OXYGEN ISOTOPES IN PHOSPHATIC FISH REMAINS FROM ISRAEL: PALEOTHERMOMETRY OF TROPICAL CRETACEOUS AND TERTIARY SHELF WATERS

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Abstract


The isotopic composition of oxygen in phosphate was measured on 24 fossil fishes from Israel and Jordan ranging in age from Cretaceous to Eocene. From these, the first paleotemperatures were calculated for shallow water Cretaceous Paleotropics. The results were compared to those of fish from other Mediterranean localities and to Late Cretaceous marine reptiles from Israel. $\delta^{18}O$ in the fish average 18.0% in the Early Cretaceous, 17.8% in the Cenomanian, 18.3% in the Coniacian, 18.2% in the Santonian, 19.4% in the Campanian, 20.1% in the Maastrichtian and 18.4% in the Early Eocene. Such variations suggest a warm Cenomanian–Turonian, cooling towards the end of the Cretaceous, and warming up again in the Eocene.

The deduced temperatures for the Cretaceous range between 20°C and 34°C, which is about 10 ° higher than Cretaceous water surface temperatures in Northern Europe. Apatite from marine reptile skeletons yields similar $\delta^{18}O$ to coexisting fish.

The good agreement between these and previously known paleotemperature data give further credence to the validity of paleotemperatures obtained by isotopic analysis of oxygen in phosphates.

The relatively large thermal gradient between middle and low paleolatitudes favors Barron’s (1983) “coolest” Cretaceous model.

Introduction

The application of oxygen isotopes to the understanding of paleoclimatology has been a major tool in the past thirty-five years. A large majority of isotopic paleotemperature data has been obtained by the isotopic analysis of oxygen from calcium carbonate of foraminifers from the deep sea and of belemnites and other molluscs. Among the major factors which have hampered the interpretation of such data was the considerable isotopic variation which may result from diagenetic and epigenetic alteration of the shells (see Urey et al., 1951; Spaeth et al., 1971; Stevens and Clayton, 1971; Savin, 1977). Because both fresh water dilution of sea water and diagenesis typically result in isotopically lighter values of the carbonates, there is a tendency to rely only on the isotopically heaviest measurements (Barron, 1983). Such selection may introduce an unwarranted bias toward low temperatures. A method which is less sensitive to alteration, would enable one to use mean isotopic measurements rather than the heaviest values.

Recent investigations by Kolodny et al. (1983), Shemesh et al. (1983), Luz et al. (1984) and Karhu and Epstein (1986) have demon-
strated the possibility of using oxygen isotopes in phosphates as paleothermometers. The previously proposed (Longinelli and Nuti, 1973) $\delta^{18}\text{O}$ in phosphate ($\delta^{18}\text{O}_p$) vs. water ($\delta^{18}\text{O}_w$) temperature scale has been reconfirmed from independent measurements, and the isotopic composition of a set of living fish from known environments was analyzed. The linear relationship between $(\delta^{18}\text{O}_p - \delta^{18}\text{O}_w)$ and temperature ($t$) is

$$t^\circ C = 111.4 - 4.3(\delta^{18}\text{O}_p - \delta^{18}\text{O}_w)$$

(1)

The major advantage in using the "$\delta^{18}\text{O}$ in phosphate" paleothermometer is in its extreme insensitivity to postdepositional isotopic exchange reactions. It has been shown (see Kolodny et al., 1983, and references therein) that temperature and water composition of the environment is recorded very sensitively in the isotopic composition of the phosphate of fish bones and teeth, which is extremely well preserved after the organism’s death. Since only enzyme-catalyzed reactions seem to enable rapid isotopic exchange of oxygen between phosphate and water it is assumed that even inorganic dissolution–precipitation or recrystallization of a fish bone at low temperature will not alter its isotopic composition. Obviously introduction of "foreign" phosphate, e.g. massive apatitic cement, may alter the original phosphate composition; but as long as care is being taken to avoid such samples, the isotopic analysis of oxygen in fishapatite has the potential of yielding a reliable record of aquatic temperatures from the Devonian to Recent. The simplest part of such an attempt would be the isotopic analysis of marine fish bones, where the assumptions on the isotopic composition of the aquatic phase ($\delta^{18}\text{O}_w$) are the least demanding. If marine fishes of different ages, and from different paleogeographic locations, can be analyzed, then paleotemperature maps of past oceans may be achievable.

A first set of fishes from Devonian to Recent age has been analyzed in our laboratory (Kolodny, in prep.). The resulting time trend is similar to that observed by Shemesh et al. (1983) viz.: a decrease in $\delta^{18}\text{O}_p$ in fish of older geological age, suggesting a gradual cooling of the Earth’s marine temperatures in the last 400 million years.

A preliminary set of measurements on younger (post-Cretaceous) shark teeth from sample sites, from which mollusc carbonate-isotopic temperatures were obtained earlier by Buchardt (1978), shows a good agreement between the two scales. It is admitted that true temperatures may be higher than the calculated carbonate temperatures, since, as explained above, only the highest $\delta^{18}\text{O}$ values are admitted in carbonate thermometry.

We sampled a limited area, Israel and some western Mediterranean outcrops, over a limited time span — between Cretaceous and Eocene. The time span extends over about 80 million years, while the area amounts to a single sampling point for most practical purposes of graphical presentation. In addition an attempt is made to check the usefulness of marine reptiles as tools for paleothermometry.

The material analyzed from the Israeli Senonian comprises specimens belonging to three major groups, viz. selachians, teleosts (Table I) and marine reptiles (Table II), whereas that from the Eocene comprises only sharks. The specimens are well preserved, allowing their osteological and systematic identification. Most of the genera dealt with are regarded as nectonic, good swimmers, that inhabited both the open sea and coastal regions. Moreover, some of them (e.g. *Squalicorax*, *Enchodus* and the mosasaurs among others) are cosmopolitan occurring from the southern United States and northern Brasil in the west to Iraq in the east (Arambourg and Signeux, 1962; Pivetau, 1955; Arambourg et al., 1959; Goody, 1969; Herman, 1975). Thus, although the specimens were found in shallow water deposits they indicate good connections to the open sea. The age determinations are based on the world wide distribution of the taxa as well as on the ages of the associated invertebrate megafossils and foraminiferids (Reiss, 1962; Raab, 1963; Herman, 1975; Shiloni et al., 1977). It should be...
TABLE I

Oxygen isotope composition of Cretaceous to Eocene fish teeth and bones

<table>
<thead>
<tr>
<th>No.</th>
<th>Group</th>
<th>Name</th>
<th>Locality</th>
<th>Age</th>
<th>$\delta^{18}O$</th>
<th>Temp. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Holost, T</td>
<td>Pycnodontidae</td>
<td>Gevaram Well</td>
<td>Berriasian–Valanginian</td>
<td>18.10</td>
<td>29</td>
</tr>
<tr>
<td>2</td>
<td>Holost, T</td>
<td>Pycnodontidae</td>
<td>Hamakhtesh Hagadol</td>
<td>Aptian-Albian</td>
<td>17.90</td>
<td>30</td>
</tr>
<tr>
<td>3</td>
<td>Shark, V</td>
<td>Unidentified</td>
<td>Ein Yabrud</td>
<td>Cenomanian</td>
<td>17.80</td>
<td>31</td>
</tr>
<tr>
<td>4</td>
<td>Teleost, B</td>
<td>Eubiodectes</td>
<td>Ein Yabrud</td>
<td>Cenomanian</td>
<td>17.80</td>
<td>31</td>
</tr>
<tr>
<td>5</td>
<td>Holost, B</td>
<td>Pycnodontidae</td>
<td>Ein Yabrud</td>
<td>Cenomanian</td>
<td>17.90</td>
<td>30</td>
</tr>
<tr>
<td>6</td>
<td>Holost, B</td>
<td>Pachyamia</td>
<td>Ein Yabrud</td>
<td>Cenomanian</td>
<td>17.80</td>
<td>31</td>
</tr>
<tr>
<td>7</td>
<td>Teleost, B</td>
<td>Unidentified</td>
<td>Kefar Shaul</td>
<td>Late Cenomanian</td>
<td>17.40</td>
<td>32</td>
</tr>
<tr>
<td>8</td>
<td>Teleost, T</td>
<td>?Palaeoalismum sp.</td>
<td>G. Rakaba, Sinai</td>
<td>Early Turonian</td>
<td>17.20</td>
<td>33</td>
</tr>
<tr>
<td>9</td>
<td>Shark, T</td>
<td>Scapanorhynchus raphidodon</td>
<td>W. Garur, Sinai</td>
<td>Early Coniacian</td>
<td>18.80</td>
<td>27</td>
</tr>
<tr>
<td>10</td>
<td>Shark, T</td>
<td>Squalicorax falcatus</td>
<td>W. Garur, Sinai</td>
<td>Early Coniacian</td>
<td>18.00</td>
<td>30</td>
</tr>
<tr>
<td>11</td>
<td>Shark, T</td>
<td>Scapanorhynchus sp.</td>
<td>N. Adasha</td>
<td>Late Santonian</td>
<td>18.30</td>
<td>28</td>
</tr>
<tr>
<td>12</td>
<td>Shark, T</td>
<td>Scapanorhynchus rapax</td>
<td>En Aqarabim</td>
<td>Santonian–Campanian</td>
<td>18.60</td>
<td>27</td>
</tr>
<tr>
<td>13</td>
<td>Shark, V</td>
<td>Scapanorhynchus rapax</td>
<td>En Aqarabim</td>
<td>Santonian–Campanian</td>
<td>17.50</td>
<td>32</td>
</tr>
<tr>
<td>14</td>
<td>Shark, T</td>
<td>Scapanorhynchus sp.</td>
<td>Har Letaot</td>
<td>L. Santonian–E. Campanian</td>
<td>18.30</td>
<td>28</td>
</tr>
<tr>
<td>15</td>
<td>Shark, T</td>
<td>?Squatina sp.</td>
<td>Mishor Rotem</td>
<td>E. Maastrichtian</td>
<td>20.30</td>
<td>20</td>
</tr>
<tr>
<td>16</td>
<td>Teleost, V</td>
<td>Unidentified</td>
<td>Mishor Rotem</td>
<td>E. Maastrichtian</td>
<td>19.80</td>
<td>22</td>
</tr>
<tr>
<td>17</td>
<td>Shark, T</td>
<td>Lamna biauriculata</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>19.30</td>
<td>24</td>
</tr>
<tr>
<td>18</td>
<td>Shark, T</td>
<td>Squalicorax haupi</td>
<td>Hamakhtesh Haqatan</td>
<td>L. Campanian</td>
<td>18.70</td>
<td>27</td>
</tr>
<tr>
<td>19</td>
<td>Teleost, T</td>
<td>Enchodus bursuaxi</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>19.70</td>
<td>22</td>
</tr>
<tr>
<td>20</td>
<td>Teleost, B</td>
<td>Enchodus libycus</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>19.10</td>
<td>25</td>
</tr>
<tr>
<td>21</td>
<td>Teleost, T</td>
<td>Enchodus bursuaxi</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>20.00</td>
<td>21</td>
</tr>
<tr>
<td>22</td>
<td>Shark, T</td>
<td>Squalicorax prisodontus</td>
<td>Ruseifa, Jordan</td>
<td>L. Campanian</td>
<td>18.40</td>
<td>28</td>
</tr>
<tr>
<td>23</td>
<td>Shark, T</td>
<td>Procarcharodon auriculatus</td>
<td>Hor Hahar</td>
<td>Early Eocene</td>
<td>18.20</td>
<td>29</td>
</tr>
<tr>
<td>24</td>
<td>Shark, T</td>
<td>Lamna obliqua</td>
<td>Hor Hahar</td>
<td>Early Eocene</td>
<td>18.10</td>
<td>29</td>
</tr>
<tr>
<td>25</td>
<td>Fish, B</td>
<td>&quot;Anaethalion&quot; vidali</td>
<td>Spain</td>
<td>Berriasian–Valanginian</td>
<td>18.10</td>
<td>29</td>
</tr>
<tr>
<td>26</td>
<td>Shark, T</td>
<td>Lamna biauriculata</td>
<td>Ruseifa, Jordan</td>
<td>L. Campanian</td>
<td>19.50</td>
<td>23</td>
</tr>
<tr>
<td>27</td>
<td>Fish, B</td>
<td>Unidentified</td>
<td>Verona, Italy</td>
<td>Early Eocene</td>
<td>18.70</td>
<td>27</td>
</tr>
<tr>
<td>28</td>
<td>Shark, T</td>
<td>Unidentified</td>
<td>Morocco</td>
<td>Early Eocene</td>
<td>18.60</td>
<td>27</td>
</tr>
</tbody>
</table>

*T = tooth, B = bone, V = vertebra. *If not marked otherwise — Israel. *Assuming constant isotopic composition of sea-water, $\delta^{18}O_w = -1\%o$.

TABLE II

Oxygen isotope composition of Senonian reptiles

<table>
<thead>
<tr>
<th>No.</th>
<th>Name</th>
<th>Locality</th>
<th>Age</th>
<th>$\delta^{18}O$, (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Leiodon cf. anceps, T</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>18.8</td>
</tr>
<tr>
<td>2</td>
<td>Leiodon cf. anceps, B</td>
<td>Zefa-Efe</td>
<td>L. Campanian</td>
<td>18.9</td>
</tr>
<tr>
<td>3</td>
<td>Globidens sp., B</td>
<td>Zefa</td>
<td>L. Campanian</td>
<td>18.7</td>
</tr>
<tr>
<td>4</td>
<td>Plesiosauria, V</td>
<td>Zin valley</td>
<td>Santonian–L. Campanian</td>
<td>17.3</td>
</tr>
<tr>
<td>5</td>
<td>Leiodon cf. anceps, T</td>
<td>Ruseifa, Jordan</td>
<td>L. Campanian</td>
<td>18.3</td>
</tr>
</tbody>
</table>

*T = Tooth, B = Bone, V = Vertebra. *Assuming constant isotopic composition of sea water, $\delta^{18}O_w = -1\%o$.

Experimental

Teeth and bones were collected aiming at the best preserved and least cemented specimens.

noted that both the Moroccan and the Transjordanian specimens examined are of Upper Campanian rather than of Maastrichtian age (see Reiss, 1962; Herman, 1975).
There is no need of cleaning the samples from non apatitic components since the applied method is selective and chemically singles out only PO\textsuperscript{2-} oxygen. The method is practically "blind" to any oxygen of other source. The analytical procedure is essentially a slightly modified Tudge (1960) method (see Kolodny et al., 1983) and consists of sample dissolution in nitric acid, its purification as ammonium phosphomolybdate and magnesium ammonium phosphate, and its precipitation as BiPO\textsubscript{4}. Bismuth phosphate is dried, fluorinated in stainless steel vessels at 150°C, oxygen converted to CO\textsubscript{2} and analyzed mass spectrometrically. The standard reproducibility of our results is ±0.3%. All results are reported in permil with respect to SMOW. All samples were analyzed in duplicate, and an average value is reported.

Results and discussion

Tables I and II summarize the analytical results. It is obvious that the total variance of \(\delta^{18}O\) values in the biogenic apatites reported here is rather small: 17.2‰ to 20.3‰. Such a range of 3.1‰ corresponds to 13°C at most, assuming a constant isotopic composition of sea water (\(\delta^{18}O_w\)). Part of the variance is certainly due to "analytical noise", and a large part due to \(\delta^{18}O_w\) variations. Since we are dealing with a shallow, epicontinental sea allowance should be made for variations in water salinity — from slightly hypersaline, to slightly brackish — and hence in water isotopic composition.

Comparing our observed \(\delta^{18}O_p\) range with the total range in biogenic fossil apatites, which is about 13‰ (12.5–25‰, Shemesh et al., 1983), it appears that the Levantine Cretaceous was an extremely stable, barely changing environment.

Cretaceous paleothermometry

Assuming a preglacial sea water isotopic composition of −1‰ (SMOW) (Savin, 1977), and applying Eq. (1), the temperature range of the Senonian of Israel is between 20 and 32°C (Table I). It seems, however, that most of the \(\delta^{18}O_p\) values for the Upper Campanian (which was sampled best) center around an average value of 19.2‰, and around a lower value (−18.2‰) for the Santonian to Lower Campanian. These values yield temperatures of 24°C for the Upper Campanian and 29°C for the Santonian. The temperatures obtained for the Cenomanian (\(\delta^{18}O_p = 17.8\)) are 31°C, for the Coniacian 28°C and the Early Eocene 29°C. The measured \(\delta^{18}O_p\) and the estimated temperatures are plotted on Fig.1 as a function of geologic age.

The small spread of values ensures that if we were to follow the logic usually applied in carbonate paleothermometry, of using only the most \(^{18}O\) enriched samples, temperature estimates would not be changed by much: 27°C for the Coniacian, 21°C for the Upper Campanian and 29°C for the Early Eocene. Obviously, the latter point suffers from some circular reasoning: as mentioned before, it is because \(\delta^{18}O_p\) is less sensitive to post depositional isotopic alteration, i.e. shows a smaller spread, that this method is being applied here.

Attention should be called to the fact that the isotopic compositions of fossil marine reptiles (Table II, Fig.1) agree well with the range of values measured for the contemporaneous fish. This suggests the use of marine reptiles for paleoenvironmental analysis.

Before attempting any comparison between our results and previous measurements it is essential to understand what is the depth habitat of the analyzed fish. In other words, at what depth in the water column is our paleothermometer located. Fortunately, paleogeographic and paleontological analyses make the answer to this question rather clear. During most of the relevant time we are dealing with an extremely shallow water body, apparently never exceeding, in the shelf region of southern Israel, several tens of meters, sometimes being reduced to a few meters only (Steinitz and Kolodny, 1978). The shallow water depth of the Campanian Mishash sea is probably part of the explanation for the
Fig. 1. Change of Cretaceous–Eocene paleotemperatures from $\delta^{18}O$ values of fish and reptiles. The solid line was drawn through average $\delta^{18}O$ calculated from fish data of each stratigraphic age, for samples from Israel. The hatched line goes through the "heaviest" points of each age. Note agreement of reptile data with general pattern.

Shemesh et al. (1983) finding that in Campanian phosphorites in southern Israel (the Negev) there is no isotopic fractionation between vertebrate bone fragments and ovulites of apatite, which were probably formed at or below the sediment–water interface. Thus the calculated temperatures obtained on skeletal elements of actively swimming fish probably reflect some averaged, integrated temperature of a rather isothermal water body. Another aspect of the thermal uniformity of the Cretaceous Sea in the Levant was also noted by Shemesh et al. (1985) who showed that over the area of the Mediterranean Tethyan phosphorite province $\delta O_p$ varied by no more than 2% between Turkey and Senegal. Again, there is little doubt that at least part of this variation is due to changes in water isotopic composition.

Specifically in equatorial areas low salinity and low $\delta^{18}O$ water is produced by heavy rainfall and river discharge. The Gulf of Guinea and the equatorial Atlantic are analogous areas in the present. In such areas, however, the variation in $\delta^{18}O_w$ is considerably smaller than that of salinity, since the $\delta^{18}O$ of equatorial rain does not differ very much from that of sea water. Moreover, it is difficult to imagine water composition varying drastically whereas $\delta^{18}O_p$ remains so stable; in other words the decrease in $\delta^{18}O_w$ being precisely compensated by temperature (see discussion of Bardawil Lagoon in Kolodny et al., 1983).

Inspection of the general trend apparent in our data (Fig. 1) seems to indicate a rather warm Cenomanian–Turonian, a cooling which begins in the Coniacian reaching a minimum temperature in the Campanian and a warming up again towards the Eocene.

We detected a similar trend in four isotopic analyses of Mesozoic fish from Western Mediterranean localities (Samples 25–28 in Table I, Figure 1). They too indicate a rather low temperature in the Upper Campanian of Morocco ($\delta^{18}O_p = 19.5\%$), which has been reached from a warmer period in the Lower Cretaceous of Spain ($\delta^{18}O = 18.1\%$) and followed again by a warming up in the Eocene, as indicated by fish from both Italy and Morocco ($\delta^{18}O_p = 18.7\%, 18.6\%$).

A general cooling trend from the Albian to the end of the Cretaceous (Maastrichtian),
followed by a temperature rise in the Eocene has been observed before in isotopic studies on planktonic foraminiferids (Savin, 1977, 1982; Douglas and Woodruff, 1981; Hallam, 1985). Generally this trend is in agreement with both belemnite isotopic analyses from uplifted marine sediments in Northern Europe and with paleontologic-paleobotanical evidence (Savin, 1977). A net cooling of "as much as 8–10°C in both Northern Europe and the Tropical Pacific Ocean" has been suggested for the Albian-Maastrichtian temperature drop (Savin, 1977, p. 331).

Specifically, the temperatures quoted here agree very well with both the values and the trend of temperatures for ocean surface in low latitudes [as reported by Douglas and Woodruff (1981) from planktonic foraminiferids and coccoliths (Fig.2)]. Our result seems to indicate that the eastern part of the Tethys was fed by roughly the same water body as the upper water mass of the Equatorial Paleo-Pacific. Such a conclusion should be qualified in view of the very shallow water depth which has been suggested for the studied area. Under such conditions the entire water column may be very warm (e.g. the present-day Persian Gulf) and possibly warmer by a few degrees than open ocean waters. In that case the estimated temperatures, though correct, would not be representative of their latitudinal band.

As mentioned above the general trend of the paleotemperature record of the Cretaceous to Eocene Middle East is similar to that of Northern Europe. This trend is characterized by an apparent maximum in the Albian followed by a very distinct maximum in the Turonian (or Coniacian) and a sharp drop of temperatures towards the Campanian–Maastrichtian (Lowenstam and Epstein, 1954; Lowenstam, 1964; Spaeth et al., 1971). The Eocene is again a period in which warm temperatures seemed to prevail in Northern Europe (Shackleton and Boersma, 1981).

Whereas trends agree, absolute values differ; most quoted Cretaceous temperatures for Northern Europe range between 10°C and 20°C. The temperatures arrived at here are 20–30°C. Such a 10°C difference persists both for the "coolest" data point (Upper Campanian–Maastrichtian) as well as the "warmest" time (Cenomanian–Turonian and Eocene).

Whereas the published North European results on uplifted marine shallow water sediments are from paleolatitudes of 30–50°N (see Bowen, 1966; Savin, 1977; Barron and Washington, 1982, fig.9), the paleolatitude of Israel in Late Cretaceous to Early Tertiary times was about 10°N (Helsley and Nur, 1970; Ron et al., 1984). Hence the reported results are probably the closest estimates of equatorial shallow water temperatures. A paleolatitudinal temperature analysis was attempted by Lowenstam (1964) on pre plate-tectonic assumptions. Figure 3 is the result of correcting his data for paleolatitudinal locations as assumed from paleomagnetic data and plate-tectonic reconstructions (Smith and Briden, 1977), including

![Figure 2](image-url)
the temperatures of Lowenstam and Epstein (1954) and adding some more data from high paleolatitudes (Savin, 1977).

A more equable temperature distribution is among the main features of the Cretaceous Earth. Barron and Washington (1982) and Barron (1983) proposed two possible extreme equator-to-pole contrasts: in the "warmest" Cretaceous a contrast of 17°C is suggested, whereas if the "coolest" Cretaceous model is valid a 26°C contrast prevailed (compare both to the 41°C contrast at present). In both models Barron and Washington (1982) point at tropical temperatures which range from 27 to 30°C. Our results seem to favor an Earth with a larger pole-to-equator thermal contrast, i.e. the "coolest" Cretaceous. A latitudinal temperature gradient which was less than half its present value was noted for the Eocene ocean by Shackleton and Boersma (1981) from the analysis of planktonic foraminifera in deep sea sediments. Their conclusion is supported by our findings.

Our findings support the concept of a more equable Cretaceous. In most general terms this also supports the popular notion of colder period showing larger latitudinal contrasts. Thus both the Cretaceous and the Eocene were both warmer and more equable than the present. The relationships depicted in Fig.3 leave, however, open the question of a more precise location of warming and cooling sites. The prevailing notion (Barron, 1983) would expect the latitudinal gradient to be roughly proportional to the temperature in high latitudes; i.e., the major temperature changes are occurring in the high latitudes, whereas the tropics remain at a more or less constant temperature. Figure 3 does not support this notion consistently. Whereas the latitudinal gradient in the Eocene is indeed the smallest, the general trend in high and low latitudes seems parallel, and there is no evidence of a larger gradient in the Maastrichtian than in the Turonian. It is premature to decide whether this pattern is just the result of too few measurements or whether it will be borne out by more data.
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