Calibration of stable isotope composition of Thoracosphaera heimii (dinoflagellate) calcite for reconstructing paleotemperatures in the intermediate photic zone

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Abstract In this study we investigate the temperature dependence of oxygen isotope ratios preserved in calcite formed by the dinoflagellate Thoracosphaera heimii, focusing primarily on the development of a geological proxy. Geochemical analysis of the calcite shells produced by this species represents a valuable proxy for reconstructing environmental conditions in the intermediate photic zone. Calibration is based on isotopic analysis from culture experiments performed in very dilute batch conditions, as well as from near-monospecific T. heimii assemblages separated from core top sediments. Results are similar for both approaches and indicate that T. heimii shells have oxygen isotope compositions close to equilibrium values predicted for inorganic calcite precipitation. This calibration of the isotopic composition of dinoflagellate calcite indicates that monospecific assemblages of T. heimii can be used to unravel paleotemperatures in the intermediate photic zone by applying isotopic transfer functions for equilibrium calcite. In culture, however, a δ^{18}O offset of ~1‰ is observed at temperatures <17°C, which falls below the natural temperature range of this species. Culture analyses also reveal a relationship between temperature and carbon isotope composition of calcite. The mechanisms behind this relationship remain to be explored, but their identification may provide a better understanding of carbon isotope systematics from both biogeochemical and geological perspectives. Comparison of the oxygen isotope composition of T. heimii shells with that of shallower dwelling organisms, such as the coccolithophores, represents a valuable proxy for determining temperature gradients within the photic zone and may enable reconstruction of the evolution of the depth of the thermocline.

1. Introduction

Generating reliable sea surface temperatures (SSTs) is a principle aim for paleoceanographers and is fundamental for successful modeling of climate sensitivity in the past, present, and future. This has motivated considerable effort in the development of paleotemperature proxies, yet uncertainties remain on the fidelity of SST estimates. Most of the studies aiming to derive paleotemperatures are based on planktic foraminifera oxygen isotope composition (δ^{18}O) because their tests can be easily isolated from sediments by hand-picking.

Calcareous phytoplankton algae, such as coccolithophores and calcareous dinoflagellates, thrive in the photic zone, and hence, calcification takes place close to the interface between the atmosphere and the ocean. The geochemistry of their shells may hence represent a valuable addition to the paleoceanography toolbox to augment data from foraminifera enabling more precise reconstruction of the physico-chemistry of the photic zone. The increasing interest in phytoplankton geochemistry has been made possible by recent development and application of sedimentological techniques enabling the isolation and geochemical analysis of their micron-sized calcified shells [Minoletti et al., 2001; Stoll and Ziveri, 2002; Minoletti et al., 2004; Minoletti et al., 2005; Stoll, 2005; Zonneveld et al., 2007; Hermoso et al., 2009; Minoletti et al., 2009; Bolton et al., 2012; Rouselle et al., 2013].

Calcareous dinoflagellates mainly thrive in warm, low-latitude waters often characterized by low nutrient levels [Zonneveld et al., 2000; Kohn and Zonneveld, 2010]. In situ measurements have shown that maximum abundance of living cells is found in the photic zone below the mixed layer although it has to be acknowledged that this observation comes from very dilute cell counts [Karwath et al., 2000; Vink et al., 2000; Wendler et al., 2002a, 2002b; Vink, 2004; Kohn and Zonneveld, 2010]. This putative deeper habitat with respect to that of the coccolithophores owes to their intolerance to high-energy water motion [Thomas and Gibson, 1990].
isotope data indicating that temperature might exert a second-order control on proxy based on monospecific T. heimii. The discrepancies with previous studies, here we attempt a mechanistic understanding of the vital effect in cascade of microcultured calcispheres, with replicates comprised within a mean 1‰ array for the same temperature. The relatively negative δ18O values indicate a vital effect toward isotopically light values. Hence, these data indicate that T. heimii calcite can be tentatively assigned to the “light group” (sensu Dudley et al. [1986]). Conversely, in the core top study, no correlation is found between temperature and δ18O, with significantly more positive δ18O values with respect to the culture data. Another contrasting feature with the coccolithophores is the mode of biomineralization of T. heimii dinoflagellates with a combined external and internal precipitation of calcite, as originally described based on morphological observations of living cells [Tangen et al., 1982]. This feature has been subsequently supported by geochemical evidence [Gussone et al., 2010; Van de Waal et al., 2013]. Morphological and molecular investigations indicate that T. heimii-like morphologies and genes are present since the Upper Cretaceous [Hildebrand-Habel et al., 1999; Gottschling et al., 2008]. This long lasting record indicates that monospecific T. heimii analyses may enable a long-term investigation of the thermal evolution of the intermediate photic zone.

To date, the comprehensive culture study of Zonneveld et al. [2007] is the only one that has measured 18O fractionation (approximated from the difference between δ18Ov and δ18Omedium) in T. heimii calcite under a wide range of growth temperatures (Figure 1). The authors observed a strong linear correlation (r² = 0.83) between temperature and oxygen isotope composition of T. heimii calcite. Compared to inorganic prediction [Kim and O’Neil, 1997], T. heimii calcite was measured with significantly negative δ18Oc, with a mean offset of −3‰. In sediments, despite the careful methodological approach presented by Zonneveld [2004], no convincing correlation could be found between temperature and δ18O of T. heimii in Atlantic Ocean core tops (r² = 0.42) (Figure 1). Values of δ18Oc obtained from separated fractions were also significantly higher (~1.5‰) than culture residues. Hence, the inconsistency between culture and core top data makes the achieved calibration hardly transferable to the geological record [Kohn and Zonneveld, 2010]. Recent reappraisals of the magnitude of the vital effect, i.e., the isotopic offset between biominerals and inorganic calcite precipitated from the same fluid, in “geologically relevant” coccoliths have highlighted the influence of the culture method on the isotopic composition of calcite (see review by Hermoso [2014]). Discrepancies in the magnitude of vital effects between studies are the likely consequence of culture artifacts—primarily due to a reservoir effect affecting the pool of dissolved inorganic carbon (DIC) [Hermoso et al., 2014]. Indeed, when cultured at very low cell density, coccolith calcite has been reported closer to equilibrium and matches the core top data [Candelier et al., 2013].

In this study, we attempt an integrated culture and core top calibration using state-of-the-art methodologies to calibrate stable isotope composition for T. heimii with a geological (Cenozoic) perspective. We have carried out very dilute batch cultures in the laboratory to minimize possible culture artifacts and applied a microseparation technique to produce near-monospecific (>90%) T. heimii fractions that is based on a cascade of micro-filtering steps [Minoletti et al., 2009]. Beyond a simple empirical calibration and discussion of the discrepancies with previous studies, here we attempt a mechanistic understanding of the vital effect in T. heimii affecting the oxygen isotope composition, and we discuss the applicability of the paleotemperature proxy based on monospecific fractions of T. heimii. In addition to oxygen isotopes, we present carbon isotope data indicating that temperature might exert a second-order control on δ13C calcite in T. heimii.

2. Material and Methods
2.1. Culture Setup
Studies based on laboratory cultures of calcifying organisms have provided calibration equations that enable translation of δ18Oc into temperature estimates for a wide range of planktic calcifiers [Dudley et al., 1986;
Benis et al., 1998; Ziveri et al., 2003; Candelier et al., 2013]. There are many methodological issues, spanning from purely inorganic to physiological aspects that can contribute to altering the isotopic values in laboratory cultures [Hinga et al., 1994; Langer et al., 2006; Barry et al., 2010; Candelier et al., 2013; Hermoso et al., 2014]. This point is well illustrated by the fact that in several cases, significantly different isotopic compositions and fractionation factors have been documented in different studies on the same species (see synthesis in Stoll and Ziveri [2004] and Hermoso et al. [2014]). As most of these artifacts originate from a drift of the carbonate system ([DIC] and pH) in the culture medium, it has been recommended to keep biomass low during batch culture experiments, implying harvesting the bioassay at very dilute cell concentration [Barry et al., 2010].

With these potential artifacts in mind, we conducted dilute batch cultures of a monoclonal strain of Thoracosphaera heimii (RCC 1511 from the Roscoff Culture Collection: http://www.roscoff-culture-collection.org) originated from Japanese coastal waters. This strain was grown at seven temperatures between 13 and 27°C. The batch cultures were undertaken in 2.7 L Nalgene polycarbonate flasks, without a headspace. The culture medium consisted of aged English Channel seawater, with nutrient, trace metal, and chelator enrichments corresponding to K/6 [Keller et al., 1987] and vitamins to that of the f/2 recipe [Guillard, 1975].

Prior to sterilization, media were bubbled with atmospheric air for 24 h. pH was adjusted to 8.2 by addition of 0.2 M NaOH, and sterilization of the culture medium was achieved by filtration through a 0.22 μm pore size Stericup filter device. Cultures were illuminated by using daylight fluorescent bulbs at a light intensity of ~150 μmol photons m⁻² s⁻¹ with a 14/10 light/dark photoperiod.

Prior to the start of the experiment, cultures were gradually acclimated (0.5°C/day) to the target temperature. The initial cell density of T. heimii in the experiments was ~50 cells per mL. Aliquots of the culture medium were sampled at the beginning and end of the experiments, filtered through a 0.22 μm disc-syringe filter and kept at 4°C for subsequent measurement of δ¹⁸Osw of the culture medium. Bubbling of the culture media at 15°C with ambient air was undertaken to ensure that the DIC level and δ¹³C of DIC were in equilibrium with the atmosphere [Zeebe et al., 1999]). By applying the same procedure, Hermoso et al. [2014] measured carbon isotope composition of the DIC in the culturing medium close to 0‰.

The culture flasks were gently shaken twice a day to allow redistribution of the cells in the flask. Cell counts were undertaken using a Sedgewick-Rafter cell containing a subsample of 1 mL of medium and observed under an inverted microscope. Cultures were harvested when cell density reached between 6000 and 15,000 cells per mL. At this stage, population growth was still at the early stage of the exponential phase and therefore under nutrient- and DIC-replete conditions. For each temperature, cultures were implemented in duplicate. At 15°C, two additional flasks were left grown until the population reached late exponential phase in order to assess the effect of the cell density and consequent chemical drift of medium composition on isotope composition in T. heimii calcite.

### 2.2. Core Top Samples

#### 2.2.1. Location of Samples Sites

Core top samples were selected from two oceanographic settings. Seven sites were recovered from intermediate latitudes at ~30°S in the Southern Atlantic and Indian oceans and between 20°N and 40°N in the Northern Atlantic and one from the Northern Somali Basin at 4°N (Table 1). Except for the Somali site, all sites are located within or at the edge of oceanic gyres. The sampling interval was always the topmost few centimeters of each core. The relatively recent age of sedimentary particles enables core top calibration with environmental parameters assumed to be close to present day conditions.

#### 2.2.2. Retrieving Environmental Parameters (T, δ¹⁸Osw) for the Calibration

Calcification of T. heimii takes place in the intermediate photic zone, typically at or just below the base of the mixed layer [Karwath et al., 2000; Vink et al., 2000; Wendler et al., 2002a, 2002b; Vink, 2004; Kohn and Zonneveld, 2010]. To assess the environmental parameters of this water mass, we first evaluated the depth of the mixed layer, the base of which was defined using seawater potential density [Levitus and de Boyer Montégut, 1994; de Boyer Montégut, 2004]. Values of mean annual temperature and δ¹⁸Osw for the intermediate photic zone were then integrated over a 30 m thick water mass below the base of the mixed layer, on average between 50 and 80 m (Table 1). Temperature and δ¹⁸Osw were retrieved from the World Ocean Atlas 2009 [Locarnini et al., 2010] and from the global 1° × 1° gridded compilation of LeGrande and Schmidt [2006]. Temperature values are comprised between 15 and 27°C, which is a similar range to that of the culture experiments (13–27°C).
### Location and Oceanographic Parameters of the Studied Sites

<table>
<thead>
<tr>
<th>Site/station</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (m)</th>
<th>Water Temperature (°C)</th>
<th>IPZ Depth (m)</th>
<th>Sedimentation Rate (cm/kyr)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD09-06</td>
<td>30°S</td>
<td>79°48'E</td>
<td>264192</td>
<td>20.4</td>
<td>359</td>
<td>1.36</td>
<td>Cléroux et al. [2013]</td>
</tr>
<tr>
<td>MD09-06</td>
<td>30°S</td>
<td>79°48'E</td>
<td>264192</td>
<td>20.4</td>
<td>359</td>
<td>1.36</td>
<td>Cléroux et al. [2013]</td>
</tr>
<tr>
<td>VM80-76</td>
<td>33°N</td>
<td>32°W</td>
<td>3567</td>
<td>18.5</td>
<td>157</td>
<td>0.50</td>
<td>Mix and Ruddiman [1985]</td>
</tr>
<tr>
<td>MD104-24</td>
<td>4°N</td>
<td>35°E</td>
<td>2576</td>
<td>25.6</td>
<td>203</td>
<td>0.49</td>
<td>Köhler et al. [1993]</td>
</tr>
<tr>
<td>SU90-03</td>
<td>40°N</td>
<td>31°W</td>
<td>2475</td>
<td>16.6</td>
<td>184</td>
<td>0.82</td>
<td>Chapman et al. [2000]</td>
</tr>
<tr>
<td>SU90-08</td>
<td>43°N</td>
<td>30°W</td>
<td>3080</td>
<td>19.4</td>
<td>184</td>
<td>0.82</td>
<td>Grousset et al. [1993]</td>
</tr>
</tbody>
</table>

In gyre oceanographic regime, production of *T. heimii* is almost constant throughout the year [Zonneveld, 2004; Kohn et al., 2011]. In this study, we used mean annual temperature and δ¹⁸Osw to represent estimates of seawater physico-chemistry. Conversely, the offshore Somali MD104-24 site corresponds to warm waters with a pronounced upwelling-driven seasonality related to a monsoon regime [Fischer et al., 1996; Peeters et al., 2002; Schott et al., 2009]. In this region, the production of *T. heimii* is higher at the end of the SW monsoon (August to September) and at the beginning of the inter-monsoon (October to November), periods that correspond to the lowest and highest annual temperatures, respectively [Wendler et al., 2002a]. As most *T. heimii* shells were formed at either extreme of the temperature range, the mean annual temperature was assumed to integrate this variability. The seasonal variability of the δ¹⁸Osw is not known at this site.

### 2.2.3. Microseparation Protocol

Shells of *T. heimii* were physically isolated from sediments according to the protocol of Minoletti et al. [2001, 2009] based on cascade microfiltering steps. Approximately 1–2 g of bulk sample was suspended in deionized water, previously adjusted to pH 8 by addition of ammonia to minimize calcite dissolution. Gentle stirring allowed complete disaggregation of the sediment. The suspension was first filtered through 315 and 160 μm sieves and then through 40 and 20 μm nylon nets to concentrate foraminifera and their fragments. Finally, the suspension was filtered onto 10 μm pore size polycarbonate membranes (Millipore). Near-monospecific *T. heimii* assemblages were thus gathered from the 10 to 20 μm interval. We chose a purity threshold of 90%, below which fractions were regarded as not sufficiently purified. For those samples in which *T. heimii* assemblages were not pure, we applied a subsequent purification step based on strong ultrasonic treatment and selective breakage of particles by strong ultrasonic supply [Minoletti et al., 2009].

After separation, smear slides were prepared for each fraction using the technique described by Koch and Young [2007]. Relative abundances were estimated by counting at least 300 particles under cross-polarized light using a Zeiss Axioscope 40 microscope equipped with a 63x Plan-Neofluar objective.

Foraminiferal shells and fragments are the most abundant calcite particle in all fractions coarser than 20 μm. The fraction <10 μm contains coccoliths, such as *Coccolithus pelagicus*, *Calcidiscus leptoporus*, *Gephyrocapsa* spp., and *Emiliania huxleyi* and their fragments. Between these two size ranges, the 10–20 μm fraction is composed mainly of *T. heimii* shells, with other rare calcareous components, such as foraminiferal fragments and rare lopadoliths of the coccolithophore *Scyphosphaera apsteinii* (Figures 2a and 2b). In samples MD09-06 and VM30-76, the proportion of *T. heimii* shells was lower than our target purity (<90%), and foraminiferal fragments may contaminate the isotopic signal. In these cases, an additional purification method based short runs (60 s) of strong ultrasonic treatment [Minoletti...
et al., 2009] was applied, and it was possible to induce fragmentation of *T. heimi*i shells while retaining foraminiferal fragments in their original size spectrum. *Thoracosphaera* fragments were subsequently collected in a <10 μm fraction from the original 10–20 μm fractions exposed to strong ultrasonic waves.

### 2.3. Stable Isotope Measurements

Carbonate stable isotopes were analyzed from about 80 μg of culture residue or near-specific *T. heimi*i assemblages from sediments on a VG Isoprime mass spectrometer. Samples were first cleaned using 10% hydrogen peroxide (H₂O₂) to remove the organic matter, rinsed three times, and dried at 60°C. In the instrument they were reacted with purified phosphoric acid (H₃PO₄) at 90°C. Calibration to V-PDB standard via NBS-19 was undertaken daily using the Marceau, a Carrara marble standard. Reproducibility of replicated standards was usually better than 0.05‰ for δ¹³C and 0.1‰ for δ¹⁸O. Carbonate data are expressed against the V-PDB standard for both oxygen and carbon isotopes (Tables 1 and 2).

Water sample measurements were made from the CO₂ in the headspace equilibrated with a 0.5 mL aliquot in Exetainer tubes using Gas Bench II coupled to a Delta V Advantage mass spectrometer. Reproducibility of replicated standards was usually better than 0.2‰ for δ¹⁸Osw. The oxygen isotope composition of water samples is expressed against V-SMOW standard (Table 1).

As δ¹³C of the DIC (predominantly HCO₃⁻) changes with temperature [Mook et al., 1974], we assessed the carbon isotope composition of each medium for a given temperature using the equation from Lynch-Stieglitz et al. [1995], and assuming a δ¹³C of the DIC at 0% at 15°C [Hermoso et al., 2014]. Data are given in Table 2.

### Figure 2...

![SEM image and smear-slide micrograph of a sedimentary *T. heimi*i assemblage obtained from core top sample MD82-449 (Southern Atlantic Ocean). This near-monospecific *T. heimi*i assemblage was obtained in the 10–20 μm fraction by the microfiltering protocol described in Minoletti et al. [2009].](image)

![SEM image and smear-slide micrograph of a sedimentary *T. heimi*i assemblage obtained from core top sample MD82-449 (Southern Atlantic Ocean). This near-monospecific *T. heimi*i assemblage was obtained in the 10–20 μm fraction by the microfiltering protocol described in Minoletti et al. [2009].](image)

### Table 2. Temperature, Specific Growth Rate, and Isotopic Data From the Culture Experiment

<table>
<thead>
<tr>
<th>Code</th>
<th>Temperature (°C)</th>
<th>μ (day⁻¹)</th>
<th>δ¹⁸OT. heimi (%) V-PDB</th>
<th>δ¹³CT. heimi (%) V-PDB</th>
<th>δ¹⁸Omedium (%) V-SMOW</th>
<th>δ¹³CDIC (%) V-PDB</th>
</tr>
</thead>
<tbody>
<tr>
<td>13A</td>
<td>13</td>
<td>0.23</td>
<td>-0.25</td>
<td>-6.36</td>
<td>0.67</td>
<td>-1.20</td>
</tr>
<tr>
<td>13B</td>
<td>13</td>
<td>0.24</td>
<td>-0.01</td>
<td>-6.03</td>
<td>0.67</td>
<td>-1.20</td>
</tr>
<tr>
<td>15A</td>
<td>15</td>
<td>0.26</td>
<td>-0.16</td>
<td>-6.62</td>
<td>0.69</td>
<td>-1.38</td>
</tr>
<tr>
<td>15B</td>
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<td>0.69</td>
<td>-1.38</td>
</tr>
<tr>
<td>17A</td>
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<td>-5.79</td>
<td>0.65</td>
<td>-1.56</td>
</tr>
<tr>
<td>17B</td>
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<td>-6.14</td>
<td>0.65</td>
<td>-1.56</td>
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<tr>
<td>20A</td>
<td>20</td>
<td>0.38</td>
<td>-1.24</td>
<td>-4.23</td>
<td>0.69</td>
<td>-1.84</td>
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<tr>
<td>20B</td>
<td>20</td>
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<td>-1.84</td>
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<tr>
<td>23A</td>
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<tr>
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<tr>
<td>25A</td>
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<td>-3.92</td>
<td>0.68</td>
<td>-2.30</td>
</tr>
<tr>
<td>25B</td>
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<td>-2.30</td>
</tr>
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<td>27A</td>
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<td>-2.48</td>
</tr>
<tr>
<td>27B</td>
<td>27</td>
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<td>-1.94</td>
<td>-3.83</td>
<td>0.67</td>
<td>-2.48</td>
</tr>
</tbody>
</table>

*δ*Isotope compositions of culture residues and δ¹⁸O of the culturing medium were measured for all temperatures. Carbon isotope compositions of the DIC for each temperature derived from the measurement at 15°C with δ¹³C of DIC = 0‰ [Hermoso et al., 2014]. For each other temperature, we used the equation given by Lynch-Stieglitz et al. [1995] to account for the temperature dependence on δ¹³C of DIC.

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2.4. Expression of Results

In the present study, the composition of T. heimii calcite is expressed for both isotope systems by using the "Δ - δ" notation, i.e., 
\[ \delta^{18}O_{\text{c}} - \delta^{18}O_{\text{sw}} \text{ (V-PDB - V-SMOW)} \] and 
\[ \delta^{13}C_{\text{c}} - \delta^{13}C_{\text{DIC}} \text{ (V-PDB - V-PDB).} \]

Two isotopic values do not correspond to fractionation coefficients but represent convenient estimates routinely used in paleoceanography to normalize microfossil geochemistry and allows comparison between regions with different seawater chemistries. Comparison with an inorganic reference is achieved via the offset of T. heimii calcite with the equilibrium calcite calculated after the equation of Kim and O'Neil [1997] approximated to quadratic expression by Bemis et al. [1998] for oxygen and after the equation of Romanek et al. [1992] for carbon.

\[ \delta^{18}O_{\text{c}} \text{ [V-SMOW]} - \delta^{18}O_{\text{sw}} \text{ [V-SMOW]} = - 0.211 \times T[°C] + 3.410 - 0.568 - 0.27 \]  

(1)

3. Results

3.1. Algal Growth and Medium Drift

Specific growth rate (μ) increased from 13 to 23°C, with an exponential trend from 0.24 to 0.52 day⁻¹ \( (r^2 = 0.97) \) (Figure 3). Growth rates diminished at 25 and 27°C, possibly indicating an ecological optimum at 24°C for this strain and applied experimental setup (light and nutrient levels). Final cell concentrations at the end of the experiments were between 4000 and 16,000 cells per mL. Even the highest cell density corresponded to very little utilization of the DIC resource in the culture medium. A conservative estimate of 5% DIC drawdown is obtained by considering ~2000 pg C per cell [Young and Ziveri, 2000]. This represents significantly lower yields compared to the end of the culture study of Zonneveld et al. [2007]. In our experiments, the typical drift in pH measured at the end of the experiments was less than 0.1 unit owing to our very dilute cell concentration strategy. This stability of the growth medium testifies for the stable composition of the carbonate system that the cells were subjected to throughout the experiments, with similar conditions at the beginning and the end of the experiments.

3.2. Oxygen Isotope Values of Culture Residues

Growth temperature strongly influences the oxygen isotope composition of T. heimii calcite. Over the 13 to 27°C interval, the range in \( \delta^{18}O_{\text{c}} \) was ~2.2‰ with isotopic ratios linearly decreasing with temperature, as expected from equilibrium prediction (Figure 4). The replicates were in good agreement with a mean standard deviation on the order of 0.10‰, indicating that obtained magnitudes of fractionation estimated from \( \delta_{\text{c}} - \delta_{\text{sw}} \) values are reproducible and show a very strong linear relationship with temperature \( (r^2 = 0.91) \). The equation linking oxygen isotope composition and temperature is:

\[ \delta^{18}O_{\text{c}} \text{ [V-PDB]} - \delta^{18}O_{\text{sw}} \text{ [V-SMOW]} = - 0.17 \times T[°C] + 1.52 \ (r^2 = 0.91) \]  

(2)
The root mean square error (RMSE) of the distribution (n = 14) indicates a residual of about 1.6°C indicating the good accuracy of the culture calibration. A 2‰ offset was recorded between the early and late exponential phase data points at 15°C (Figure 4). Taken at face value, this offset would represent an unrealistic ~12°C discrepancy. This result illustrates that performing very dilute cultures in batch is a prerequisite for measuring meaningful δ¹⁸O fractionation factors. It is worth noting that the oxygen isotope measurements obtained for temperatures lower than 17°C depart from the inorganic reference. Disregarding the two lowest temperatures, also corresponding to minimum growth rate of *T. heimii*, our culture calibration is parallel and indistinguishable from the inorganic reference (Figure 4):

\[
\delta^{18}O_{c} \text{ [%o V-PDB]} - \delta^{18}O_{sw} \text{ [%o V-SMOW]} = - 0.19 \times T \text{[°C]} + 2.09 \quad (r^2 = 0.86)
\]

The RMSE corresponds to the “internal” reliability of the calibration, in the sense that it is based on the distribution of measurements and their degree of scattering with respect to the linear regression line. The same approach can be used to assess an “external” uncertainty of δ¹⁸O composition in *T. heimii* compared to the inorganic calcite (equation (1)). The RMSE using predicted values that derive from equation (1) is 3.0°C (n = 14). Not surprisingly, the lowest temperatures (13 and 15°C) correspond to maximal residuum of the population (Figure 4). Disregarding these coldest temperatures, RMSE in the range of 17–27°C is significantly reduced and quantified to 2.3°C (n = 10). These contrasting vital effects for oxygen isotopes on both sides of the 17°C cutoff temperature challenge the univocal assignment of *T. heimii* to an isotopically “light group” [Zonneveld et al., 2007; Ziveri et al., 2012; Van de Waal et al., 2013]. However, our data support a relatively light isotopic composition at 15°C, as previously documented at this temperature [Hermoso et al., 2014].

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**Figure 4.** Oxygen isotope composition of *T. heimii* calcite (expressed as δ¹⁸O₅₅[V-PDB] – δ¹⁸O₅₅[V-SMOW]) versus temperature in laboratory culture experiments. Data from dilute bioassays are represented by red circles. High cell density 15°C batch cultures (2‰ isotopically lighter than the very dilute 15°C assay) are indicated by red crosses. Overall, the data show a very good linear regression coefficient (r² = 0.91). The equilibrium calcite line was calculated after *Kim and O’Neil* [1997] and adjusted for pH 8.2 (see equation (1) in text). At temperatures greater than 15°C, the agreement with the equilibrium reference is remarkably good, while at lower temperatures, δ¹⁸O values are offset by about 1‰ toward negative values. The culture regression of *Zonneveld et al.* [2007] is indicated with the dashed line; other culture data sets are indicated with black stars [Ziveri et al., 2012] and black triangles [Van de Waal et al., 2013]. Numbers next to symbols correspond to the carbonate ion concentrations in the medium expressed in μmol kg⁻¹ sw. For comparison, our cultures were grown at [CO₃²⁻] of ~250 μmol kg⁻¹ sw. For comparison, our cultures were grown at [CO₃²⁻] of ~250 μmol kg⁻¹ sw.
3.3. Carbon Isotope Values of Culture Residues

The $\delta^{13}C_{T. heimii}$ values of $T. heimii$ are linearly correlated with temperature ($r^2 = 0.92$) (Figure 5). Because temperature and $\delta^{18}O_c$ are also correlated, there was a strong relationship between carbon and oxygen values ($r^2 = 0.86$). We measured very negative $\delta^{13}C_c$ for $T. heimii$ shells that fall between the composition of equilibrium calcite (calculated after Romanek et al. [1992]) and that of aqueous CO2 assumed to be the substrate assimilated by the cell (around $\sim 10.2\%$o) calculated after Lynch-Stieglitz et al. [1995].

3.4. Oxygen Isotope Values of Microseparated $T. heimii$ Fractions

For $T. heimii$ fractions, $\delta^{18}O_c - \delta^{18}O_{sw}$ values range from $-0.43\%o$ at 15.2°C for North Atlantic site SU90-08 to $-2.66\%o$ at 25.6°C for site MD104-24 (Offshore Somalia) and are negatively correlated to temperature (Figure 6). The equation linking oxygen isotope fractionation to temperature in the core top approach is:

$$\delta^{18}O_c [\%o \text{ V-SMOW}] = -0.19 	imes T[°C] + 2.16 \quad (r^2 = 0.82)$$

Core top fractions are in very good agreement with culture data over the entire investigated temperature range, i.e., including the lowest temperature at 15.2°C. When plotted on the same graph, the samples from Zonneveld [2004] that where of comparable purity to the present study (>90%) confirm this regression line while less pure $T. heimii$ fractions exhibit scattered and overall, more positive $\delta^{18}O_c$ values (Figure 6).

3.5. Carbon Isotope Values of Microseparated $T. heimii$ Fractions

The $\delta^{13}C_c$ values of $T. heimii$ fractions exhibit relatively light values, ranging from $-1.31\%o$ to $-3.67\%o$ (Table 1). This observation is in agreement with data from cultured $T. heimii$ shells (Figure 5) and with evidence from the geological record [Friedrich and Meier, 2003; Kohn et al., 2011]. As relatively easily done for the oxygen isotope system (see section 2.2.2.), retrieving oceanic $\delta^{13}C$ of DIC relevant for $T. heimii$ found in core tops has been found to be challenging. First, the coverage and resolution of available databases for
Present-day carbon isotope of DIC in seawater are rather scarce [Gruber et al., 1999]. Second, *T. heimii* shells laid down on the sea floor were produced before the infusion of isotopically light carbon into the surface waters (as referred to as the Suess effect). Last, *T. heimii* thrives in the intermediate photic zone layer while available data correspond to surface waters. As a result, data from present-day waters do not accurately account for $\delta^{13}C_{DIC}$ values to which *T. heimii* from our microseparated fractions were exposed to. These methodological and conceptual issues prevent us to express carbon isotope fractionation using the "$\delta^{13}C_{c}/C_{0}\delta^{13}C_{DIC}$" notation. Consequently, taking $T. heimii$ $\delta^{13}C_{c}$ as a proxy for carbon isotope fractionation in core tops can only be ascertained if all aforementioned parameters are not themselves temperature dependent.

Estimates of paleo-$\delta^{13}C_{DIC}$ that would reflect the composition of $T. heimii$‘s depth habitat could be made using the carbon isotope signature of the foraminifera *Glabarotalia menardii* or *Neoglobquadrina dutertrei* extracted from the same core top sediments. Species-specific coefficient allowing conversion of foraminiferal $\delta^{13}C_{c}$ into $\delta^{13}C_{DIC}$ as determined by Spero et al. [2003] might be applied to constrain $\delta^{13}C_{c}$ of the intermediate photic zone. Unfortunately, we did not have access to the coarse fraction of our core top sediments, nor found previously-published $\delta^{13}C$ of foraminifera for these sites. With all these problems in mind, we refrained from interpreting sedimentary $\delta^{13}C_{c}$ values of *T. heimii*. Nevertheless, raw isotopic data are presented in Table 1.

**4. Discussion**

**4.1. Oxygen Isotope Composition: “Light” Versus “Equilibrium” Group?**

Our culture data indicate that *T. heimii* precipitates calcite in near-equilibrium conditions (Figure 4). The main source of DIC assimilation in *T. heimii* has been demonstrated to be CO$_2$ [van de Waal et al., 2013]. The fact
that $\delta^{18}O_C$ was not shifted toward more positive values indicates that the very heavy oxygen isotope composition of CO$_2$ ($\sim +7\%$O with respect to HCO$_3^-$ [Zeebe, 1999]) is not preserved in calcite produced by T. heimii. This implies total re-equilibration of the oxygen isotope system (H$_2$O/CO$_2$HCO$_3^-$/CO$_3^{2-}$). In solution, total re-equilibration is attained within approximately 12 h at 15°C [Zeebe and Wolf-Gladrow, 2001].

The oxygen isotope system of fast growing coccolithophores such as E. huxleyi is not completely re-equilibrated at time of calcification and their calcite therefore reflects to some extent the isotopic composition of the isotopically heavy CO$_2$ source [Hermoso et al., 2014]. This thermodynamic feature typifies the so-called “heavy group” [Dudley et al., 1986]. The near equilibrium $\delta^{18}O_C$ of T. heimii might be explained by the proven presence of carbonic anhydrase inside the cell and in the boundary layer [Van de Waal et al., 2013], which speeds up the re-equilibration between DIC and H$_2$O [Uchikawa and Zeebe, 2012]. Additionally, a long residence time of the intracellular DIC pool between assimilation and calcification could account for equilibrium composition, as for the coccolithophore Coccolithus pelagicus [Hermoso et al., 2014], which, like T. heimii, has a relatively slow growth rate.

At low temperatures (13 and 15°C), offsets from equilibrium of about $-1\%$O were recorded in T. heimii calcite, as observed by Hermoso et al. [2014]. Over this temperature range, T. heimii can hence be assigned to the isotopically “light group”, like the coccolithophore Calcidiscus leptoporus. The change in oxygen isotope values at the lowest temperatures corresponds to the ecological lower limit for this species in terms of temperature tolerance (Figure 3). Lower growth rate would leave more time for thermodynamic re-equilibration of the DIC with H$_2$O, but this process is also slower at low temperatures, so an explanation based on kinetics of the oxygen isotope system can be ruled out.

From a chemical point of view, $\delta^{18}O_C$ lower than equilibrium can result from a higher relative proportion of CO$_3^{2-}$ compared to HCO$_3^-$ at the site of calcification (relative to the ratio of the external fluid) or from kinetic effect [Watkins et al., 2014]. This change in the DIC system would be accompanied by increased pH. We hypothesize that a less active carbon concentrating mechanism and/or less intense calcification at low temperature could account for higher pH. With a lowered net CO$_2$ flux, as evidenced by limited growth rate at low temperature, calcification would occur in a less acidic environment. Under this assumption, calcite grown at lower temperatures would reflect the higher pH and concomitant higher abundance of CO$_3^{2-}$ with respect to HCO$_3^-$ and therefore would have lower oxygen isotope values. An alternative hypothesis is that the re-equilibration time is not achieved completely below 17°C due to the reduced activity of carbonic anhydrase at low temperature. Further biogeochemical work is required to test these hypotheses, especially to characterize passive diffusion of CO$_2$ versus HCO$_3^-$ uptake in T. heimii and more broadly the carbon concentrating mechanisms (or CCMs) under a range of temperatures.

4.2. Comparison With Previous Culture Studies

4.2.1. Implications of the Culture Technique for the Magnitude of the Vital Effect

The culture strategy (dilute versus dense) has a strong effect on the isotopic composition of T. heimii calcite. This point is well illustrated by the batch cultures at 15°C that reached the late exponential phase with cell division being hampered by unfavorable living conditions. Low $\delta^{18}O_C$ are unlikely to be the consequence of changed $\delta^{18}O$ of the medium during growth, as initial and final measurements were within analytical error. Rather, we explain this drift by an altered chemistry of the medium with substantial elevation of pH due to the consumption of CO$_2$. This “culture artifact” may hence correspond to the combination of reduced [CO$_2$]$_{aq}$ and high pH, at which enzymatic activity, including those involved in the CCMs, is ineffective. Other environmental parameters could be involved in such large shifts toward negative $\delta^{18}O_C$ values, including the effect of pH [Zeebe, 1999, 2001] and potentially the influence of reduced light and nutrient levels.

4.2.2. The Pioneering Work of Zonneveld et al. [2007]

Zonneveld et al. [2007] performed an extensive culture campaign investigating $^{18}O$ fractionation in T. heimii under a wide range of temperatures (Figure 1). They documented the evolution of carbonate chemistry during growth, which can be used to better interpret the oxygen isotope compositions of T. heimii calcispheres produced in the laboratory. The authors calculated a linear equation for $^{18}O$ fractionation according to temperature from their culture examination of two strains of T. heimii. They grew T. heimii using an initial pH of 8.0, which is 0.2 pH unit higher than in our experimental condition.

In the culture study of Zonneveld et al. [2007], the pH drifts were significant, by +0.4 pH unit. This drift is similar to that we observed in our late exponential assays. As cell density and hence calcification, follow an
From these perturbation experiments, it is thus difficult to establish a direct link between CO$_3^{2-}$ concentrations, pH, and $\delta^{18}$O change with temperature, as more than one parameter was modified. Our experiments were not intended to examine the effect of the status of the carbonate system on $\delta^{18}$O, but rather to investigate the temperature record in conditions as close as possible to the natural environment, i.e., keeping all other parameters constant by using very dilute batch cultures.

4.3. Reconciled Culture and Core Top Data

4.3.1. Temperature and $\delta^{18}$O of Dinoflagellate Calcite

There is a good concordance between culture and core top data in our study (Figures 4 and 6). In both data sets, strong linear correlation coefficients demonstrate the foremost control of temperature on oxygen isotope composition in T. heinii calcite. This concordance in the slopes and intersects of the equations...
(equations (2)–(4)) means that our calibration is transferable to natural environments and that δ18O of *T. heimii* calcite may be used as a robust paleothermometer for assessing seawater temperatures of the intermediate photic zone.

From our culture and core top studies, it is clear that calcification of *T. heimii* is made in near isotopic equilibrium with ambient seawater for the oxygen isotopes. This absence of apparent vital effect potentially indicates a relatively low degree of biological control on calcification in *T. heimii* for this isotopic system. This geochemical observation may be related to the suggested partially external mode of biomineralization in *T. heimii* [Tangen et al., 1982; Inouye and Pienaar, 1983; Gussone et al., 2010].

Below 17°C, δ18O values of the calcite of cultured *T. heimii* are slightly offset from equilibrium (~ −1‰), but this departure was not recorded in core top data. Indeed, at 15.2°C, which is the mean annual temperature of site SU90-08, *T. heimii* exhibited equilibrium δ18O values. This slight discrepancy would suggest that the calcite of cultured *T. heimii* is isotopically shifted due to a process that is not operating in the natural environment. Further biogeochemical work would be useful to confirm the existence of this differential vital effect and to understand its origin. Examining core tops where calcification took place in temperatures colder than 15°C would also help reconcile datasets and confirm whether *T. heimii* belongs to the isotopically “light group” at low temperature. Nevertheless, *T. heimii* mainly thrives in low-latitude waters characterized by temperature corresponding to the range where its δ18O composition is not offset [Karwath et al., 2000; Kohn and Zonneveld, 2010]. Considering the long-term cooling of the Cenozoic, it could be inferred that temperatures of *T. heimii* habitat in the pelagic environment have never been lower than 15°C. As such, the possible limitation of our calibration may have no effect on the applicability of paleotemperature reconstruction with a geological perspective.

In their attempt to reconstruct seawater temperatures over the last 45 ka, Kohn et al. [2011] measured δ18Oc of *T. heimii* and two planktic foraminifera, *Globigerinoides ruber* and *Globigerina bulloides*, from the tropical North Atlantic Ocean. Oxygen isotope values of *T. heimii* shells from the Late Holocene fell between −0.5‰ and −1‰. The authors calculated seawater temperatures using the equation of Zonneveld [2004] (Figure 1) and found unrealistically low temperatures. As an alternative, they assumed that calcite precipitation of calcareous dinoflagellates occurred in isotopic equilibrium with ambient water, using the raw equation of Kim and O’Neil [1997]. The results appeared to be more consistent with what was expected for the thermal status of the upper water column of this interval, especially when compared to foraminiferal data.

Yet in the study by Kohn et al. [2011], the δ18O of *T. heimii* were measured −0.5‰ heavier than *G. ruber* (var. pink), a symbiotic foraminifera that spends its entire life cycle in the mixed layer [Deuser and Ross, 1989; Farmer et al., 2007; Pearson, 2012]. Reconstructing similar temperatures from *T. heimii* and *G. ruber* is still a contentious result, as these organisms calcify in distinct water masses. Instead, the recorded oxygen isotope offset between both taxa that calcify in equilibrium with seawater supports relatively deep habitat for *T. heimii*. At the sites of Kohn et al. [2011], the present-day temperature difference between mixed layer and intermediate photic zone waters is ~3°C. Applying our calibration to *T. heimii* δ18O would reconcile the data. Indeed, this alternative correction of δ18O values in *T. heimii* would lead to a decrease of ~2°C in isotopic temperature estimates relative to *G. ruber* (var. pink). This figure would make *T. heimii*-derived temperatures compatible with a calcification depth deeper than *G. ruber*. Overall, this downcore study case of *T. heimii* δ18O confirms the validity of our calibration for the geological record.

### 4.3.2. A New Proxy for Thermal Gradients Within the Photic Zone

Recent methodological advances, like that used in the present study, now allow microseparation of coccoliths and calcareous dinoflagellate shells in sediments. These sedimentological techniques enable near-monospecific geochemical measurements on these nanofossils and hence examination of the vital effect in the geological record [e.g., Bolton and Stoll, 2013]. This effort contributes to the ongoing development of new proxies for evaluating accurate temperatures of the uppermost water column. Coccoliths and some surface-dwelling foraminifera species, such as *G. ruber*, are ideal for deriving the temperature of the mixed layer, corresponding to SSTs. In addition, there is a promising opportunity to decipher parameters other than temperature by undertaking cross-species comparison of δ18O of coccolith taxa [Rickaby et al., 2010; Hermoso et al., 2014]. Due to its deeper habitat, the calibration of *T. heimii* geochemistry is of primary importance since it enables characterization of the intermediate part of the photic zone, i.e., the upper part of the thermocline. Hence, measuring paired coccolith and dinoflagellate calcite isotopes might be used to
constrain the thermal structure of the photic zone. Comparison of isotopic signals in calcite of the coccolithophore Coccolithus pelagis before its poleward migration [Bown, 1998; Sato et al., 2004] with those of T. heimi would be of particular help for reconstructing thermal gradients within the photic zone and for capturing long-term variability in the thermocline depth. The Δ18Osw variability within the photic zone is small and usually lower than analytical errors of isotope measurements (±0.1‰); hence, differences in isotopic signals may reflect a temperature signal, after a suitable correction accounting for the vital effect is achieved. 4.3.3. A Temperature Component Driving δ13C of T. heimi Calcispheres There is a positive correlation between carbon isotope composition of T. heimi and temperature in our batch cultures (Figure 5). In our cultures, δ13C of aqueous CO2 was constant around −10.2‰, whereas the δ13C of DIC, predominantly consisting of HCO3−, and hence that of inorganic calcite slightly decreases (Figure 5) [Romanek et al., 1992; Lynch-Stieglitz et al., 1995]. Under the safe assumption that T. heimi assimilates CO2 [van de Waal et al., 2013], variation of carbon isotope composition in calcite cannot be explained by the decrease in the isotopic composition of the carbon substrate assimilated by the cell and subsequently used for calcification. Increase in δ13C of T. heimi may not correspond to a change in the magnitude of 12C/13C fractionation. With elevation in temperature and accompanying higher carbon fixation in organic matter [van de Waal et al., 2013], we hypothesize that the DIC pool used for calcification becomes heavier due to more intense Rayleigh fractionation linked to Rubisco fractionation that favors 12C-carbon. Hence, we hypothesize that the linear relationship between temperature and δ13C observed in our cultures may also be driven by a modulation of the Rayleigh fractionation of the carbon isotope system with higher 12C partitioning into organic matter at high temperature [van de Waal et al., 2013]. This biogeochemical feature represents a commonality with the carbon isotope system in symbiobearing foraminifera [Bernis et al., 2000]. Overall, covariation in δ18O/δ13C signals in the geological record should not always be regarded as a diageneric feature [Marshall, 1992], as this work demonstrates that they may convey a pristine biogeochemical signal. Further investigation of the coupling between the two isotopic systems (oxygen and carbon) should be conducted. Notably, a full mechanistic understanding of stable isotope fractionation in calcite-producing dinoflagellates would also require separate analyses of the internal versus external calcite layers. It would be particularly interesting to determine how the inner layer that is precipitated intracellularly (hence presumably with higher coupling between the photosynthetic and calcification carbon pools) compares isotopically with the outer layer that is supposed to be under a lesser degree of biological control.

5. Conclusions
This study represents a reconciled culture and core top calibration of the temperature record on δ18O of the calcareous dinoflagellate T. heimi. Previous studies have produced conflicting data between culture results and the natural environment. The combination of two methodologies implemented here, very dilute batch cultures and microseparation of extremely enriched (>90%) T. heimi assemblages from core top sediments, enabled this successful calibration. Differences with previous studies may originate from a drift in the composition of the growth medium and insufficiently purified sedimentary material, respectively.

Our work demonstrates the near-equilibrium composition of T. heimi calcite for oxygen isotopes. In our culture data, however, there seems to be a mismatch with lower (1‰) δ18O with respect to equilibrium below 17°C, which is not seen in core top data. The cause of this discrepancy remains elusive, although reduced DIC assimilation and consequently reduced growth rates may explain this change of the vital effect with higher pH values at the site of calcification. As the occurrence of this species at such low temperatures is rare, this discrepancy does not represent a pitfall for our calibration, although this offset should be kept in mind if T. heimi are extracted from temperate regions. Downcore investigation of stable isotopes from T. heimi microfractions represents a promising means to reconstruct the evolution of temperature of the intermediate photic zone. When combined with the shallower dwelling coccolithophores, interspecies δ18O may help unraveling the thermal structure of the mixed layer.

The very negative δ13C values in calcite found in this study confirm previous culture and geological evidence. This geochemical feature contrasts with observations made in coccolithophores in which a relatively high degree of coupling between calcification and calcification lead to more positive δ13C. Furthermore, as both carbon and oxygen isotopes seem to be under a common temperature control, this may help understanding and interpreting covariation between δ13C and δ18O in the geological record.
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