Comparison of species-specific oxygen isotope paleotemperature equations: Sensitivity analysis using planktonic foraminifera from the Cariaco Basin, Venezuela

Katherine E. Wejnert a,⁎, Robert C. Thunell a,b, Yrene Astor c

⁎ Corresponding author at: School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, GA 30332-0340, USA. Tel.: +1 404 894 3942.
E-mail addresses: kwejnert@geol.sc.edu (K.E. Wejnert), thunell@geol.sc.edu (R.C. Thunell).

A R T I C L E  I N F O

Article history:
Received 6 December 2011
Received in revised form 4 March 2013
Accepted 17 March 2013

Keywords:
Cariaco Basin
Planktonic foraminifera
Oxygen isotopes
Paleotemperature equations

A B S T R A C T

The oxygen isotope compositions of seven species of planktonic foraminifera (Globigerinoides ruber (pink), Globigerina bulloides, Globorotalia menardii, Globigerinoides sacculifer, Orbainia universa, Globorotalia crassaformis, and Neogloboquadrina dutertrei) were measured from biweekly sediment trap samples collected in Cariaco Basin, Venezuela between 2003 and 2008. Sixteen published paleotemperature equations were used to estimate calcification temperatures from this δ18O data. As part of the CARIACO time series, routine in situ water column chemistry measurements and hydrographic monitoring has occurred since 1995. The calcification temperature estimates were compared with these monthly water column temperature measurements to ascertain species depth habitats and to determine which equations yielded the most realistic temperature estimates. No single equation provided accurate temperature estimates for all species. Rather, the equations yielded temperature estimates that varied by 2–4 °C for each species. Overall species-specific paleotemperature equations based on plankton tow samples yield the coldest temperature estimates, while generic equations produce the warmest temperatures. Our results indicate that when using planktonic foraminiferal δ18O to make paleotemperature estimates, equations must be carefully selected on a species by species basis. Here we provide general guidelines for selecting paleotemperature equations. We also evaluated how well the δ18O difference between the surface dwelling G. ruber and the deeper dwelling N. dutertrei estimates the surface to thermocline temperature gradient. The Δδ18O is small (0.61‰) during winter due to upwelling and increases (1.37‰) during summer as upwelling ceases and surface waters become thermally stratified.

© 2013 Elsevier B.V. All rights reserved.

1. Introduction

Urey (1947) first proposed that the temperature of the ambient environment controls the oxygen isotopic composition of carbonates. McCrea (1950) and Epstein et al. (1953) verified the theoretical work of Urey (1947) empirically using inorganically precipitated calcite and mollusk shells. Emiliani (1955) first applied this concept to foraminiferal δ18O assuming a predominately temperature signal, with Shackleton (1967) subsequently demonstrating that a large part of the foraminiferal δ18O signal is related to past changes in the δ18O of seawater (δ18Osw), which were larger than previously thought. Since then, the oxygen isotope composition of planktonic foraminifera has become one of the most commonly used paleothermometers and is frequently used in conjunction with Mg/Ca measurements to separate the temperature and δ18Osw signals (see Lea, 2003; Ravelo and Hillaire-Marcel, 2007; Katz et al., 2010 for reviews).

Since these pioneering studies, many empirically derived paleotemperature equations have been developed on both inorganically precipitated carbonates and individual species of foraminifera (Erez and Luz, 1983; Bouvier-Soumagnac and Duplessy, 1985; Kim and O’Neil, 1997; Bomis et al., 1998; Mulitza et al., 2003). Most equations were developed using cultured samples, while others are based on plankton tow samples or living samples pumped from surface waters. While some studies have used a single generic equation to estimate paleotemperatures from the δ18O of multiple species (e.g. Hendy and Kennett, 1999; Ganssen and Kroon, 2000), the recent trend has been to use species-specific equations to reconstruct past ocean temperatures (Bemis et al., 1998, 2002). Although most species-specific paleotemperature equations have basic similarities, they can produce significantly different temperature estimates when applied to the same isotope datasets (Bemis et al., 2002; King and Howard, 2005; Saraswat and Khare, 2010). In addition, the relationship between temperature and δ18O can vary between species and with changing environmental conditions (Spero et al., 1997; Bemis et al., 1998; Bijma et al., 1999; Zeebe, 1999; Bemis et al., 2002; King and Howard, 2005), thus it is important to test paleotemperature equations in a range of natural environments in order to select an appropriate equation for estimating temperatures from the sedimentary record.
In this study, we present δ¹⁸O data from seven species of planktonic foraminifera from sediment trap samples collected over a five-year period (May 2003–May 2008) in Cariaco Basin, Venezuela and build on the earlier work of Tedesco et al. (2007). Both generic and species-specific equations are used to estimate calcification temperatures, which are then used to ascertain the depth habitats of different species and to assess the performance of each paleotemperature equation. We also test how well δ¹⁸O differences between surface and thermocline species can be used to reconstruct upper ocean thermal stratification.

2. Regional setting

Cariaco Basin is a large anoxic basin (~10⁶ km²) located along the northern continental margin of Venezuela (Fig. 1). The basin is approximately 1400 m deep and is divided into two sub-basins separated by a saddle at 900 m water depth. The basin is isolated from the Caribbean Sea by a shallow sill (~150 m) that restricts the exchange of deeper water. Vertical mixing is also restricted due to a strong pycnocline below the mixed layer (Scranton et al., 1987). The restricted mixing coupled with high productivity results in anoxia below ~250 m (Deuser, 1973; Richards, 1975).

The two primary watermasses entering the Cariaco Basin are the North Equatorial Current (NEC) and the Subtropical Underwater (SUW). The principal source of surface water in the basin is the NEC, which enters the Caribbean Sea from the east through the passages between the islands of the Lesser Antilles and continues to flow westward as the Caribbean Current (Hernández-Guerra and Joyce, 2000; Johns et al., 2002). Beneath the surface waters is the SUW, which originates in the central tropical Atlantic and can reach depths shallower than 50 m in Cariaco Basin during the winter/spring upwelling season (Astor et al., 2003).

Cariaco Basin is located within the trade wind belt and as a result, its climatology is driven by the seasonal migration of the Inter-Tropical Convergence Zone (ITCZ). During the winter and early spring the ITCZ is in its southernmost position and the Cariaco Basin experiences strong easterly winds, which cause coastal upwelling, minimum sea surface temperatures (SST), maximum nutrient concentrations, and increased primary production. During the summer and early fall the ITCZ migrates northward, causing the easterly winds to diminish, upwelling to cease, an increase in SST, and a decrease in nutrient concentrations and primary production (Goñi et al., 2003; Tedesco and Thunell, 2003; Scranton et al., 2006). A brief secondary upwelling event also frequently occurs during late summer (Muller-Karger et al., 2004; Astor et al., 2005). Typically precipitation is highest during the summer and fall when the ITCZ is in its northernmost position and upwelling is weakest (Tedesco and Thunell, 2003). Since 1995, hydrographic and water column data, including temperature and salinity measurements, are collected monthly in the basin. During 2003–2008 surface salinity ranged from 35.8 to 37.1 and surface temperature varied from 23.3 to 29.2 °C. During this period the salinity of the upper 200 m ranged from ~36.4 to 36.9 and temperature ranged from ~18 to 27 °C (Fig. 2).

3. Material and methods

As part of the Cariaco Basin time series study (Muller-Karger et al., 2001) a sediment trap mooring has been deployed since November 1995 in the deepest part (~1400 m) of the eastern sub-basin (10°30′N, 64°40′W; Fig. 1) (Thunell et al., 2000, 2007). The mooring includes five automated sediment traps (~150, 230, 410, 810, and 1200 m) which are programmed to collect consecutive two-week-long samples. For this study, we utilized samples from the two uppermost traps (~150 and 230 m) collected from 2003 to 2008. In addition, hydrographic and water column data are collected monthly at the sediment trap mooring site (Muller-Karger et al., 2001). For this study, we used the monthly temperature and salinity profiles, which are measured continuously throughout the water column (0–1310 m) using a Seabird (SBE-25) CTD. Time series data are available online at www.imars.usf.edu/CAR/. Periods of upwelling were determined using an upwelling index developed by Astor et al. (2013). Seasonal upwelling is directly related to the intensity of the trade winds (Muller-Karger et al., 2004), thus the upwelling index is calculated from zonal wind stress (Astor et al., 2013).

Prior to deployment, each sample cup is filled with a buffered formalin solution to poison and preserve the samples. After the trap samples are collected, they are split using a precision rotary splitter, stored in buffered deionized water, and refrigerated until processed. The samples were wet sieved over a 125 μm screen and seven species of planktonic foraminifera (Globigerinoides ruber (pink), Globigerina bulloides, Globorotalia menardii, Globigerinoides sacculifer, Orbulina universa, Globorotalia cassisformis, and Neogloboquadrina dutertrei) were wet-picked. Due to seasonal changes in species abundances and composition, all seven species are not present in each sample. Oxygen isotope analyses were conducted using a GV IsoPrime stable isotope ratio mass spectrometer, which is equipped with an automated carbonate preparation system. The long-term standard reproducibility is 0.07‰ for δ¹⁸O and results are reported relative to Vienna Pee Dee Belemnite (V-PDB).

Fig. 1. Map of the Cariaco Basin, Venezuela showing the location of the sediment trap at 10°30′N, 64°40′W and at 150 m water depth.
Calcification temperatures were calculated using sixteen published paleotemperature equations (Table 1) and the results are compared to monthly temperature measurements to estimate depth habitats for each planktonic foraminiferal species. Salinity in Cariaco Basin is measured monthly as part of the CARIOCO time series project and ranges from 35.8 to 37.1 at the surface and ~36.4–36.9 over the upper 200 m during our study period (Fig. 2). In order to determine the oxygen isotope composition of the seawater, we used the \( \delta^{18}O_{sw} : \) salinity relationships determined by McConnell et al. (2009) for upwelling (Eq. (17)) and non-upwelling conditions (Eq. (18)).

\[
\delta^{18}O_{sw} = 0.80(\pm 0.08) \times \text{salinity} - 28.53(\pm 3.0) \quad (17)
\]

\[
\delta^{18}O_{sw} = 0.27(\pm 0.04) \times \text{salinity} - 8.77(\pm 1.3) \quad (18)
\]

The uncertainty in the \( \delta^{18}O_{sw} : \) salinity equations yields an error in \( \delta^{18}O_{sw} \) of \( \pm 0.02 \) for upwelling conditions and \( \pm 0.01 \) for non-upwelling conditions (McConnell et al., 2009), which equates to errors in the temperature estimates of \( \pm 0.05 \) °C and \( \pm 0.10 \) °C respectively. The \( \delta^{18}O_{sw} \) values are scaled from SMOW to PDB by subtracting 0.20‰ when using the paleotemperature equations of Epstein et al. (1953) and Bouvier-Soumagnac and Duplessy (1985), 0.22‰ when using the Erez and Luz (1983) equation, and 0.27‰ when using the more recent paleotemperature equations (see Bemis et al., 1998).

McConnell et al. (2009) measured \( \delta^{18}O_{sw} \) in Cariaco Basin during May 1996 through February 1997 and during December 2005 through May 2006. These measurements do not cover the entire time period used in this study (2003–2008) and because there is significant annual variability in \( \delta^{18}O_{sw} \), we calculated \( \delta^{18}O_{sw} \) from the monthly salinity measurements. McConnell et al. (2009) found significant differences in the \( \delta^{18}O_{sw} : \) salinity relationship between upwelling and non-upwelling periods, which is caused, in part, by seasonal changes in riverine input into the basin. During the rainy season, which corresponds to periods of non-upwelling, there is an increase in discharge into Cariaco Basin from the local rivers, which include the Tuy, Unare, Neverí, and the Manzanares Rivers (Lorenzoni, 2005; McConnell et al., 2009; Martinez et al., 2010). The
input of freshwater, which has lower salinity and δ18O, into the basin significantly alters the δ18O, salinity relationship and therefore it is necessary to use different equations for upwelling versus non-upwelling conditions (McConnell et al., 2009). McConnell et al. (2009) determined δ18O, salinity relationships for both the surface and the upper 25 m. In this study, we used the surface δ18O, salinity relationships, which were recommended by McConnell et al. (2009). There is uncertainty in the temporal and spatial variabilities in the δ18O,salinity relationship. Previous studies in other locations have found that surface and intermediate waters have different δ18O,salinity relationships as a result of mixing water masses (Fairbanks et al., 1982; Benway and Mix, 2004). In Cariaco Basin, δ18O, generally increases from the surface through the mixed layer and then decreases (McConnell et al., 2009). However, the equations determined by McConnell et al. (2009) for the surface and the upper 25 m are statistically indistinguishable, indicating that spatial variability in the δ18O,salinity relationship is not as important as in other locations. More significant in Cariaco Basin is the temporal variability in the δ18O,salinity relationship caused by the seasonal changes in riverine discharge into the basin (McConnell et al., 2009). Since the McConnell et al. (2009) δ18O,salinity equations for upwelling and non-upwelling conditions are consistent between years, the largest uncertainty in the δ18O,salinity relationship will be during the transitions between upwelling and non-upwelling periods. The upwelling and non-upwelling equations produce δ18O estimates for a given salinity that differ by ~0.2–0.4‰ with the difference increasing with decreasing salinity. This would equate to a temperature estimate difference of ~1–2 °C; however any affect of error in calculating δ18O from salinity will be mostly confined to the short transitional periods between upwelling and non-upwelling and will have minimal impact on the analyses presented here.

In addition, seasonal changes in the environmental conditions will affect the foraminiferal δ18O. During winter–spring, upwelling brings water with lower carbonate ion concentration to the surface, which can impact the foraminiferal δ18O (Spero et al., 1997; Zeebe, 1999). Typically the carbonate ion concentration in Cariaco Basin during upwelling is ~40–50 μmol/kg lower than during non-upwelling periods, which based on the relationships determined by Spero et al. (1997), would relate to ~0.1–0.2% change in foraminiferal δ18O. However, the exact effect is difficult to quantify because the impact of carbonate ion on δ18O is species-specific and δ18O,[CO32−] relationships have only been determined for G. bulloides and O. universa (Spero et al., 1997; Bijma et al., 1999). In addition, symbiont activity can locally modify the carbonate chemistry of the foraminifera complicating the δ18O,[CO32−] relationship (Bijma et al., 1999).

### 4. Results and discussion

#### 4.1. δ18O data

The oxygen isotope data clearly display an ordering or grouping among the species (Fig. 3; Table 2). G. ruber and G. bulloides have the lowest values followed by G. sacculifer, with G. menardii, N. dutertrei, O. universa and G. crassaformis having the highest values. In addition, each species shows seasonal changes in δ18O that are fairly consistent from year to year with the δ18O values being higher during winter upwelling than during the more thermally stratified summer (Table 3). All seven species are present throughout the year, although typically in higher abundance during upwelling, and therefore are useful for recording seasonal changes in δ18O and temperature at their calcification depth (Tedesco and Thunell, 2003). The magnitude of the seasonal change in foraminiferal δ18O is a function of the species depth habitat with the surface dwelling species G. ruber and G. sacculifer, having average annual δ18O ranges (i.e. the difference between maximum upwelling and minimum non-upwelling δ18O) of 0.93 and 0.94‰ respectively, while the thermocline dwelling species G. menardii, N. dutertrei, and O. universa have slightly smaller average annual ranges (~0.80–0.90‰). This range of δ18O values equates to a temperature range of 5–6 °C (Tables 3 and 4), which is equivalent to the observed difference between winter and summer SST (23–29 °C) in Cariaco Basin over the study period (Fig. 2). The species with the deepest habitat, G. crassaformis, has a smaller average annual range (0.45‰%), which reflects a temperature range of approximately 3 °C. Interestingly, the surface dwelling G. bulloides has the smallest average annual range of 0.31‰, which equates to only ~2 °C. The small δ18O change relative to the annual SST range suggests that G. bulloides alters its depth habitat seasonally in order to maintain a relatively constant thermal environment, which has been suggested by previous studies (Sautter and Thunell, 1991b; Peeters et al., 2002; Tedesco et al., 2007; Wejnert et al., 2010).

G. ruber is a subtropical to tropical species (Niebler et al., 1999) that has a broad temperature tolerance of 14–32 °C (Bijma et al., 1990) and an optimum temperature preference of 27 °C (Mulitza et al., 1998). The pink variety, as used in this study, prefers warmer temperatures than the white variety (Wilke et al., 2009). Based on plankton tow samples, G. ruber is known to live in the surface mixed layer in Cariaco Basin (Miro, 1971), which is consistent with other studies (Ravelo and Fairbanks, 1992; Field, 2004; Farmer et al., 2007); however, while G. ruber mostly lives in the upper 25 m, it has been shown to migrate into deeper waters when SSTs are high and the chlorophyll maximum

---

### Table 1

Paleotemperature equations used to estimate calcification temperature from foraminiferal δ18O. Errors on the slope and intercept are 95% confidence intervals.

<table>
<thead>
<tr>
<th>Equation number</th>
<th>Reference Source</th>
<th>T (°C) = a + b(δ18Ow − δ18Oi) + c(δ18Ow − δ18Oi)²</th>
<th>Calibration temperature</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Epstein et al. (1953)</td>
<td>16.9 ± 4.2 ± 0.13</td>
<td>7–29 °C</td>
</tr>
<tr>
<td>2</td>
<td>Erez and Luz (1983)</td>
<td>17.0 ± 1.3 ± 0.74</td>
<td>14–30 °C</td>
</tr>
<tr>
<td>3</td>
<td>Bouvier-Soumagnac and Duplessy (1985)</td>
<td>16.4 ± 3.4 ± 0.57</td>
<td>20–25 °C</td>
</tr>
<tr>
<td>4</td>
<td>Bouvier-Soumagnac and Duplessy (1985)</td>
<td>15.4 ± 5.8 ± 0.93</td>
<td>20–25 °C</td>
</tr>
<tr>
<td>5</td>
<td>Bouvier-Soumagnac and Duplessy (1985)</td>
<td>14.6 ± 6.7 ± 1.09</td>
<td>22–29 °C</td>
</tr>
<tr>
<td>6</td>
<td>Bouvier-Soumagnac and Duplessy (1985)</td>
<td>10.5 ± 12.1 ± 2.64</td>
<td>24–30 °C</td>
</tr>
<tr>
<td>7</td>
<td>Bijma et al. (1999)</td>
<td>16.5 ± 0.2 ± 0.16</td>
<td>15–25 °C</td>
</tr>
<tr>
<td>8</td>
<td>Bijma et al. (1999)</td>
<td>14.9 ± 0.1 ± 0.08</td>
<td>15–25 °C</td>
</tr>
<tr>
<td>9</td>
<td>Bijma et al. (1999)</td>
<td>13.2 ± 0.3 ± 0.21</td>
<td>15–25 °C</td>
</tr>
<tr>
<td>10</td>
<td>Spero et al. (2000)</td>
<td>13.4 ± 4.48</td>
<td>9–24 °C</td>
</tr>
<tr>
<td>11</td>
<td>Spero et al. (2000)</td>
<td>15.7 ± 4.46</td>
<td>9–24 °C</td>
</tr>
<tr>
<td>12</td>
<td>Spero et al. (2003)</td>
<td>14.9 ± 5.13</td>
<td>9–24 °C</td>
</tr>
<tr>
<td>13</td>
<td>Spero et al. (2003)</td>
<td>12.0 ± 5.67</td>
<td>9–24 °C</td>
</tr>
<tr>
<td>14</td>
<td>Multizzi et al. (2003)</td>
<td>14.62 ± 0.6 ± 0.41</td>
<td>2–25 °C</td>
</tr>
<tr>
<td>15</td>
<td>Multizzi et al. (2003)</td>
<td>14.91 ± 1.4 ± 0.49</td>
<td>16–31 °C</td>
</tr>
<tr>
<td>16</td>
<td>Multizzi et al. (2003)</td>
<td>14.2 ± 0.6 ± 0.23</td>
<td>16–31 °C</td>
</tr>
</tbody>
</table>
is deep (Tedesco et al., 2007; Wilke et al., 2009). For our study period, *G. ruber* had the lowest δ¹⁸O values, ranging from −2.08 to −0.85 with an average value of −1.37. *G. ruber* δ¹⁸O values have the strongest correlation with SST, which reflects its preference for living in the surface waters (Table 2).

*G. bulloides* has a wide geographic distribution ranging from the poles to the low latitudes and characterizes upwelling zones (Prell and Curry, 1981; Sautter and Thunell, 1991a; Conan and Brummer, 2000; Tedesco and Thunell, 2003). As such, it has a wide temperature tolerance (Žarić et al., 2005), but tends to dwell closer to the surface during upwelling and then migrates to deeper waters during warmer seasons (Sautter and Thunell, 1991b; Peeters et al., 2002; Tedesco et al., 2007; Wejnert et al., 2010). Cariaco Basin has very warm temperatures (~27 °C) during the late summer and early fall and *G. bulloides*’ growth rate is known to sharply decline after 25 °C (Lombard et al., 2009); therefore it is probable that *G. bulloides* is migrating seasonally to remain near its optimum growing temperature of 20–25 °C (Žarić et al., 2005; Lombard et al., 2009). In Cariaco Basin, the δ¹⁸O of *G. bulloides* ranges from −1.98 to −0.90‰ and has a mean of −1.29‰, which is similar to *G. ruber*; however *G. bulloides* has lower seasonal variability and shows no significant relationship to SST (Table 2). This is surprising because Black et al. (2004) found a statistically significant correlation between the Hadley SST data set and down core δ¹⁸O measurements of *G. bulloides* in Cariaco Basin; however this discrepancy can be explained by differences in sampling resolution. The Black et al. (2004) study was based on year-long sediment assemblages, whereas this study uses samples collected over a two-week period. If these biweekly samples are degraded to yearly resolution, then the δ¹⁸O of *G. bulloides* has a correlation with temperature at 35 m (r² = −0.55) that is similar to that reported by Black et al. (2004) (r² = −0.61). Thus while the δ¹⁸O of *G. bulloides* accurately tracks temperature at longer time scales (i.e. yearly), it does not track short-term changes (i.e. seasonal), possibly because it is altering its depth habitat seasonally to remain near its optimum growing temperature or because of food availability (Mortyn and Charles, 2003).

*G. sacculifer* is a subtropical to tropical species that dwells in the surface mixed layer and near the chlorophyll maximum (Fairbanks et al., 1982; Niebler et al., 1999; Schmidt and Mulitza, 2002; Farmer et al., 2007). *G. sacculifer* has a similar temperature and salinity tolerance range as *G. ruber* (Bijma et al., 1990; Žarić et al., 2005; Lombard et al., 2009), but has a lower temperature optimum of 22 °C (Mulitza et al., 1998). This is consistent with our results that *G. sacculifer* has higher δ¹⁸O values than *G. ruber*, ranging from −1.66 to −0.24‰ and averaging −0.90‰ (Table 2).

*O. universa* is a tropical to subpolar species that dwells in both the mixed layer and thermocline, while *N. dutertrei* and *G. menardii* are subtropical to tropical species that primarily live at thermocline to subthermocline depths (Sautter and Thunell, 1991b; Ravelo and Fairbanks, 1992; Niebler et al., 1999; Spero et al., 2003; Field, 2004; Farmer et al., 2007; Wejnert et al., 2010). All three of these species have similar δ¹⁸O ranges and average values that are significantly higher than those of *G. ruber*, *G. bulloides*, and *G. sacculifer*, indicating that these species calcify in deeper, colder waters (Table 2).

*G. crassaformis* is a deep-dwelling species (Kemle-von Mücke and Oberhänsli, 1999; Niebler et al., 1999) that has been recorded in plankton tows from 100 to 500 m depth in Cariaco Basin (Miro, 1971) and is present year-round (Tedesco and Thunell, 2003). The δ¹⁸O for *G. crassaformis*
Table 3

<table>
<thead>
<tr>
<th>Year</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
<th>Upwelling</th>
<th>Non-upwelling</th>
</tr>
</thead>
<tbody>
<tr>
<td>2003</td>
<td>-0.87</td>
<td>-0.95</td>
<td>-1.29</td>
<td>-1.23</td>
<td>-1.17</td>
<td>-1.09</td>
<td>-1.46</td>
<td>-1.38</td>
<td>-0.77</td>
<td>-1.05</td>
<td>-1.26</td>
<td>-1.38</td>
</tr>
<tr>
<td>2004</td>
<td>-1.02</td>
<td>-0.95</td>
<td>-1.16</td>
<td>-1.23</td>
<td>-1.40</td>
<td>-1.09</td>
<td>-1.74</td>
<td>-1.38</td>
<td>-0.77</td>
<td>-1.05</td>
<td>-1.26</td>
<td>-1.38</td>
</tr>
<tr>
<td>2006</td>
<td>-1.29</td>
<td>-0.95</td>
<td>-1.17</td>
<td>-1.09</td>
<td>-1.46</td>
<td>-1.38</td>
<td>-1.52</td>
<td>-0.91</td>
<td>-0.77</td>
<td>-1.05</td>
<td>-1.26</td>
<td>-1.38</td>
</tr>
<tr>
<td>2007</td>
<td>-1.29</td>
<td>-0.95</td>
<td>-1.17</td>
<td>-1.09</td>
<td>-1.46</td>
<td>-1.38</td>
<td>-1.52</td>
<td>-0.91</td>
<td>-0.77</td>
<td>-1.05</td>
<td>-1.26</td>
<td>-1.38</td>
</tr>
<tr>
<td>Average</td>
<td>-1.29</td>
<td>-0.95</td>
<td>-1.17</td>
<td>-1.09</td>
<td>-1.46</td>
<td>-1.38</td>
<td>-1.52</td>
<td>-0.91</td>
<td>-0.77</td>
<td>-1.05</td>
<td>-1.26</td>
<td>-1.38</td>
</tr>
</tbody>
</table>

4.2. Paleotemperature equations

For each of the seven species used in this study, calcification temperatures were calculated using a selection of the sixteen equations listed in Table 1 (Table 4; Fig. 4). The calcification temperatures were then used to estimate depth habitats. The generic Epstein et al. (1953) equation was used to estimate temperatures for all seven species, while the species-specific equations were used only on the species for which they originally were calibrated and for species which previous studies had considered appropriate for that equation (e.g. Thunell et al., 1999; Faull et al., 2000; Spero et al., 2003; Tedesco et al., 2007; Weinert et al., 2010). Calcification depths were capped at 200 m for O. universa, N. dutertrei, and G. menardii and 300 m for G. crassiformis because below that the water column is so isothermal that it is difficult to differentiate depth habitats using δ¹⁸O data.

The error calculated for the temperature estimates includes the analytical error associated with the measurements of salinity and δ¹⁸O (±0.003 and ±0.07‰ respectively) and the statistical uncertainty associated with the δ¹⁸Ow: salinity and paleotemperature equations. In addition to the quantifiable sources of error, there are unquantifiable errors as well including the previously discussed seasonal [CO₂⁻] variations and the temporal and spatial variabilities in the δ¹⁸Ow: salinity relationship. Additionally symbionts can impact foraminiferal δ¹⁸O; however it is difficult to quantify the impact of symbionts on foraminiferal carbonate chemistry because it is dependent upon the concentration of symbionts and their activity, which is dependent upon light availability and temperature (Spero and Parker, 1985; Bemis et al., 2000; Peeters et al., 2002; Köhler-Rink and Kühl, 2005). These factors can vary seasonally and even within the life span of a planktonic foraminifera (Spero and Parker, 1985; Spero, 1998).

Aside from the Epstein et al. (1953) equation, eq. (1), two other paleotemperature equations were applied to the G. ruber data: 1) eq. (8), which was developed for O. universa under high light conditions (Bemis et al., 1998), but has been previously used to estimate temperatures for G. ruber (Thunell et al., 1999; Spero et al., 2003; Tedesco et al., 2007; Wejnert et al., 2010) and 2) eq. (16), which was developed using surface pump/plankton tow samples of G. ruber (Mulitza et al., 2003). All three equations used to predict calcification temperatures for G. ruber yield values within the surface mixed layer. However, eq. (1) frequently results in temperatures that are warmer than measured surface values, which is consistent with previous findings (Williams et al., 1981). Equation (8) yields temperatures associated with depths ranging from 0 to 65 m, with an average depth of 17 m. Similarly, eq. (16) results in temperatures that span depths ranging from 0 to 100 m, with an average of 35 m. Temperature estimates for G. ruber from eqs. (1) and (16) were ~3.1 ± 1.0 °C different, translating to an average depth difference of ~30 ± 15 m (Table 4). G. ruber is known to dwell in the surface mixed layer, predominately in the upper 50 m (Ravelo and Fairbanks, 1992; Niebler et al., 1999; Farmer et al., 2007), but may migrate into deeper water (Field, 2004; Wilke et al., 2009). While both eqs. (8) and (16) provide realistic results, eq. (8) produces temperatures for depths that best reflect G. ruber’s known depth preferences. Although eq. (8) was calibrated for O. universa, it yields the most realistic temperature estimates and resultant depth habitats for G. ruber. This may be due to the fact that both O. universa and G. ruber have dinoflagellate symbionts (Hemleben et al., 1989).

For G. bulloides, three paleotemperature equations were used besides the generic eq. (1): 1) eq. (9) developed by Bemis et al. (1998), 2) eq. (10) developed by Spero et al. (2000) and used by Bemis et al. (2002); and 3) eq. (14) developed by Mulitza et al. (2003). All three equations were developed specifically for G. bulloides, the first two using culture samples and the latter equation using plankton tow samples. As with
G. ruber, eq. (1) frequently yields temperatures that are warmer than the observed SST for G. bulloides δ¹⁸O data, particularly during the upwelling season. Likewise, eq. (14) yields temperatures that occasionally exceed measured SST values during winter–spring upwelling periods. Equations (9) and (10) yield similar temperature results for G. bulloides and predict calcification depths ranging from 0 to 110 m and 0–120 m with averages of 35 m and 45 m, respectively. The difference between the warmest temperature estimates, eq. (1) and the coldest, eq. (10) is ~3.8 ± 1.1 °C yielding an average depth habitat difference of ~40 ± 20 m for these temperature estimates, eq. (1) and the coldest, eq. (10) is ~3.8 ± 1.1 °C, which is equivalent to a ~0.06‰ difference in δ¹⁸O. However, the temperature estimates from eq. (2) are likely to be the most accurate because eq. (2) was developed for G. sacculifer and therefore would incorporate G. sacculifer’s strong species-specific fractionation.

The only oxygen isotope paleotemperature equation developed specifically for N. dutertrei is eq. (6), which was determined by Bouvier-Soumagnac and Duplessy (1985) using plankton tow samples. Other equations have been applied to N. dutertrei δ¹⁸O data including the generic eq. (1), and eq. (2), which was developed for G. sacculifer by Erez and Luz (1983) but applied previously to N. dutertrei data (Tedesco et al., 2007), and eq. (7), which was developed for O. universa in low light conditions, but also applied to N. dutertrei (Faul et al., 2000; Wejnert et al., 2010). For N. dutertrei, eqs. (1), (2), and (7) yield similar temperature results with calcification depths ranging from ~15 to 120 m and average depths of ~50–60 m. Conversely, eq. (6) yields much colder temperatures and consequently deeper depths ranging from ~60 to 200 m and an average of 165 m. The average difference between the warmest results from eq. (2) and coldest from eq. (6) is ~3.9 ± 1.8 °C, which is equivalent to a ~0.06‰ difference in δ¹⁸O. However, the temperature estimates from eq. (2) are likely to be the most accurate because eq. (2) was developed for G. sacculifer and therefore would incorporate G. sacculifer’s strong species-specific fractionation.
(dinoflagellate versus chrysophyte; Spero, 1998), they still accurately estimate calcification temperature. As the temperature estimates from eq. (1) provide a good average of the three equations yielding similar results, we used them in our attempt to reconstruct surface to thermocline temperature gradients (see Section 4.3).

Five equations have been developed for *O. universa*, although only four were evaluated in this study. Bouvier-Soumagnac and Duplessy (1985) determined two equations for this species, with eq. (3) derived from culture experiments and eq. (4) based on plankton tow samples. Bemis et al. (1998) also developed two equations for *O. universa* based on culturing experiments, one using low light conditions, eq. (7), and the other using high light conditions, eq. (8); however only the high light equation, eq. (8), was applied here because previous work at this location has used this equation (Tedesco et al., 2007). Spero et al. (2000) also determined a medium light equation (eq. 11) using cultured *O. universa* samples. *O. universa* dwells in the mixed layer and

![Graph showing depth habitats](image-url)

**Fig. 4.** Three point running averages of depth habitats estimated from calcification temperatures calculated from shell δ¹⁸O using the paleotemperature equations in Table 1 for *G. ruber* (A), *G. bulloides* (B), *G. sacculifer* (C), *O. universa* (D), *G. menardii* (E), *N. dutertrei* (F), and *G. crassaformis* (G). Dashed lines indicate the known depth habitat for each species. The X markers at the top of each plot indicate where data is available.
the thermocline at depths predominately ranging from 0 to 70 m, but as deep as 150 m (Niebler et al., 1999; Field, 2004; Farmer et al., 2007). The temperature results from eq. (1) and (3) are consistent with these observations with estimated depths ranging from ~5 to 160 m and average depths of ~55 m. Equations (4) and (11) yield temperatures that result in depth habitats ranging from ~15 to 200 m, with an average depth of ~75 m, which are slightly deeper than the known depth habitat. Lastly, eq. (8), the Bemis et al. (1998) high light O. universa equation yields even deeper calcification depths ranging from 30 m to 200 m, with an average of 100 m. The warmest O. universa temperatures were estimated by eq. (1) and the coldest by eq. (8) with an average difference between them of 1.8 ± 0.8 °C, which translates to a depth habitat difference of nearly 50 ± 20 m (Table 4). Based on the depth habitat results, eqs. (1) and (3) provide realistic temperature estimates, while the other equations, eqs. (4), (8), and (11) yield calcification depths that appear to be too deep for O. universa. The Bouvier-Soumagnac and Duplessy (1985) culture equation, eq. (3), is similar to the low light Bemis et al. (1998) equation, eq.
(7), and provides realistic temperature estimates, whereas the medium and high light equations yield colder temperatures. Thus it appears that low light paleotemperature equations for O. universa are the most appropriate for Cariaco Basin, which is consistent with our G. sacculifer results.

Aside from generic eq. (1), four other equations were used to estimate temperatures from G. menardii δ18O data: 1) eq. (2), which was developed by Erez and Luz for G. sacculifer, but has been previously applied to G. menardii (Tedesco et al., 2007), 2) eq. (5), which was derived for G. menardii by Bouvier-Soumagnac and Duplessy (1985) from plankton tow data, 3) eq. (7), which was developed by Bemis et al. (1998) using O. universa under low light conditions, but has also been applied to G. menardii (Faul et al., 2000; Wejnert et al., 2010), and 4) eq. (12), which was derived using culture samples of G. menardii (Spero et al., 2003). G. menardii lives within the thermocline to subthermocline region at depths ranging from 50 to 200 m (Oberhänsli et al., 1992; Niebler et al., 1999; Farmer et al., 2007; Mohtadi et al., 2009). Equations (5) and (12) yield temperatures that are frequently found above this depth, whereas δ18O data (Tedesco et al., 2007). We also evaluated the suitability of applying the two equations developed for G. menardii, eq. (5) and eq. (12), to G. crassaformis, since both are Globorotalia species. Equations (5) and (12) predict temperatures associated with depths ranging from ~65 to 300 m and an average depth of ~165 m, consistent with the known depth habitat of 100–500 m for this species in Cariaco Basin (Miro, 1971). Alternately, eqs. (1) and (2) result in shallower depth estimates of ~15–140 m with average depths of ~90 m. These four equations yield temperature estimates for G. crassaformis that differ by 1.9 ± 1.4 °C and depth habitats that are ~80 ± 50 m different (Table 4). Both eqs. (1) and (2) yielded calcification temperatures that equate to depth habitats that are too shallow for this species. Equation (1) consistently produced warm temperature estimates and eq. (2) was calibrated on G. sacculifer, which is known to have a large symbiont photosynthesis effect (Spero and Lea, 1993), whereas G. crassaformis is symbiont-barren (Hemleben et al., 1989). As with G. menardii, eqs. (5) and (12) yielded similar realistic temperatures.

No single oxygen isotope temperature equation yields accurate calcification temperatures and depth habitat estimates for all seven planktonic foraminiferal species examined in this study. Rather, our results indicate that equations must be selected carefully for each individual species. Average temperature estimates for each species varied by ~2–4 °C when using different equations (Table 4). This is consistent with the results of Bemis et al. (2002), in which it was found that species-specific equations are preferable because generic equations can yield misleading interpretations. Aside from eq. (14), the G. bulloides equation developed by Mulitza et al. (2003), the equations derived from plankton tow samples yielded some of the coldest temperature estimates, which is consistent with the results of Lin et al. (2011), who found that foraminifera from plankton tow samples typically had lower δ18O values relative to specimens of the same species from sediment trap samples. During the later stages of ontogeny, most planktonic foraminiferal species calcify at deeper depths, where colder temperatures lead to calcite with higher δ18O values (Duplessy et al., 1981; Erez and Honjo, 1981). Conversely, foraminiferal specimens from plankton tow samples often are collected before this 18O-enriched calcite is formed. Thus, equations developed using isotopically lighter plankton tow samples would yield colder temperatures when applied to the isotopically heavier sediment trap or sediment core samples. Conversely, the generic Epstein et al. (1953) equation, eq. (1), yielded some of the warmest temperatures, which is consistent with previous findings (Williams et al., 1981). Since eq. (1) was developed using mollusk data, it would not incorporate the effect of symbiont photosynthesis on foraminiferal δ18O. As the symbionts photosynthesize they remove CO2 from the microenvironment surrounding the calcifying shell, which increases pH (Jørgensen et al., 1985; Rink et al., 1998; Köhler-Rink and Kühl, 2005) and consequently lowers foraminiferal δ18O (Spero et al., 1997). As a result, foraminiferal δ18O, at least for those species containing symbionts, should be lower relative to the mollusk δ18O data used to calibrate eq. (1) and this would lead to warmer temperature estimates.

Since paleotemperature equations can produce significantly different temperature estimates when applied to the same isotope dataset (Bemis et al., 2002; King and Howard, 2005; Saraswat and Khare, 2010), one must carefully select the most appropriate equation for that species and location to estimate past temperatures. For Cariaco Basin, low light
paleotemperature equations appear to be the most appropriate. This may be because the euphotic zone in Cariaco Basin is typically only ~35–45 m (Lorenzoni et al., 2011) and since the planktonic foraminifera frequently dwell deeper, they would therefore be living in low light conditions. For the near surface dwelling species (**G. ruber**, **G. bulloides**, and **G. sacculifer**), there is no single equation that is appropriate for all species. Rather, species-specific culture equations provide the best temperature estimates. For the thermocline dwelling species (**O. universa** and **N. dutertrei**), the low light Bemis et al. (1998) **O. universa** equation (eq. 7) provides accurate temperature estimates. For the deeper dwelling species (**G. menardii** and **G. crassaformis**), equations calibrated on surface or thermocline dwelling species provide inaccurate results, whereas equations developed specifically for **G. menardii** (eqs. 5 and 12) provide the best results.

### 4.3. Surface water reconstructions

Documenting changes in upper ocean stratification is essential for understanding past climatic conditions and is commonly estimated by determining the difference in oxygen isotope composition of thermocline and surface-dwelling species of planktonic foraminifera (Spero et al., 2003; Cléroux et al., 2007; Farmer et al., 2007; Lin and Hsieh, 2007; Steph et al., 2009). Here we use the $\delta^{18}O$ difference ($\Delta\delta^{18}O$) between the surface dwelling **G. ruber** and the thermocline...
dwellng *N. dutertrei* to estimate the surface to thermocline temperature gradient in the Cariaco Basin for the period from 2005 to 2008 (Fig. 5). Temperature estimates for *G. ruber* were derived using eq. (8), while eq. (1) was used to estimate temperatures for *N. dutertrei*. The largest Δδ¹⁸O, and hence largest temperature gradient usually occurs during the summer when upwelling decreases and a strong thermal stratification develops. For the four-year study period, the average maximum summer season Δδ¹⁸O was 1.37 ± 0.1‰ (Fig. 6), which equates to an average surface to thermocline temperature difference of −4.5 ± 0.8 °C. During the winter the Δδ¹⁸O decreases to a minimum of 0.61 ± 0.1‰ (−1.3 ± 0.8 °C), which reflects the return of upwelling conditions. The Δδ¹⁸O was compared to an Ekman upwelling index for Cariaco Basin (Astor et al., 2013) and periods of strong upwelling clearly have the lowest Δδ¹⁸O (Fig. 6). There has been significant variability in upwelling intensity over the past 15 years, with strong upwelling from 1996 to 1997, and 2001–2003, and weak upwelling from 1998–2000 and 2004–2010 (Astor et al., 2013).

For our study period only 2003 had particularly strong upwelling, and this is reflected by the lowest SSTs, the highest sea surface salinities, and a small Δδ¹⁸O. Conversely 2004–2008 is characterized by higher SSTs, lower surface salinities, and larger Δδ¹⁸O (Figs. 2, 6). This is consistent with the findings of Tedesco et al. (2007), who found the smallest Δδ¹⁸O during 1997, a year of strong upwelling, whereas 1998–1999, which had weak upwelling, had higher Δδ¹⁸O.

5. Conclusions

Oxygen isotope measurements were conducted on seven species of planktonic foraminifera (*G. ruber, G. bulloides, G. sacculifer, N. dutertrei, O. universa, G. menardii, and G. crossoformis*) collected in biweekly sediment trap samples in the Cariaco Basin, Venezuela over a six-year period (2003–2008). Sixteen generic and species-specific δ¹⁸O temperature equations were applied to the data. The resultant calcification temperature estimates were compared to water column temperature measurements to determine depth habitats, which were used to assess the performance of the paleotemperature equations. No single equation can be accurately applied to all seven species, rather equations must be chosen for each species individually. The paleotemperature equations yield temperature estimates that varied by 2–4 °C, emphasizing the need to test paleotemperature equations in a range of environments in order to properly select an equation and interpret the sedimentary record. The coldest temperature estimates came from equations calibrated on plankton tow data, while the warmest temperature estimates were derived from the generic paleotemperature equations.

The differences in δ¹⁸O between the surface dwelling *G. ruber* and the thermocline dwelling *N. dutertrei* were used to assess upper ocean stratification. The differences were largest (1.37‰) during the summer when thermal stratification is strongest and smallest (0.61‰) during winter when upwelling mixes the water column. The Δδ¹⁸O was compared to an upwelling index for Cariaco Basin and the lowest Δδ¹⁸O occurred during strong upwelling. The strongest upwelling occurred in 2003, which had the lowest SSTs, highest sea surface salinities, and low Δδ¹⁸O.

Acknowledgments

We thank Eric Tappa for coordinating the Cariaco Basin sediment trap sample collection and for overseeing the stable isotope analyses. We are grateful to David Black and an anonymous reviewer whose insightful comments greatly improved this manuscript. This research was supported, in part, by the U.S. National Science Foundation award OCE-0752037.

Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.marmicro.2013.03.001.

References


