Flat meridional temperature gradient in the early Eocene in the subsurface rather than surface ocean

Sze Ling Ho and Thomas Laepple
Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research,
Telegrafenberg A43, D-14473 Potsdam, Germany.

(E-mail: ling.ho@uib.no; tlaepple@awi.de)

This supplementary file contains:

• Discussion on “Non-thermal effects on TEX86”.
• Discussion on “Implications and mechanisms of mesopelagic GDGT export”.
• Supplementary Figures S1 to S10.
• Supplementary Table S1.

Calibration ensemble derived from variability comparison is available in a separate Microsoft Excel file. The non-Eocene data compilation used in variability analyses is available in PANGAEA (https://doi.pangaea.de/10.1594/PANGAEA.149998).
Non-thermal effects on TEX$_{86}$

There is a growing body of evidence$^{1-5}$ that in addition to temperature, non-thermal secondary effects, including physiology, might have a strong control on individual TEX$_{86}$ records. Although unlikely to yield a systematic effect across the broad temporal and spatial range covered by the proxy records constituting the power spectra (Fig. 1) and thus our derived calibration equation, these factors likely contribute to TEX$_{86}$-based temperature reconstructions from individual sites. They might therefore also contribute to the model - TEX$_{86}^H$-derived seawater temperatures mismatch that remains after our proposed recalibration (Fig. 3).

Recent studies demonstrated that in addition to temperature, TEX$_{86}^H$ values in archaeal lipids are also sensitive to physiological factors such as growth phase$^4$, oxygen availability$^5$ and archaeal ecotypes$^{1-3}$.

In spite of a considerable TEX$_{86}^H$ change throughout archaeal growth, most of the GDGTs that are incorporated in sediments are plausibly from the stationary growth phase, regardless of climate states. Thus no systematic effect on the TEX$_{86}^H$ variability across time-scales from millennia to millions of years is expected. We further see no direct evidence that growth phase influences should lead to the relatively high Eocene TEX$_{86}^H$-temperatures in the Southern Ocean.

An influence of oxygen changes on TEX$_{86}^H$$^5$ might lead to non-temperature variations in records from specific regions, for example at the boundary of oxygen minimum zones. Given that we find a widespread variability mismatch between U$^{K-37}$ and TEX$_{86}^H$ i.e. at 20 of 22 globally distributed sites, it is likely not the major factor for the variability mismatch between both proxies. A warm bias due to low
oxygenation might lead to an overestimation of the temperature of warm climate states as the solubility of oxygen is reduced in warm waters\(^5\). However, it does not explain why the largest mismatch occurs in the Southern Ocean (Fig. 3), as here, signs of bioturbation and the lack of lamination in sediments\(^6\), suggest that the sediments were oxygenated during early Eocene. Furthermore, given the localities of these sites in the Southern Ocean (i.e. deepwater formation region), the bottom waters overlaying the sediments were probably relatively young and therefore should not be depleted of oxygen.

Changes in archaeal assemblages due to differences in regional oceanography and/or changes in the habitat depth could provide an explanation for the high \(\text{TEX}_{86}^H\)-derived temperature in high southern latitudes. Although the \(\text{TEX}_{86}^H\)-temperature relationship in Southern Ocean surface sediments (Pacific sector) agrees well with the global calibration\(^7\), the archaeal assemblage here might be different during the early Eocene. This is plausible in light of the substantially different temperature response of two strains of archaea despite originating from the same hydrographic setting\(^1\) (Benguela upwelling system). Temporal evolution in archaeal assemblage can introduce either a warm or cold bias depending on the main constituent of the assemblage. In this case, neither a global nor regional calibration based on the modern spatial temperature distribution would be able to account for these effects. Ultimately, a better mechanistic understanding of the thaumarchaeotal ecophysiology as well as the continued analysis of multiple independent proxy evidences is needed to reduce the uncertainty of past climate reconstructions.

**Implications and mechanisms of mesopelagic GDGT export**
Numerous studies suggest that mesopelagic (defined here as 200–1000m) GDGTs are exported to the marine sediments\textsuperscript{2,8–11}. The transport mechanism of mesopelagic GDGTs to the seafloor is likely via passive pathways, such as marine snow and faecal pellet formation. Mid-water maxima in faecal pellet flux are ubiquitous in the global ocean\textsuperscript{12,13}, suggesting an active food-web in dark mesopelagic ocean (also known as the twilight zone). For instance, GDGTs have been found in the guts of amphipods collected at depths (540m–1024m)\textsuperscript{14}. As archaea are too small to be grazed directly by amphipods, the GDGTs might have been ingested accidentally, or were embedded in larger particles (e.g. faecal pellets of smaller zooplankton) grazed by the amphipods. The latter is in agreement with a study in the Southwest Pacific which found picoplankton (of which archaea is a subset) within flagellates and copepod faecal pellets, both of which were incorporated in organic aggregates\textsuperscript{15}. Mesopelagic GDGTs may also be incorporated into large sinking particles via collision, due to differential settling\textsuperscript{16}, or attachment by sticky mucus / exopolymers\textsuperscript{17}.

The efficiency of downward export of the GDGT produced in the mesopelagic waters is lower than that from the epipelagic zone (<200m), where a substantially more active food web is present. However, its relative importance to the sedimentary GDGT pool may be amplified by the extensive loss (>90%) of epipelagic zone derived particle flux during its descent in the water column, with the majority of particles lost between 100m and 500m\textsuperscript{18}. In summary, the evidence from water column studies on suspended and sinking particulate matters, zooplanktons, as well as surface sediment studies, is consistent with our independent statistical evidence of a subsurface signal in TEX\textsubscript{86} records.

Some recent studies also suggest that different archaeal communities thriving at different water depths\textsuperscript{19,20} may produce GDGTs with different temperature
relationships\textsuperscript{21–24}. This complicates the interpretation of both TEX\textsubscript{86} values in water column studies and the depth origin of sedimentary TEX\textsubscript{86}. To investigate whether such a depth relationship would systematically bias our results or could be incorporated into the global calibration, we tested the influence of the water depth at the core-sites for the core-top calibration, as well as for the variability comparison of $U_{37}^{K}$ vs. TEX\textsubscript{86}, assuming that shallow sites only receive a minimal contribution from the deepwater GDGTs. Dividing the global calibration dataset\textsuperscript{25} in two sets of less and more than 1000m water depth does not result in a significantly different calibration slope ($p=0.56$, TEX\textsubscript{86} = $a\times$SST+$b$); similarly, dividing the downcore records of the variability comparison into shallow (<1000m) and deep records does not affect the ratio of $U_{37}^{K}$ to TEX\textsubscript{86} (<5\% ratio change).
Figure S1: Locations of proxy records in our three compilations, namely paired $U_{37}^K$-TEX$_{86}^H$ sites (proxy details in Fig. S2), paired surface-subsurface foraminiferal Mg/Ca sites (proxy details in Fig. S6), and early Eocene sites (proxy details in Tab. S1).
Figure S2: Paired $U_{37}^{K}$ and TEX$^{II}$$_{86}$ records constituting the composite power spectra in Fig. 1.

a. Sediment core details and proxy calibrations used in the original publications to convert index values into temperatures.

<table>
<thead>
<tr>
<th>Site</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Time coverage (kyr BP)</th>
<th>#samples (TEX86$^H$ + U37$^K$)</th>
<th>Reference for TEX86$^H$ calibration$^*$</th>
<th>Reference for U37$^K$ calibration$^*$</th>
<th>Reference for the proxy record</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD02-2515</td>
<td>27.48</td>
<td>-112.07</td>
<td>6.9–24.9</td>
<td>482</td>
<td>2</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>TTR17_293G</td>
<td>36.17</td>
<td>-2.75</td>
<td>0.3–19.4</td>
<td>180</td>
<td>26</td>
<td>29</td>
<td>29</td>
</tr>
<tr>
<td>NIOP905</td>
<td>10.78</td>
<td>51.93</td>
<td>0–22.2</td>
<td>216</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>SO42-74KL</td>
<td>14.32</td>
<td>57.33</td>
<td>0.4–22.1</td>
<td>146</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>GeoB7702-3</td>
<td>31.65</td>
<td>34.07</td>
<td>0.4–24.2</td>
<td>100</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>MD97-2146</td>
<td>20.12</td>
<td>117.38</td>
<td>0.4–28</td>
<td>144</td>
<td>33</td>
<td>34.35</td>
<td>34.35</td>
</tr>
<tr>
<td>MD98-2195</td>
<td>31.64</td>
<td>128.94</td>
<td>0.5–41.9</td>
<td>136</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>GeoB7926-2</td>
<td>20.22</td>
<td>-18.45</td>
<td>0.2–48.3</td>
<td>494</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>MD03-2607</td>
<td>-38.96</td>
<td>137.41</td>
<td>1.1–134.8</td>
<td>340</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP977A</td>
<td>36.03</td>
<td>-1.96</td>
<td>147.2–244.2</td>
<td>394</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1241</td>
<td>5.84</td>
<td>-86.44</td>
<td>1.8–147.9</td>
<td>118</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>MD97-2151</td>
<td>8.73</td>
<td>109.87</td>
<td>1–153.5</td>
<td>222</td>
<td>44</td>
<td>43</td>
<td>43</td>
</tr>
<tr>
<td>GeoB9528-3</td>
<td>9.17</td>
<td>-17.66</td>
<td>6.8–192</td>
<td>386</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1146</td>
<td>19.45</td>
<td>116.27</td>
<td>67–350.7</td>
<td>508</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1147</td>
<td>18.84</td>
<td>116.55</td>
<td>0.3–356.2</td>
<td>130</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1239</td>
<td>-0.67</td>
<td>82.08</td>
<td>0.4–431</td>
<td>488</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>MD06-2048</td>
<td>-26.17</td>
<td>34.02</td>
<td>0.4–790.1</td>
<td>354</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1143</td>
<td>9.36</td>
<td>113.29</td>
<td>82.1–3987.9</td>
<td>184</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP806</td>
<td>0.319</td>
<td>159.36</td>
<td>100–5260</td>
<td>108</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP882</td>
<td>50.37</td>
<td>167.60</td>
<td>479.5–5865</td>
<td>106</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP850</td>
<td>1.30</td>
<td>-111.52</td>
<td>0–11880</td>
<td>160</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>ODP1085</td>
<td>-29.38</td>
<td>13.99</td>
<td>15–13708</td>
<td>132</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
</tbody>
</table>

$^*$ Referring to the proxy calibration used in the original studies where the proxy records were presented.

b. Proxy time-series as published shown as temperature anomalies with a vertical scaling given by the vertical bar in the third column.
**Figure S3**: Single spectral estimates of $U_{37}^{K'}$- and $\text{TEX}^{H}_{86}$-inferred SST variability at study sites reported in Fig. S2. Single spectra show that $\text{TEX}^{H}_{86}$-variability (red) is larger than $U_{37}^{K'}$-variability (black) at 20 out of 22 sites, and that the discrepancy cannot be explained by independent noise as artificially adding white noise to $U_{37}^{K'}$ time series (green) does not result in similar spectral shape as $\text{TEX}^{H}_{86}$. PSD denotes power spectral density.
Figure S4: Comparison of power spectral density (PSD) derived from $U_{37}^K$ and TEX$_{86}$ SST; the latter based on the Bayesian calibration BAYSPAR$^{57}$. Dividing the TEX$_{86}$ spectral estimate by a factor of 4 leads to a similar spectra as $U_{37}^K$ SST. The stronger difference (factor of 4 instead of 3) between $U_{37}^K$ and TEX$_{86}$ SST is caused by BAYSPAR having on average a higher temperature sensitivity$^{58}$ than the classical TEX$_{86}^H$ calibration from Kim et al$^{25}$. Paired proxy records constituting the power spectra are as in Fig. 1, with details listed in Fig. S2.
**Figure S5**: Ratio of surface and subsurface (0–600 m) seawater temperature change in different climate change scenarios as simulated by the MPI-ESM model in the CMIP5/PMIP3 experiments. Shown are the ratios for **a**, Last Glacial Maximum (LGM) simulation minus preindustrial; and **b**, 4×CO$_2$ simulation minus preindustrial. Every black dot is one average ratio calculated from N randomly picked sites, where N varies on the x-axis and the calculation is repeated 100 times for each N. For single sites (N=1), the subsurface and surface change can be substantially different, whereas when averaged over many sites, the ratio converges to one.
**Figure S6:** Power spectral density (PSD) of paired Mg/Ca temperature records based on sea surface- and subsurface-dwelling foraminifera, in comparison with the spectra of $U_{37}^{K}$ and $TEX_{86}^{H}$-SST.

**a,** Proxy record details and foraminiferal species in the Mg/Ca records constituting the power spectra in **b** using the calibrations proposed by the original studies.

<table>
<thead>
<tr>
<th>Site</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Surface dweller</th>
<th>Subsurface dweller</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD01-2390</td>
<td>6.64</td>
<td>113.41</td>
<td><em>G. ruber</em></td>
<td><em>P. obliquiloculata</em></td>
<td>59</td>
</tr>
<tr>
<td>MD05-2904</td>
<td>19.46</td>
<td>116.25</td>
<td><em>G. ruber</em></td>
<td><em>P. obliquiloculata</em></td>
<td>59</td>
</tr>
<tr>
<td>ODP1240</td>
<td>0.02</td>
<td>-86.46</td>
<td><em>G. ruber</em></td>
<td><em>N. dutertrei</em></td>
<td>62</td>
</tr>
<tr>
<td>SO213-59/2</td>
<td>-45.83</td>
<td>-116.88</td>
<td><em>G. bulloides</em></td>
<td><em>G. truncatulinoides</em></td>
<td>64</td>
</tr>
<tr>
<td>GeoB10038-4</td>
<td>-5.94</td>
<td>103.25</td>
<td><em>G. ruber</em></td>
<td><em>N. dutertrei</em></td>
<td>59</td>
</tr>
<tr>
<td>ODP999A</td>
<td>12.74</td>
<td>-78.74</td>
<td><em>G. sacculifer</em></td>
<td><em>N. dutertrei</em></td>
<td>65</td>
</tr>
<tr>
<td>ODP1241C</td>
<td>5.84</td>
<td>-86.45</td>
<td><em>G. sacculifer</em></td>
<td><em>N. dutertrei</em></td>
<td>50</td>
</tr>
<tr>
<td>MD01-2378</td>
<td>-13.08</td>
<td>121.79</td>
<td><em>G. ruber</em></td>
<td><em>P. obliquiloculata</em></td>
<td>68</td>
</tr>
<tr>
<td>MD06-3067</td>
<td>6.51</td>
<td>126.50</td>
<td><em>G. ruber</em></td>
<td><em>P. obliquiloculata</em></td>
<td>68</td>
</tr>
<tr>
<td>VM12-107</td>
<td>11.33</td>
<td>-66.63</td>
<td><em>G. ruber</em></td>
<td><em>G. crassaformis</em></td>
<td>69</td>
</tr>
<tr>
<td>MD01-2461</td>
<td>51.75</td>
<td>-12.92</td>
<td><em>G. bulloides</em></td>
<td><em>N. pachyderma</em></td>
<td>59,71</td>
</tr>
<tr>
<td>SO18460</td>
<td>9.09</td>
<td>129.24</td>
<td><em>G. ruber</em></td>
<td><em>P. obliquiloculata</em></td>
<td>67</td>
</tr>
<tr>
<td>ODP1446</td>
<td>19.46</td>
<td>116.27</td>
<td><em>Gs. Subquadratus + Gs. obliquus</em></td>
<td><em>Gs. Trilobus</em></td>
<td>73</td>
</tr>
<tr>
<td>WIND28K</td>
<td>-10.15</td>
<td>51.01</td>
<td><em>G. ruber</em></td>
<td><em>N. dutertrei</em></td>
<td>74,75</td>
</tr>
<tr>
<td>64PE-174P13</td>
<td>-29.76</td>
<td>2.40</td>
<td><em>G. ruber</em></td>
<td><em>G. truncatulinoides</em></td>
<td>76</td>
</tr>
<tr>
<td>GeoB12610-2</td>
<td>-4.82</td>
<td>39.42</td>
<td><em>G. ruber</em></td>
<td><em>N. dutertrei</em></td>
<td>77</td>
</tr>
</tbody>
</table>

**b,** Raw average power spectra for $TEX_{86}^{H}$, $U_{37}^{K}$ and Mg/Ca data of surface- and subsurface-dwelling foraminifera
Figure S7: Histogram of TEX$^H_{86}$ to temperature slopes derived from comparing the U$^K_{37}$ and TEX$^H_{86}$ variability. The distribution includes the uncertainty of the TEX$^H_{86}$ to U$^K_{37}$ ratio, as well as the uncertainty of the U$^K_{37}$ to SST calibration. The resulting slopes are significantly smaller than the commonly used TEX$^H_{86}$ calibrations.
**Figure S8**: Implications of recalibrated TEX$^H_{86}$ on selected published time series across timescales. For visual purposes, only the mean calibration and not the full calibration ensemble is shown.

**a**, TEX$^H_{86}$ records in our paired U$^{K_{37}}_{37}$-TEX$^H_{86}$ compilation (Fig. S2) consisting of time series spanning at least the Last Glacial Maximum to as far back as the Miocene. All time series are anomalies relative to the time period shown. We note that the same dataset was used in the construction of the calibration but as only a single parameter was estimated from 22 paired time-series, the artificial skill is very small.

**b**, Implications of recalibrated TEX$^H_{86}$ for transient changes at site ODP1172 during Middle Eocene Climatic Optimum (MECO) derived from U$^{K_{37}}_{37}$ and TEX$^H_{86}$. MECO is as defined in the original publication. The recalibration results in a convergence of the amplitude of MECO warming between proxies. All time series are anomalies relative to the time period shown.
**Figure S9:** Comparison of ECHAM5/MPI-OM with other EoMIP models studied by Lunt et al.\(^9\).

a, Eocene warming.

**Comparison of EoMIP models**

![Graph showing comparison of EoMIP models](image)

b, As in Fig. 3a, with the addition of ECHAM5 temperature at TEX\(^\text{H}_{15}\) sites, and EoMIP ensemble mean temperature for both the zonal mean and at TEX\(^\text{H}_{15}\) sites (except West Siberia Sea site because paleolongitude data is not available in the original publication; refer to Supplementary Table S1).

**Early Eocene sea surface**

![Graph showing early Eocene sea surface temperatures](image)
Figure S10: Seasonal range of simulated Eocene zonal mean surface and subsurface (0–600m) temperatures. The seasonal range is negligible for the subsurface case and is thus not considered in the proxy-model comparison shown in Fig. 3b.
Table S1: Site details for early Eocene temperature data in Fig. 3. The compilation is an extension to the EoMIP proxy compilation (Lunt et al.94), with the addition of two new TEX$_86^H$ sites.

<table>
<thead>
<tr>
<th>Site</th>
<th>Present-day Latitude (°N)</th>
<th>Present-day Longitude (°E)</th>
<th>Paleo-latitude during 49–55Ma (°N)</th>
<th>Paleo-latitude during 49–55Ma (lower bound)</th>
<th>Paleo-latitude during 49–55Ma (upper bound)</th>
<th>Proxy</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>ODP 690</td>
<td>-65.2</td>
<td>1.2</td>
<td>-71.9</td>
<td>-80.1</td>
<td>-64.1</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>74</td>
</tr>
<tr>
<td>ODP 738</td>
<td>-62.7</td>
<td>82.8</td>
<td>-62.3</td>
<td>-68.6</td>
<td>-56.2</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>81/81</td>
</tr>
<tr>
<td>ODP 1172</td>
<td>-44.0</td>
<td>149.9</td>
<td>-57.8</td>
<td>-62.8</td>
<td>-52.8</td>
<td>TEX$_86^H$</td>
<td>84</td>
</tr>
<tr>
<td>Waipara River, NZ</td>
<td>-43.2</td>
<td>172.8</td>
<td>-47.8</td>
<td>-52.0</td>
<td>-44.1</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>84</td>
</tr>
<tr>
<td>Tanzania (TDP)</td>
<td>-7.8</td>
<td>39.1</td>
<td>-22.4</td>
<td>-24.4</td>
<td>-20.2</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>85</td>
</tr>
<tr>
<td>Hatcheligbee</td>
<td>32.0</td>
<td>-88.0</td>
<td>29.8</td>
<td>26.4</td>
<td>33.2</td>
<td>Clumped isotopes</td>
<td>86</td>
</tr>
<tr>
<td>ACEX (302-4A)</td>
<td>87.9</td>
<td>136.2</td>
<td>77.4</td>
<td>68.6</td>
<td>88.0</td>
<td>TEX$_{cl}$</td>
<td>87</td>
</tr>
<tr>
<td>Bass River</td>
<td>39.6</td>
<td>-74.4</td>
<td>34.1</td>
<td>31.9</td>
<td>36.2</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>88</td>
</tr>
<tr>
<td>Wilson Lake</td>
<td>39.7</td>
<td>-75.1</td>
<td>34.2</td>
<td>32.1</td>
<td>36.4</td>
<td>TEX$_{cl}$</td>
<td>90</td>
</tr>
<tr>
<td>ODP 865</td>
<td>18.4</td>
<td>-179.6</td>
<td>11.4</td>
<td>9.8</td>
<td>12.8</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>90</td>
</tr>
<tr>
<td>DSDP 527</td>
<td>-28.0</td>
<td>1.8</td>
<td>-41.2</td>
<td>-43.6</td>
<td>-38.4</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>91</td>
</tr>
<tr>
<td>ODP 1209</td>
<td>32.7</td>
<td>158.5</td>
<td>29.0</td>
<td>26.6</td>
<td>31.3</td>
<td>Foraminifera $\delta^{18}O$</td>
<td>92</td>
</tr>
<tr>
<td>IODP U1356</td>
<td>-63.0</td>
<td>135.0</td>
<td>-57.6</td>
<td>-62.7</td>
<td>-52.6</td>
<td>TEX$_{cl}$</td>
<td>93</td>
</tr>
<tr>
<td>West Siberia Sea</td>
<td>53.5</td>
<td>73.5</td>
<td>51.5</td>
<td>47.8</td>
<td>56.2</td>
<td>TEX$_{cl}$</td>
<td>93</td>
</tr>
</tbody>
</table>
References


39. Lopes dos Santos, R. A. et al. Comparison of organic (U$_{K'37}^+$, TEX$^H_{86}$, LDI) and faunal proxies (foraminiferal assemblages) for reconstruction of late


