e sea level high stand.

2.1. Preliminary remark on the timing of benthic $\delta^{18}$O and sea level

Benthic $\delta^{18}$O can be decomposed into a global term reflecting changes in global ice volume or sea level, and a local term reflecting changes in local deep water temperature and $\delta^{18}$O. There are currently two sources of deep waters in the North Atlantic basin: (1) relatively warm (2°C) and saline (> 34.9 practical salinity units) North Atlantic deep water (NADW) formed at high northern latitudes; (2) colder (< 0°C) and less saline (< 34.7psu) deep waters formed around Antarctica. At depths of more than ~ 2500m, past deep-water temperature at a given site of the North Atlantic basin depends on the proportion of each of these water masses reaching the site, so that deep water temperature increases when NADW formation intensifies (e.g., Skinner et al., 2003). Changes in surface conditions and ocean circulation during the Younger Dryas and the 3kyr time interval preceding the Bolling–Allerød in the North Atlantic resulted in the formation of brine-generated intermediate waters characterized by low $\delta^{18}$C and $\delta^{18}$O (Labeyrie et al., 2005; Waelbroeck et al., 2006). Therefore, benthic $\delta^{18}$O signals from North Atlantic cores are strongly influenced by changes in deep water temperature and $\delta^{18}$O, that in turn depend on changes in ocean circulation occurring during glacial–interglacial transitions (Skinner et al., 2003; Skinner and Shackleton, 2006). These changes in circulation have been shown to occur rapidly with respect to the slow decrease in mean ocean $\delta^{18}$O resulting from ice-sheets melting (Elliot et al., 2002; McManus et al., 2004; Gherardi et al., 2005). As a consequence, the timing of benthic $\delta^{18}$O signals from North Atlantic cores can not be assumed to follow that of global ice volume or sea level changes across glacial–interglacial transitions. Note that this conclusion is valid for the entire world ocean since benthic $\delta^{18}$O signals from other regions of the world ocean are also sensitive to changes in deep water temperature and $\delta^{18}$O resulting from circulation changes across deglaciations (e.g., Labeyrie et al., 2005; Skinner and Shackleton, 2005).

When examining the sequence of events during the last deglaciation in well-dated North Atlantic marine records, we indeed see that the beginning of the Holocene benthic
δ\(^{18}\)O plateau markedly precedes that of the Holocene sea level high stand. North Atlantic AMS-dated benthic δ\(^{18}\)O records (Table 1) are plotted in Fig. 1. Calendar ages were computed using the Calib5.0.1 software (Stuiver and Reimer, 1993), the 2004 marine calibration curve (Stuiver et al., 1998), a surface reservoir age of 400 y for cores MD95-2037 and SU81-18, and larger reservoir ages at higher latitudes prior to the end of the Younger Dryas (YD) (Waelbroeck et al., 2001). We identify the beginning of the Holocene benthic δ\(^{18}\)O plateau as the first occurrence of benthic δ\(^{18}\)O values equal or lower than the 0–9 calendar ky BP (ka) average. We obtain ages comprised between 9.3ka ± 0.4kyr and 10.5ka ± 0.4kyr, resulting in an average age of 10.1ka ± 0.5kyr (2 sigma) (Fig. 1, Table 1). On the other hand, when U–Th ages are recomputed (Eisenhauer et al., 1993) using decay rates given by (Cheng et al., 2000), U–Th dating of coral terraces yields an age of 6.4ka ± 1kyr for the end of the last deglaciation sea level rise. There is thus a 3.7 ± 1.1kyr lead of the beginning of the Holocene benthic δ\(^{18}\)O plateau with respect to that of the Holocene sea level high stand. This reflects the impact of the resumption of NADW and resulting deep water warming on North Atlantic benthic δ\(^{18}\)O records. This large phase shift

### Table 1
AMS-dated North Atlantic deep-sea cores

<table>
<thead>
<tr>
<th>Core</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Water depth (m)</th>
<th>Beginning of Holocene benthic δ(^{18})O plateau (ka)</th>
<th>Age error (2σ) (ky)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>NA87-22</td>
<td>55°30' N</td>
<td>14°42' W</td>
<td>2161</td>
<td>10.1</td>
<td>0.7</td>
<td>Waelbroeck et al., 2001</td>
</tr>
<tr>
<td>CH69-K09b</td>
<td>41°45' N</td>
<td>47°21' W</td>
<td>4100</td>
<td>10.3</td>
<td>0.5</td>
<td>Waelbroeck et al., 2001</td>
</tr>
<tr>
<td>SU81-18</td>
<td>37°48' N</td>
<td>10°10' W</td>
<td>3135</td>
<td>10.5</td>
<td>0.4</td>
<td>Duplessy et al., 1992</td>
</tr>
<tr>
<td>MD95-2037</td>
<td>37°05' N</td>
<td>32°02' W</td>
<td>2159</td>
<td>9.3</td>
<td>0.4</td>
<td>Labeyrie et al., 2005</td>
</tr>
</tbody>
</table>

\(^{a}\) Errors include the 2σ error computed by the Calib5 0.1. software and the error resulting from sampling resolution.

\(^{b}\) The mean between the beginning of the Holocene δ\(^{18}\)O plateau given by the Melonis and Cibicides curve was taken.

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\[\text{Fig. 1. North Atlantic benthic } \delta^{18} \text{O and ice-volume equivalent sea level (Lambeck and Chappell, 2001) versus calendar age over the last deglaciation. Complete coordinates and references of benthic records are given in Table 1. The grey stripe indicates the beginning of the Holocene benthic } \delta^{18} \text{O plateau in North Atlantic cores.}\]
between benthic $\delta^{18}O$ decrease and sea level rise demonstrates that North Atlantic benthic $\delta^{18}O$ can not be used as a sea level proxy at suborbital time scales, so that corals U–Th ages cannot be directly transferred to North Atlantic foraminifera $\delta^{18}O$ curves with a precision better than about 5kyr.

2.2. Comparison of the sequence of climatic events over Termination II and I

In order to examine whether phase relationships measured over the last deglaciation could potentially be also valid over the penultimate deglaciation, we selected North Atlantic marine records in which sea surface temperature (SST) has been reconstructed and for which benthic and planktonic foraminifera isotopes have been measured over both terminations at sufficiently high resolution (Fig. 2). Comparing the two terminations, we reach the following conclusions:

- the end of *Globigerinoides bulloides* $\delta^{18}O$ decrease takes place at the same time as the end of the benthic $\delta^{18}O$ decrease for both terminations within ± 0.3 ky (Fig. 2);
- rapid SST increases occur synchronously with *G. bulloides* $\delta^{18}O$ decreases for both terminations. This confirms that *G. bulloides* $\delta^{18}O$ may be considered as a first order proxy of SST in the North Atlantic region (Shackleton et al., 2000). An uncertainty of ± 0.5 ky is associated with the use of *G. bulloides* $\delta^{18}O$ as a proxy of SST in the North Atlantic region (Landais et al., 2006).

Similar phasing between *G. bulloides* and benthic $\delta^{18}O$ records over both terminations (Fig. 2) indicates that changes in North Atlantic surface and deep water temperature followed the same sequence during Termination II and I. This reflects the fact that the response time of the deep ocean with respect to changes in surface conditions is the same during Termination II and I and involves the same physical mechanisms. The similarity between Termination II and I has been noted in various previous studies (Lotostkaya and Ganssen, 1999; Skinner and Shackleton, 2006).

We assume in what follows that the phase between the beginning of the sea level high stand and the rapid increase in North Atlantic SST was similar over both terminations. We are aware that there are some uncertainties with respect to this assumption and that some aspects of the two terminations are different, such as the presence of the Younger Dryas cold episode during the last deglaciation, or the fact that the orbital configuration was not the same over both terminations. In what follows, we will thus examine the implications on dating of the above assumption. We will subsequently discuss whether our conclusions are consistent with other dated archives or not.

2.3. Link between North Atlantic marine records and ice core records

Precise absolute dating of the rapid temperature increase marking the transition out of the Younger Dryas cold episode has been established by annual layer counting of Greenland ice cores. The transition was dated at 11.65ka ± 0.10kyr in the GRIP and NorthGRIP ice cores on the latest GICC05 age scale (Rasmussen et al., 2006). This rapid warming has also been recorded in $^{14}C$-dated North Atlantic SST records and in terrestrial air temperature records at the same calendar date (Duplessy et al., 1992; Bjork et al., 1996; Waelbroeck et al., 2001). Severinghaus et al. (1998) showed that abrupt increases in Greenland air temperature over the last glacial period and last deglaciation were synchronous with abrupt increases in atmospheric $CH_4$. There is thus a $5.2 \pm 1$kyr lead of the abrupt increase in surface temperature in the North Atlantic region and in atmospheric $CH_4$ marking the end of the last deglaciation with respect to the end of the sea level rise or beginning of the Holocene sea level high stand.

2.4. Radiometric dating of the beginning of the last interglacial sea level high stand

We compiled published uranium-series isotope measurements on corals from the period extending from stage 6 to the middle of the last interglacial, collecting data from different sites in order to free ourselves from potential local tectonic effects and obtain a true indication of the timing of global sea level high stand. The compiled data include proximal (Bahamas, Barbados) as well as distal (Western Australia, Huon Peninsula, Hawaii) sites with respect to the massive glacial ice sheets, some of which being tectonically relatively stable (Bahamas, Western Australia), and others tectonically active (Barbados, Huon Peninsula, Hawaii).

It has long been known that U-series isotope measurements on corals are incompatible with closed-system decay (Bard et al., 1992; Stirling et al., 1995). Frank et al. (2006) recently tested three published models that correct for observed open-system behavior of U-series nuclides of unaltered corals (Thompson et al., 2003; Villemant and Feuillet, 2003; Scholz et al., 2004) and demonstrated that the slope of the observed $^{234}U/^{238}U$ and $^{230}Th/^{238}U$ correlation in different reef sections recovered from New
Caledonia, is best reproduced by the Thompson et al. model (Thompson et al., 2003). We thus computed for all compiled U–Th data the open-system ages using Thompson et al. model (Thompson et al., 2003), as well as conventional ages and initial $^{234}$U/$^{238}$U ratio ($\delta^{234}$U), using half-lives given by (Cheng et al., 2000). Recently published values of modern seawater $\delta^{234}$U range from 146.6 ± 1‰ (Robinson et al., 2004 using ICP-MS) to 149.6 ± 1‰ (Delanghe et al., 2002 using TIMS), so we opted for an intermediate value of 148‰. In order to minimize age errors from re-crystallization and other processes not accounted for by the open-system model of (Thompson et al., 2003), we discarded samples with calcite content higher than 2%, $^{232}$Th content higher than 2ppb and initial $\delta^{234}$U different from 148 ± 10‰. The resulting data subset is listed in the supplementary table online. Fig. 3 displays fossil corals initial elevation versus computed open-system age for supplementary table samples. As some coral species live in deeper waters, we drew a tentative sea level curve through highest
elevation corals of a given age. Following (Thompson and Goldstein, 2005), we did not include Barbados sample UWI93-2 (Gallup et al., 2002) in the sea level curve, as its level is in conflict with other data in the 128–129ka age range.

In order to define the age of the beginning of the last interglacial sea level high stand, we identified the oldest corals found at or above present-day sea level in different geographical locations. Two Barbados coral samples with initial elevations of 19 and 25m below the modern sea level (bsl) have ages comprised between 127 and 131ka (Fig. 3). The coral species of the Barbados sample found at 25m bsl (UWI-101) is Acropora palmata, meaning that sea level must have been within 5m above that level. Corals found in Alladin Cave (Huon Peninsula) at an initial elevation of 75m bsl have ages comprised between 123 and 133ka. The latter coral samples are Porites sp. and Faviidae sp., which can live at water depths of more than 27m. Sea level seems thus to have dropped to a level intermediate between 20 and 75m bsl at some time during this age interval, either before or after the episode marked by the Barbados Acropora coral found at 20m bsl. Thereafter, the first samples found at or above present-day sea level are samples from the Bahamas (open-system age of 126.4ka ± 2.8kyr), Western Australia (open-system age of 126.1ka ± 0.8kyr) and Hawaii (open-system age of 125.5ka ± 1.2kyr). In cases where the age uncertainty interval of a sample was completely included within the uncertainty interval of an older sample, we retained the most precise age. The first Barbados sample found at or above present-day sea level is significantly younger (open-system age of 123.8ka ± 1.4kyr) than those from the Bahamas, Western Australia and Hawaii. Therefore, we pooled the ages found for the three first sites and defined the beginning of the last interglacial sea level high stand by concatenating the corresponding ages. This statistical procedure is best suited for small data sets and allows one to estimate a realistic error by averaging the respective error bars rather than deriving standard deviations through normal distribution laws. The resulting age of the beginning of stage 5e sea level high stand is 126ka ± 1.6kyr.

This age is substantially older than the open-system age of 123.7ka ± 1.3kyr computed by Thompson and Goldstein (Thompson and Goldstein, 2005, 2006) from Barbados corals. There are two reasons for this discrepancy: (1) we used a value of 148 ‰ for the modern seawater δ234U, whereas Thompson and Goldstein (Thompson and Goldstein, 2005, 2006) used a value of 145 ‰ and hence obtained open-system ages that are systematically about 1.2kyr younger than ours; (2) we did not see any reason to discard coral data from Western Australia, Bahamas and Hawaii, that have been stratigraphically and morphologically described as certainly belonging to stage 5e high sea level stand by their authors (Chen et al., 1991; Stirling et al., 1995, 1998; Muhs et al., 2002). Therefore, although we did not retain cobble or rubblestone samples, our data set contains stage 5e coral...
data from other sites than Barbados that were not taken into account in (Thompson and Goldstein, 2005, 2006).

It should be noted that quoted open-system age errors are often underestimated (Scholz and Mangini, 2007). It is however not possible to estimate by how much these errors are underestimated when the coral sample has not been divided into sub-samples. In the present study, we could compute the age variability among different sub-samples for 11 corals out of the 80 corals listed in the supplementary table. We found an age variability larger than the quoted error for only three corals, with no incidence on the estimated age of the beginning of stage 5e sea level high stand. In contrast, one can easily assess that a 1.5‰ uncertainty on the modern seawater δ²³⁴U value translates into an additional uncertainty on open-system ages of about 600 y. Accounting for this additional source of uncertainty, our final estimate of the age of the beginning of stage 5e sea level high stand is 126ka ± 1.7kyr.

3. Results and discussion

Following our assumption of similar phasing between the beginning of the sea level high stand and the rapid increase in North Atlantic SST at the end of Termination II and I, we thus estimate that the abrupt increase in North Atlantic surface temperature and atmospheric CH₄ that marked the end of Termination II took place 5.2 ± 1kyr earlier, i.e., at 131.2ka ± 2kyr.

3.1. Consequences on glaciological timescales

Because of the rapid mixing time of the atmosphere (∼1 year between hemispheres) and 10years lifetime of atmospheric CH₄, large-scale changes in atmospheric CH₄ concentration are essentially globally synchronous. This synchronism provides a tool for correlating Greenland and Antarctica ice core chronologies (Blunier et al., 1998; Blunier and Brook, 2001; Delmote et al., 2004).

The new age marker we just derived can thus be compared with the age currently given by ice core chronologies for the rapid increase in CH₄ at the end of Termination II in Antarctica, that is, 126.8ka in EPICA Dome C EDC2 chronology (EPICA, 2004), 129.2ka in EPICA Dome C EDC3 chronology (Parrenin et al., 2007), and 129ka in Vostok FGT1 gas age scale (Parrenin et al., 2004). Among these ages, only EDC2 age is significantly younger than the age derived from open-system coral dates. This indicates that the glaciological model used to derive EDC2 chronology did not succeed in correctly simulating ice thinning and/or snow accumulation over that portion of the record. Parrenin et al. (2007) indeed showed that the formulation used in EDC2 resulted in overestimated snow accumulation values during glacial.

The timing of changes in δ¹⁸O of atmospheric O₂ (δ¹⁸O atm) with respect to changes in foraminifera δ¹⁸O can be used as an independent evaluation of the assumption that the abrupt CH₄ rise was indeed synchronous with the rapid North Atlantic SST increase at the end of the penultimate glacial–interglacial transition. Changes in δ¹⁸O atm over terminations reflect changes in average surface water δ¹⁸O and in the ratio of terrestrial versus marine primary production, with a variable time lag of 1 to 2kyr resulting from O₂ residence time in the atmosphere (Bender et al., 1994; Jouzel et al., 2002). Assuming that changes in δ¹⁸O atm over terminations primarily reflect changes in mean ocean δ¹⁸O, the δ¹⁸O atm signal should thus lag benthic and planktonic δ¹⁸O over Termination II and I. This can be easily verified over the last deglaciation since absolute dating is available over that period for ice and marine core records (Fig.4). In order to visualize the impact of the new age constraints on the various ice core and marine core records, we linearly modified Vostok FGT1 age scale over the 245–75ka time period by assigning an age of 131.2ka to the rapid increase in CH₄ marking the end of Termination II (Fig. 4). Similarly, we suppressed the orbital control point over Termination II in marine cores SPECMAP age scales and assigned an age of 131.2ka to the rapid increase in North Atlantic SST (or decrease of G. bulloides δ¹⁸O) at the end of Termination II. Fig. 4 shows that the new ice and marine cores chronologies are mutually consistent over Termination II and do not violate the causal relationship implying that δ¹⁸O atm should lag foraminifera δ¹⁸O over terminations if it mainly reflects changes in mean ocean δ¹⁸O.

Furthermore, Fig. 4 illustrates that the proposed ice core chronology derived from open-system coral ages implies that the phase between δ¹⁸O atm and northern hemisphere summer insolation is not constant from one termination to another. This is in contradiction with previous ice core chronologies based on the assumption that the phase between δ¹⁸O atm and northern hemisphere summer insolation or precession is constant over time (Petit et al., 1999; Shackleton, 2000) and confirms studies indicating that phasing between δ¹⁸O atm and insolation could be variable over time (Malalizé et al., 1999; Hoffmann et al., 2004).

3.2. Consequences on deep-sea cores timescales

Adopting a definition of the beginning of the last interglacial benthic δ¹⁸O plateau similar to the one we
used to define the beginning of the Holocene benthic $\delta^{18}O$ plateau, we obtain an age of 130.4ka ± 2kyr for the beginning of the North Atlantic last interglacial benthic $\delta^{18}O$ plateau. This age is significantly older than the SPECMAP age of 125.2ka ± 2.9kyr (Martinson et al., 1987). In other words, the proposed new dating derived from open-system coral ages implies that the phase between the precession and obliquity components of benthic $\delta^{18}O$ and insolation is not constant over time. Estimating the phase between the orbital parameters and the beginning of the sea level high stand over both terminations, we obtain a lead of 5 ± 1kyr of the maximum in northern hemisphere June insolation (or the minimum in precession) with respect to the beginning of the Holocene sea level high stand (mid-June insolation at 65°N is maximum at 11.4ka and sea level reaches 0m at 6.4ka ± 1kyr Eisenhauer et al., 1993), whereas, this lead is of only 2.1 ± 1.7kyr over the penultimate deglaciation (mid-June insolation at 65°N is maximum at 128.1ka and sea level reaches present-day levels at 126ka ± 1.7kyr).

Fig. 4. A. Obliquity (thin gray line) and 65°N mid-June insolation (bold black line) computed versus calendar age according to Berger’s formulas (Berger, 1978) with the Analyses software (Paillard et al., 1996). B. GRIP atmospheric CH4 record versus calendar age over Termination I (Dansgaard et al., 1993) and Vostok atmospheric CH4 records versus this study age scale over Termination II (Parrenin et al., 2004; Delmotte et al., 2004). C. G. bulloides $\delta^{18}O$ for North Atlantic core CH69-K09 (bold gray line) (Labeyrie et al., 1999; Waelbroeck, et al., 2001) and MD95-2042 (thin black line) (Shackleton et al., 2003) versus calendar age over Term. I, and versus the new age scale over Term. II. D. Same as in C, but for mixed benthic foraminifera $\delta^{18}O$ in core MD95-2042, and for Melonis barleanum $\delta^{18}O$ corrected by +0.4‰ (Duplessy et al., 1980) in core CH69-K09. E. Vostok $\delta^{18}O$ of atmospheric O2 versus calendar age over Term. I (Sowers et al., 1991), and versus this study age scale over Term. II (Malaizé et al., 1999; Parrenin et al., 2004).
Similarly, the lead of the maximum in obliquity with respect to the beginning of the sea level high stand is different for the last and penultimate deglaciation (3 ± 1 kyr versus 5.2 ± 1.7 kyr, respectively), although this difference in phasing is smaller than between sea level high stand and precession or northern hemisphere summer insolation.

These results reinforce the conclusions of previous studies that provided evidence against a constant phase between the climatic response and orbital parameters (Parrenin et al., 2001; Parrenin and Paillard, 2003). In particular, this proposed new chronology is consistent with some results indicating that the midpoint of sea level rise over Termination II took likely place around 135 kyr (Henderson and Slowey, 2000; Gallup et al., 2002), as opposed to the SPECMAP age of 129.8 kyr ± 3 kyr for the midpoint of planktonic or benthic δ18O transition (Martinson et al., 1987).

3.3. Comparison with speleothems records

A few U-Th dated speleothems isotopic records (δ18O and δ13C) cover the last interglacial time interval in different parts of the World: United Kingdom (Gordon et al., 1989); Norway (Lauritzen, 1995); Romania (Lauritzen and Onac, 1999); Israël (Frumkin et al., 1999; Bar-Matthews et al., 2003); Austria (Holzkämper et al., 2004; Spötl et al., 2007); Italy (Drysdale et al., 2005); China (Yuan et al., 2004; Cheng et al., 2006; Kelly et al., 2006). However, only isotopic records that are relatively well understood and cross-cover the whole stage 6/5 transition can be used for an accurate estimation of the age of the penultimate glacial–interglacial transition. The δ18O records of the Chinese caves exhibit an extremely abrupt stage 6/5 transition, dated at 129.0 kyr ± 0.9 kyr in the Dongge Cave (Kelly et al., 2006) and 129 kyr ± 1 kyr in the Hulu Cave (Cheng et al., 2006). Closer to the North Atlantic area, the δ18O record of the Corchia cave displays a 6/5 transition between 134.0 kyr ± 2 kyr and 129.0 kyr ± 1 kyr (Drysdale et al., 2005). The 134.0 kyr ± 2 kyr age coincides with the beginning of the δ18O change, but not with the most abrupt transition which occurs just after 130.0 kyr ± 2 kyr. This is consistent with speleothem records from a high-elevation Alpine cave (Spötl et al., 2007). Similarly, the δ18O record of the Peqin Cave in Northern Israel indicates that the end of the stage 6/5 transition took place at 130.9 kyr ± 2.7 kyr (Bar-Matthews et al., 2003). In all these cases, the actual 2σ error of the transition should be theoretically slightly larger due to the linear interpolation between the dated points and to the growth rate changes that occurred during the climatic transition. This error is likely small in records that have a high density of measured ages, which is the case here, but we must keep in mind that just before the 6/5 transition, growth rates are generally much lower than during stage 5 so that the time interval covered by each sample used for U-Th dating can in certain cases be larger than the analytical error on the age.

In order to ascertain the meaning of the climatic signal, we only retain in this discussion dated isotopic records that also cover the last 20 to 30 kyr and record a climatic signal that is in phase with the Greenland ice isotopic signal over the last deglaciation: the Peqin Cave (Israël, Bar-Matthews et al., 2003), Dongge Cave (China, Yuan et al., 2004; Kelly et al., 2006), and Hulu Cave (China, Cheng et al., 2006). In these records, the δ18O variation is clearly interpreted by the authors as an “amount effect”: more rainfall leading to lighter δ18O. Heavy rainfall periods are related with humid and generally warm episodes: Sapropel events in the Mediterranean area and Asian Monsoon for the China records.

Because speleothems δ18O depends on several factors (e.g., the amount of precipitation, the air temperature, the changes in depression pathways and altitudinal gradients), the correlation with other proxies extracted from other archives, such as the atmospheric CH4 or the surface temperature in the North Atlantic region, is not straightforward. However, recent modeling studies have shown that there is a physical basis for a direct response of the Asian Monsoon to changes in climatic conditions in the North Atlantic area. A decrease in the extent of northern hemisphere ice sheet (Chiang et al., 2003), and a strengthening of the Atlantic thermohaline circulation (Zhang and Delworth, 2005), have been shown to induce a northward displacement of the Inter Tropical Convergence Zone.

Considering that the above speleothem δ18O records can be matched with surface temperature in the North Atlantic region, as shown by (Genty et al., 2003) for stages 4 and 3 on SW-France speleothems, we see that the age marker derived from coral open-system ages is consistent with speleothem datings within error bars. We thus have here an a posteriori verification of our assumption of similar phasing between the beginning of the sea level high stand and the rapid increase in CH4 and North Atlantic temperatures over Termination II and I.

4. Summary and conclusions

Assuming similar phasing between the beginning of sea level high stand and the rapid increase in CH4 and North Atlantic temperatures over Termination II and I, we obtained a radiometric age marker for ice cores and North Atlantic deep-sea cores, based on open-system
coral dating of the beginning of the penultimate interglacial sea level high stand.

Open-system dating of stage 5e corals found at or above present-day sea level yields an age of 126ka ± 1.7kyr for the beginning of the penultimate sea level high stand. This age translates into an age of 131.2ka ± 2kyr for the rapid increase in CH4 and North Atlantic temperatures marking the beginning of last interglacial.

In the resulting new chronology, the beginning of the last interglacial benthic δ18O plateau is significantly older than in the SPECMAP age scale. We also show that in the proposed dating, the phase between the climatic response and northern hemisphere summer insolation is not constant from one termination to the other (Fig. 4). Northern hemisphere summer insolation alone can thus not explain the timing of terminations, contrarily to what was first suggested by Milankovitch (1941).

The age marker derived in the present study is consistent with U–Th dates of the penultimate glacial–interglacial transition recorded in speleothems from sites where speleothems isotopic records are synchronous with North Atlantic temperature records over the last deglaciation (i.e., Peqin Cave in Israël Bar-Matthews et al., 2003, Hulu and Dongge Caves in China Yuan et al., 2004; Cheng et al., 2006). Our results are thus compatible with the assumption that abrupt changes in monsoon are different from the CH4 or the North Atlantic temperature records over the last deglaciation intervals. Geochim. Cosmochim. Acta 67, 3181 –3199. Bassinot, F., Labeyrie, L., Vincent, E., Quideulleur, X., Shackleton, N., Lancelot, Y., 1994. The astronomical theory of climate and the age of the Brunhes–Matuyama magnetic reversal. Earth Planet. Sci. Lett. 126, 91–108.


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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2007.10.006.

References


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