Climate- and gateway-driven cooling of Late Eocene to earliest Oligocene sea surface temperatures in the North Sea Basin

Kasia K. Śliwińska1, Erik Thomsen2, Stefan Schouten3,4, Petra L. Schoon3 & Claus Heilmann-Clausen2

During the late Eocene, the Earth’s climate experienced several transient temperature fluctuations including the Vonhof cooling event (C16n.1n; ~35.8 Ma) hitherto known mainly from the southern oceans. Here we reconstruct sea-surface temperatures (SST) and provide δ18O and δ13C foraminiferal records for the late Eocene and earliest Oligocene in the North Sea Basin. Our data reveal two main perturbations: (1), an abrupt brief cooling of ~4.5 °C dated to ~35.8 Ma and synchronous with the Vonhof cooling, which thus may be a global event, and (2) a gradual nearly 10 °C temperature fall starting at 36.1 Ma and culminating near the Eocene-Oligocene transition at ~33.9 Ma. The late Priabonian temperature trend in the North Sea shows some resemblance to the IODP Site U1404 from the North Atlantic, offshore Newfoundland; and is in contrast to the more abrupt change observed in the deep-sea δ18O records from the southern oceans. The cooling in the North Sea is large compared to the pattern seen in the North Atlantic record. This difference may be influenced by a late Eocene closure of the warm gateways connecting the North Sea with the Atlantic and Tethys oceans.

Following the Early Eocene Climatic Optimum, the Earth’s climate entered a phase of decreasing temperatures culminating at the Eocene-Oligocene transition (EOT) with the formation of large ice sheets on Antarctica (the Earliest Oligocene Glacial Maximum; EOGM)1,2. The onset of the EOGM is marked by a positive δ18O excursion in deep-sea benthic foraminifera close to the Chron C13r–C13n boundary and known as the Oi-1 event2–6. However, the δ18O values are influenced by both temperatures and volume of continental ice and the exact magnitude of the temperature decrease across the EOT is far from clear. Estimates based on various proxies indicate that the deep sea generally cooled between 3 °C and 5 °C1,7,8, while the surface waters cooled from less than 2 °C to 6 °C with large geographical variations7–10. Records of δ18O of benthic foraminifera from the southern oceans indicate that the long-term middle and late Eocene cooling was superimposed by several smaller transient temperature fluctuations. One of the most distinct of these is a brief cooling dated to ~35.8 Ma and referred to as the Vonhof cooling event11,12. The Vonhof cooling event has hitherto been observed mainly in southern oceans. Extra-terrestrial spherules present at the onset of the event at a number of these sites initially led to suggestions that the Chesapeake Bay and Popigai impacts triggered the cooling11. However, this assumption was later challenged by others (e.g. ref.13; see below). Until now the geological records of the Vonhof cooling event are scarce and the nature of the event is not fully understood.

Several recent temperature records that document the climatic changes in the Northern Hemisphere during the late Eocene to early Oligocene are at odds with the rather abrupt changes indicated by the deep-sea δ18O records from the southern oceans. A study from offshore Newfoundland shows no change across EOT10, while a study from the Greenland-Scotland Ridge indicates a gradual change stretching over nearly 3 Ma9. A recent high-resolution δ18O and δ13C benthic foraminifera study from the Atlantic Ocean, including sites from...
the southern Labrador Sea, argues for deep water formation sourcing from the Norwegian-Greenland Sea, which pre-dated the Antarctic glaciation. Several terrestrial studies, mainly from Europe, have been carried out, but they are often ambiguous as regards the magnitude and abruptness of the temperature fall during the EOT.

Here we reconstruct changes in surface water temperature (SST) in the eastern North Sea Basin during the late Priabonian to earliest Rupelian (38.6–33.5 Ma), utilizing the TetraEther index of 86 carbon atoms, TEX86 (Methods). We evaluate the TEX86-derived temperatures in relation to δ18O records measured on benthic and planktic foraminifera and compare the data from the North Sea with previously published results from the Atlantic Ocean.

The study is based on the well-calibrated middle Eocene (Barthonian) to earliest Oligocene (earliest Rupelian) succession in the Kysing-4 borehole located in the eastern part of the North Sea Basin (Fig. 1A). The site is unique, because it penetrates the most complete marine record of the upper Eocene to lowermost Oligocene in this part of the North Sea Basin. The paleolatitude of the core site during the late Priabonian was ~50°N cf. ref.27. Kysing-4 is at the moment the most northerly located site where both SST as well as benthic foraminiferal δ13C and δ18O records are available cf. ref.14. During the late Eocene, the North Sea and the Norwegian-Greenland Sea formed an elongate sea connected to the world oceans through a number of shallow gateways (Fig. 1A).

The middle and upper Eocene deposits in Kysing-4 consist of fine-grained calcareous nannofossil ooze interbedded with a sharply delimited 10 m thick unit of dark, coarser grained siliciclastic mud dated to Chron C16n.1n (Moesgaard Member) (Fig. 1B). The nannofossil ooze is overlain by a lower Rupelian unit of dark mud similar to the Moesgaard Member. The ooze is rich in planktic and benthic foraminifera and the lower part below the Moesgaard Member contains palynofacies strongly dominated by marine dinoflagellates and was probably deposited at a water depth of 300–400 m. The Moesgaard Member and the Oligocene muds lack planktic foraminifera and yield a minor amount of reworked dinoflagellate cysts (up to ~3% of the total assemblage). The palynofacies consists of mixed marine and terrestrial particles and both units are associated with sea-level falls.

Results

The Eocene calcareous nannofossil ooze is characterized by low BIT values ranging from 0.08–0.15 (Fig. 2C, Table S1) suggesting relatively low terrestrial input. Our record shows two BIT excursions with maximum values of 0.4 (Fig. 2C), which both correlate with the dark, muddy units, implying a significant rise in the riverine input from land.
Regardless of the TEX86 calibration, the sea surface temperature (SST) records show overall similar patterns (Fig. 2B,G). The TEX86–derived SST shows stable high values of ~28 °C (using the calibration in ref. 24) during the late Bartonian and early Priabonian (38.6–36.1 Ma) and low values of ~15–22 °C during the late Priabonian and early Rupelian (35.6–33.7 Ma) (Fig. 2B,G; Table S1). The temperature decline from middle Priabonian (~36.1 Ma) to earliest Rupelian (~33.7 Ma) is over 10 °C (from ~28 °C to ~15 °C). The most striking feature in our record is a distinct transient temperature minimum (of mean SST$_{TEX86H}$ = 20.5 °C or SST$_{BAYSPAR}$ = 22.5 °C) corresponding roughly with the Moesgaard Member. This dark siliceous mud unit was deposited during the early part of Chron C16n.1n over a period of approximately 100,000 years27. The C16n.1n cooling is followed by a brief recovery (to ~25 °C) in the late part of Chron 16n. From here the temperatures gradually decrease to a minimum (~15 °C or 19 °C) during the earliest Oligocene (latest Chron C13r).

The trend of the benthic foraminiferal δ¹⁸O record closely follows the fluctuations shown by the TEX86-derived SST (Fig. 2D). The TEX86-derived SST shows stable high values of ~28 °C (using the calibration in ref. 24) during the late Bartonian and early Priabonian (38.6–36.1 Ma) and low values of ~15–22 °C during the late Priabonian and early Rupelian (35.6–33.7 Ma) (Fig. 2B,G; Table S1). The temperature decline from middle Priabonian (~36.1 Ma) to earliest Rupelian (~33.7 Ma) is over 10 °C (from ~28 °C to ~15 °C). The most striking feature in our record is a distinct transient temperature minimum (of mean SST$_{TEX86H}$ = 20.5 °C or SST$_{BAYSPAR}$ = 22.5 °C) corresponding roughly with the Moesgaard Member. This dark siliceous mud unit was deposited during the early part of Chron C16n.1n over a period of approximately 100,000 years27. The C16n.1n cooling is followed by a brief recovery (to ~25 °C) in the late part of Chron 16n. From here the temperatures gradually decrease to a minimum (~15 °C or 19 °C) during the earliest Oligocene (latest Chron C13r).

The trend of the benthic foraminiferal δ¹⁸O record closely follows the fluctuations shown by the TEX86-derived SST (Fig. 2D). The planktic δ¹⁸O record also shares basic similarities with the record based on organic proxies, despite a gap across the Moesgaard Member, which is barren of planktic foraminifera (Fig. 2D,E). The δ¹³C record is virtually a mirror image of the δ¹⁸O data between 37.3 and 35.7 Ma is attributed to Milankovitch cycles. Milankovitch cycles are present in most of the core27, but they are not a part of this investigation.

Discussion

North Sea temperatures during the early Priabonian. The TEX86 SST and the benthic foraminiferal oxygen isotope records are very similar and suggest that the δ¹⁸O values were primarily controlled by the temperature (Fig. 2). However, the oxygen isotope composition of the ambient water has also a significant influence on the δ¹⁸O values of foraminiferal carbonate, and several studies indicate that the δ¹⁸O$_{water}$ of the world oceans fluctuated during the late Eocene due to the build-up of transient continental ice-sheets on Antarctica4,32. These global changes also affected the North Sea as indicated by the paleontological and sedimentological shifts that characterize the upper Eocene and lower Oligocene deposits of Kysing-4. In this study we refrain from using the δ¹⁸O values as a temperature proxy except for the lowermost sequence deposited between 37.4 and 37.1 Ma (late Chron C17n.1n) (Fig. 2B,G). This part of the succession was probably deposited in a stable open marine environment as indicated by the uniform sedimentary facies and micropaleontology27,28. We calculated water temperatures for...
in the upper part of C17n.1n to the lower part of C16n.2n (Fig. 2). The Danish hiatus can be correlated with the European–North American Bart2/Pr1 sequence16,20,37 (Fig. 3B). It is also in agreement with the early Priabonian sea surface temperatures of ~26–28 °C derived from alkenones from the northwest North Atlantic10 (Fig. 3C), with sea surface temperatures of 23–24 °C calculated for the central Greenland Sea using the TEX86 proxy37 and with mean annual ice temperatures of about 13–15 °C estimated for the adjacent land area in East Greenland on the basin of the horse lizards15 (Fig. 3B) and spore-pollen assemblages16. The temperature difference between Kysing-4 and central Greenland is smaller than today, but it is in accord with the lower Eocene latitudinal temperature gradient e.g. ref.39.

The overall agreement between the TEX86-derived SST and the terrestrial records may suggest a seasonal bias of the TEX86 proxy in our record towards summer. A similar summer bias has been observed in other TEX86 studies, although the proxy mostly is considered to represent a mean annual temperature (see discussion in ref.36). Foraminiferal δ18O records seem generally also to reflect summer temperatures40.

The EPI-1/PROM cooling event. In all investigated upper Eocene sections in Denmark, including Kysing-4, NP18 is missing or very thin indicating the presence of a hiatus41,42. In Kysing-4, the hiatus comprises the upper part of C17n.1n to the lower part of C16n.2n (Fig. 2). The Danish hiatus can be correlated with the Belgian Bassevelle 1 depositional sequence, which is of very limited extent as compared to sequences below and above43,44. The Bassevelle 1 sequence is correlated with the European–North American Bart2/Pr1 sequence of Hardenbol et al.45,46. The sea-level fall associated with the major sequence boundary at the base of the Bart2/Pr1 sequence is associated with a positive δ18O excursion recorded in Chron C17n.1n at ODP Site 689 (Southern
Ocean) and named the EPI-1 event\(^{54,46}\). More recently, a cooling event based on a benthic foraminiferal \(^{81}^{8}O\) excursion tentatively placed in Chron C17n.1n (also based on correlation with ODP Site 689) has been indicated in the Southern Ocean (ODP Site 738)\(^{47}\). The event was named the PrOM-event and appears to be the same as the EPI-1 event.

In the benthic foraminiferal \(^{81}^{8}O\) record of Kysing-4 there is no indication of a positive excursion in the preserved part of C17n.1n (Fig. 2) and it is most probable that the EPI-1/PrOM event falls within the regional hiatus comprising the upper part of C17n.1n and the lower part of C16n.2n, thus supporting previous suggestions connecting the event to a glacioeustatic sea-level fall.

**The Vonhof/C16n.1n cooling event.** The most distinct climatic event in the 4 million years long record is a transient (100,000 years) cooling dated to ~35.8 Ma (Chron 16n.1n) (Fig. 2G). The onset of the cooling is marked by a SST drop of ~5–8 °C (Fig. 2). The cooling coincides with a shift from hemipelagic nanofossil ooze to siliclastic mud (Moesgaard Member), with elevated BIT values (Fig. 2C), and with an increase in the amount of terrestrial organic particles\(^{28,30}\), altogether indicating a significant sea-level fall\(^{27–29}\) (Fig. 1). The cooling is also observed in the benthic \(^{81}^{8}O\) record, which shows an increase of about 0.7‰, closely following the trend of the TEX\(_{86}\)-derived SST (Fig. 2).

The combination of a brief sea-level fall, a drop in \(^{81}^{8}O\), and a significant decrease in SST is most simply explained by an increase in the volume of continental ice, causing a glacio-eustatic, hence global, sea level fall. Using preliminary \(^{81}^{8}O\) data a glacio-eustatic model has previously been proposed for the deposition of the Moesgaard Member\(^{27,30}\). In both of these studies, the Moesgaard Member was correlated with the Vonhof cooling event\(^{14,15}\). The Vonhof event is marked by a distinct increase in the \(^{81}^{8}O\) values of benthic foraminifera in several ODP boreholes from the high latitude southern oceans\(^33,34\). The event is less prominent in the benthic \(^{81}^{8}O\) compilation of ref.\(^48\) (Fig. 3F). We observe that sea surface temperatures of sites 925 and 336 (Fig. 3C,D), although of low resolution, also show a decrease that may correspond to the Vonhof/C16n.1n event in the North Sea.

Because of a peak of extra-terrestrial spherules at the base of the event in several ODP boreholes\(^{31,12}\) and also in the Massignano section, Italy\(^{50,51}\), the Vonhof cooling event has been linked to the Chesapeake Bay and the Popigai impact events\(^{13}\). However, the potential cooling effect of these two impacts is far from clear as the spherule layer in several cores is associated with short-term temperature rises\(^{13,52–54}\). No spherules were detected in Kysing-4.

The glacio-eustatic nature of the sea-level fall and the cooling in the North Sea Basin suggests that the event is connected to formation of ephemeral Antarctic ice sheets. Based primarily on sea-level records and on the distribution of ice-rafted debris (IRD), several studies have argued that the southern, and possibly also the northern high latitudes, experienced short-lived glaciations through most of the late Eocene and possibly also the late mid-Oligocene\(^{33,37}–38\). The Vonhof/C16n.1n event is connected to formation of ephemeral Antarctic ice sheets. Based primarily on deep-sea oxygen isotope records from the southern oceans. These records generally exhibit a relative abrupt shift with majority of the changes occurring during the EOT from ~34 to ~33.4 Ma\(^{49}\). Independent temperature proxies are few, but a study from the Kerguelen Plateau in the Southern Ocean based on the Mg/Ca temperature proxy yielded an abrupt 2–3 °C cooling in deep surface waters at ~34 Ma\(^{37}\). However, not all existing records show an abrupt shift and several studies of sea surface temperatures show a more gradual cooling trend or no cooling at all. Below we compare the TEX\(_{86}\) temperature records from Kysing-4 with some of the most detailed surface water records from the Atlantic Ocean and Norwegian-Greenland Sea (Fig. 3).

The overall cooling trend indicated in ODP Site 511 from the South Atlantic\(^9\), although of low resolution, show similarity to the pattern in Kysing-4 regarding both the time span and the magnitude (Fig. 3A,E,G). At Site 511, similar to our record, the temperature decreases by ~10 °C between ~36.3 and 33.6 Ma. The temperature record at IODP Site U1404 from the eastern North Atlantic (offshore Newfoundland), shows a very different pattern from that of Site 511 (Fig. 3E). The temperature fall is minor (~2 °C), very gradual, and there is no evidence of a surface cooling directly coinciding with the EOT. Liu et al.\(^{10}\) considered the surface water temperature records of sites 511 and U1404 as representative of the southern and northern Atlantic Ocean, respectively. The sea surface temperature record of ODP Site 913 from the central Norwegian-Greenland Sea is not considered here, as the data covering our time frame are from two independent proxies, with no overlap interval (Fig. 3B). Kysing-4 show both similarities and differences relative to the North Atlantic Site U1404 (Fig. 3A,C). The main similarities between these two sites are the overall gradual cooling patterns during the late Priabonian and the lack of a significant temperature fall across the EOT. The main difference is in the magnitude of the total cooling ranging from only ~2 °C in U1404 to at least 10 °C in Kysing-4. The differences may be related to the semi-enclosed nature of the Norwegian-Greenland Sea – North Sea system. During the Priabonian, these interconnected basins were only
connected to the outside oceans through shallow seaways (Fig. 1A), which during the Priabonian were affected by eustatic sea-level changes and plate tectonic movements (see below). A seaway corresponding approximately to the location of the English Channel today connected the Hampshire–Dieppe Basin in the southwestern North Sea Basin with the warm waters of the eastern North Atlantic (Fig. 1A). It was open during most of the Bartonian as indicated by the presence of marine sediments in the Hampshire Basin65. The Priabonian and lowermost Rupelian deposits are marginal marine and non-marine indicating that the seaway became more restricted at that time. King29,61 suggested that the connection was closed from the mid Priabonian.

A wide southeastern seaway between the North Sea and the warm Tethyan Realm was also severed during the late Priabonian (Fig. 1A) as a result of a combination of uplift of the Alpine–Carpathian foldbelt and the eustatic sea-level fall62,63. The connection between the Norwegian–Greenland Sea and the North Atlantic Ocean across the Greenland–Scotland Ridge was probably of minor importance during the late Priabonian as the sill depth over the Scotland–Greenland Ridge was only 30–50 m64. The initial separation of Greenland from the Svalbard area began at ~35 Ma65. However, the spreading zone presumably remained emerged until ~25 Ma, when shallow-water exchange became possible66.

Altogether, it appears that the connections to the warmer waters of the Atlantic Ocean and the Tethys Ocean became closed during the latest Eocene, while the shallow seaway to the Atlantic across the Greenland–Scotland Ridge remained unchanged. A connection to the Arctic Ocean at that time was apparently not yet established. We suggest that the closure of these connections may have influenced the development in the North Sea thus accentuating the temperature decrease in Kysing-4 during the late Priabonian.

The limited connection of the North Sea and the North Atlantic is supported by a new study from the southern Labrador Sea (ODP Site 647)14. However, a comparison between the two regions is difficult as the two records represent different paleodepths: 2000–3000 m in the Labrador Sea as compared to 0.300–400 m for the North Sea and planktic foraminifera at Site 647 are sparse14.

One of the more remarkable observations at Site 647 is the unusually low benthic δ18O values during the late Eocene14. They are on an average 0.5–1‰ lower than at all more southerly Atlantic sites. The late Eocene δ13C values of Kysing-4, including both the benthic and the planktic records, are significantly more positive (Fig. 2E) and are in more line with values from sites in the southern oceans. Judging from the δ13C records, it appears that the development of water masses and productivity in the two regions during the late Eocene were very different.

The δ18O records of benthic foraminifera indicate that there are also similarities between the North Sea and the Labrador Sea. The onset of the long-term cooling in the North Sea at ~36.4 Ma (Fig. 2G) coincides with the beginning of a long-term increase in the benthic δ18O values at Site 647. Increasing δ18O values are generally indicative of decreasing temperatures, but as oxygen isotopic composition is also affected by salinity, and thus the magnitude of the temperature decrease at Site 647 is uncertain. According to ref.14, the start of the increasing benthic δ18O values in the Labrador Sea coincides with the beginning of deep water formation in the northwest Atlantic and probably also with a sea surface warming. To which degree these oceanographic changes influenced the development in the North Sea remains unclarified.

Methods

Age Model. We apply the existing age model for the Kysing-4 borehole27 updated to the geological time scale GTS201228. The upper Priabonian to lower Rupelian interval in the Kysing-4 borehole is relatively condensed with a rather weak magnetic signal27 (Fig. 2). Potential uncertainties in the age model for this critical interval are evaluated below.

In the upper Priabonian deposits of Kysing-4 we observe two important nanofossil events, namely the last occurrences (LO) of the two rosette-shaped discoasters, Discosta barbadiensis and D. saipanensis27 (Fig. 2B,E,G). These two species are generally considered to disappear almost simultaneously68,69, but the extinction events have been shown to occur considerably earlier in high latitudes than in low latitudes70. In the Massignano section (Italy), the LO of the two species is in the lower third of Chron C13r51,54 at a level dated to ~34.6 Ma in the GTS201228. In Kysing-4, the LO of the group (here D. barbadiensis) occurs close to the Chron C15n-C13r boundary at a level with an estimated age of ~35 Ma. Considering the diachrony of the LO of the group, the observations in Kysing-4 are in good agreement with the observation from the Italian section and clearly supports the age model of ref.27.

Organic proxies. 20 sediment samples were collected from the interval between 6.5 m and 55.08 m. The total lipid extract was obtained from mechanically powdered and freeze-dried samples with the accelerated solvent extraction (ASE) technique using dichloromethane/methanol (9:1 [v/v]). The lipid extract was separated over a silica gel column (150 × 2.1 mm; 3 μm; Alltech, Deerfield, IL, USA) maintained at 30 °C. GDGTs were eluted with 99% hexane and 1% isopropanol for 5 min, followed by a linear gradient to 98% hexane and 2% isopropanol at a flow rate of 0.2 mL/min. Detection was achieved using single-ion monitoring. Relative qualification of the compounds was achieved by manual integration of the peaks in the mass chromatograms in the Agilent ChemStation manager software. In order to evaluate the source and the distribution of GDGTs, we calculated a number of indices: the BIT index72, % GDGT-073, the Methane Index74, the f_cren/f_cren+cren index75 and the Ring Index vs TEX67. The results imply that in all sediments ammonia-oxidizing Thaumarchaeota are the main source
of GDGT. For sea surface temperature estimations, we applied the TEX\textsubscript{86} and BAYSPAR calibrations\textsuperscript{25,26}. Out of 20 samples, nine were analysed in duplicate and two in triplicate. All the results are shown in Table S1.

The TEX\textsubscript{86} as sea surface temperature (SST) proxy. The TetraEther indeX of 86 carbon atoms (TEX\textsubscript{86}) is an organic paleothermometer, which is based on the distribution of the isoprenoid glycerol dialkyl glycerol tetraethers (isoGDGT)\textsuperscript{23}. The original definition for calculating TEX\textsubscript{86} is as follows:

$$TEX_{86} = \frac{(GDGT - 2 + GDGT - 3 + GDGT - crenarchaeol)}{(GDGT - 1 + + GDGT - 2 + GDGT - 3 + GDGT - crenarchaeol)} \quad (S1)$$

IsoGDGTs are membrane lipids spanning the cell membranes of archaea. One of the most ubiquitous isoGDGTs is crenarchaeol characterized by presence of a cyclohexane moiety. Crenarchaeol is produced by the marine archaea belonging to the phylum Thaumarcheota e.g.\textsuperscript{77,78}. The Thaumarchaeota also synthesizes other common isoGDGTs: GDGT-0 (with no cyclopentane moiety) and GDGT with 1 to 3 cyclopentane moieties. Structures of isoGDGTs are shown on Fig. S1a. Studies on Thaumarchaeota suggest that many of them are chemoautotrophs and ammonia oxidizers e.g.\textsuperscript{79,80}.

Schouten et al. (ref.\textsuperscript{23}) recognized that temperature is the main factor influencing the distribution of the sedimentary GDGTs. However, several studies recognized that the distribution of GDGTs can be influenced by other, non-thermal factors, such as: terrestrial input, oxic degradation or thermal alternation. Therefore, in order to ensure that GDGTs origin from ammonia-oxidizing Thaumarchaeota, it is important to evaluate the distribution and source of the GDGT for potential bias. For that purpose, we have utilized a number of indices.

The branched and isoprenoid tetraether (BIT) index. The BIT index is calculated as a ratio between the branched GDGTs (brGDGTs are synthesized by soil and river bacteria; for more see e.g.\textsuperscript{36} and references cited therein) versus crenarchaeol. The index values are calculated as described in ref.\textsuperscript{72}:

$$BIT = \frac{(GDGT - Ia + GDGT - IIa + GDGT - IIIa)}{(GDGT - Ia + GDGT - IIa + GDGT - IIIa + crenarchaeol)} \quad (S2)$$

The roman numerals refer to individual brGDGT structures (for details see ref.\textsuperscript{36}). Structures of brGDGTs are shown on Fig. S1b. The index aims to estimate the terrestrial input of the GDGT pool in marine environments and serves as a proxy for the relative input of soil and river organic material into marine settings\textsuperscript{22,24,41-43}. BIT values span from close to 0 (absence of brGDGTs, typical for open marine environments) to 1 (absence of crenarchaeol, characteristic for mineral soils and peat)\textsuperscript{36,72}. It is generally accepted, that TEX\textsubscript{86} estimates where BIT > 0.3 can potentially be influenced by soil-derived GDGT signal, and thus should not be used for SST reconstruction\textsuperscript{31}. However, this depends on the particular location, i.e. the TEX\textsubscript{86} value of the terrestrial GDGTs transported to the marine environment as well as the mass spectrometer settings (see discussion in ref.\textsuperscript{36}).

The BIT values are between 0.1 and 0.4, with a mean value of 0.2 (Table S1). We use a cut-off value of 0.4 and thus include all samples in further analysis.

The %GDGT-0 index. The ammonia-oxidizing Thaumarchaeota may not be the only source of GDGTs in the marine settings. GDGT-0 and smaller quantities of GDGT-1, GDGT-2 and GDGT-3 can be synthesized by other archaea including sedimentary methanogenic archaea. In some settings, the methanogenic GDGTs can be substantial e.g.\textsuperscript{84,85} and therefore can bias the TEX\textsubscript{86}. For constraining a methanogenic input of GDGTs Sinninghe Damsté et al.\textsuperscript{73} suggested applying the %GDGT-0 index:

$$%GDGT - 0 = \left(\frac{GDGT - 0}{GDGT - 0 + crenarchaeol}\right) \times 100 \quad (S3)$$

Studies on enrichment cultures of Thaumarchaeota suggest that when %GDGT-0 values reach values above 67% the sedimentary GDGT pool may be affected by an additional (probably methanogenic) source of GDGTs. Our %GDGT-0 values range between 28.8 and 48.3 with a mean value of 43 (Table S1) suggesting that the GDGT pool is most probably not influenced by methanogenic GDGTs.

The Methane Index (MI). It has also been suggested that some of the GDGTs preserved in the sediments may be produced by methanotropic Eurarchaeota (ref.\textsuperscript{36} and references cited therein). This is especially observed in settings where gas-hydrate-related anaerobic oxidation of methane is taking place\textsuperscript{34}. To identify the methanotrophic source of GDGTs, Zhang et al.\textsuperscript{74} proposed the Methane Index (MI), which is calculated using the formula:

$$MI = \frac{(GDGT - 1 + GDGT - 2 + GDGT - 3)}{(GDGT - 1 + GDGT - 2 + GDGT - 3 + crenarchaeol + crenarchaeol)} \quad (S4)$$

For SST calculations, it is recommended to exclude all samples where MI > 0.5. In our material MI varies from 0.18 to 0.25, with mean value of 0.21 (Table S1) and thus suggest no input of methanotrophic Archaea.

The Ring Index vs TEX\textsubscript{86}. Prior to calculating TEX\textsubscript{86}-SST proxy it is also crucial to eliminate samples which may have been influenced by non-thermal factors and/or deviate from modern analogues e.g.\textsuperscript{78}. To achieve that, Zhang et al.\textsuperscript{78} proposed the Ring Index (RI), which is calculated as follows:
\[ RI = 0 \times \left( \frac{GDGT - 0}{\sum GDGT} \right) + 1 \times \left( \frac{GDGT - 1}{\sum GDGT} \right) + 2 \times \left( \frac{GDGT - 2}{\sum GDGT} \right) + 3 \times \left( \frac{GDGT - 3}{\sum GDGT} \right) + 4 \times \left( \frac{crenarchaeol}{\sum GDGT} \right) + 4 \times \left( \frac{crenarchaeol'}{\sum GDGT} \right) \]  

(S5)

Where:
\[ \sum GDGT = GDGT - 0 + GDGT - 1 + GDGT - 2 + GDGT - 3 + crenarchaeol + crenarchaeol' \]  

(S6)

The formula for \( RI \) estimates a weighted average of the ring numbers in GDGT compounds. Zhang et al.\(^7\) demonstrated that in the modern core-top dataset, \( RI \) and TEX\textsubscript{86} are significantly correlated. This strong relationship is expressed as:
\[ RI_{TEX} = -0.77(\pm 0.38) \times TEX_{86} + 3.32(\pm 0.34) \times (TEX_{86})^2 + 1.59(\pm 0.10) \]  

(S7)

Zhang et al.\(^7\) furthermore suggest that TEX\textsubscript{86}-SST values deviating by more than |0.3| from the modern TEX\textsubscript{86}-RI relationship should be excluded, as they may be impacted by non-thermal factors\(^7\).\(^7\).
\[ \Delta RI = RI_{TEX} - RI_{sample} \]  

(S8)

The \( \Delta RI \) values in our dataset are between −0.17 and 0.10 with the mean value of 0 (Table S1) suggesting that the TEX\textsubscript{86} follows modern day behaviour.

The relative abundance of crenarchaeol isomer \( f_{Cren':Cren} \) In order to identify anomalous GDGT distributions O'Brien et al.\(^7\) suggested a new ratio:
\[ f_{Cren':Cren} = \frac{crenarchaeol'}{crenarchaeol + crenarchaeol'} \]  

(S9)

The ratio in our dataset is between 0.03 and 0.09 (Table S1) which is close to the lower values of the modern (0.00–0.16) core-top sediments\(^7\).

**TEX\textsubscript{86} calibration.** The first calibration of TEX\textsubscript{86} as SST proxy was linear\(^2\). Following this, Kim et al.\(^2\) presented two logarithmic calibrations, TEX\textsubscript{86\textsuperscript{H}} and TEX\textsubscript{86}\textsuperscript{L}, where TEX\textsubscript{86\textsuperscript{H}} is more applicable in high latitude settings. Considering the mid latitude setting for our site we calculated TEX\textsubscript{86\textsuperscript{H}} values using the calibration given in ref.\(^2\):
\[ TEX_{86\textsuperscript{H}} = \log \left( \frac{(GDGT-2 + GDGT-3 + crenarchaeol')}{(GDGT-1 + GDGT-2 + GDGT-3 + crenarchaeol')} \right) \]  

(S10)

Raw TEX\textsubscript{86\textsuperscript{H}} values for the studied sediments are between 0.51 and 0.72 with mean value of 0.59. Sea surface temperatures were calculated as follows:
\[ Temp[^\circ C] = 68.4 \times (TEX_{86\textsuperscript{H}}) + 38.6 \]  

(S11)

Samples analysed in duplicate show reproducibility better than 0.5 °C and in most cases better than 0.25 °C (Table S1). The residual standard error for the TEX\textsubscript{86\textsuperscript{H}} calibration model is 2.5 °C\(^2\).

One of the most recent approaches is based on a spatially varying, TEX\textsubscript{86} Bayesian regression model (BAYSPAR)\(^3\).\(^6\). BAYSPAR model SST predictions were obtained from the online GUI at http://bayspar.geo.arizona.edu using the modern-day coordinates for the Kysing site (56.0107° N, 10.2566° E). For the “deep-time” calibration we have applied the mean of the tolerance which is equal to the mean of the TEX\textsubscript{86} value (mean = 0.05973). The number of iterations to perform at each analogue site is set as default (i.e.\( \pm 2000\)). Modern analogues for our dataset show low to mid latitudinal settings (Fig. S2).

The TEX\textsubscript{86}\textsuperscript{H}-derived SST for the studied interval range between 18.8 °C and 28.8 °C (±2.5 °C) for TEX\textsubscript{86\textsuperscript{H}} (Fig. 2), 15 °C and 36 °C (±5.8 °C to 8.2 °C) for BAYSPAR (Figs 2 and S3). Regardless of the calibration, the SST record derived both calibrations shows the same trend and reveals two minima. The ΔSST between the two calibrations (ΔSST = BAYSPAR-SSTTEX\textsubscript{86\textsuperscript{H}}) is below 2.6 °C (Table S1), with mean value of −1.5 °C.

The TEX\textsubscript{86} has also been shown to be reflecting subsurface rather than SST e.g.\(^8\).\(^8\). However, since this setting is relatively shallow we assume that the trends mostly reflect upper water column conditions rather than deep water. Indeed, recent studies show that TEX\textsubscript{86} gives reasonable SST estimates with respect to other proxies such as Mg/Ca and Δ47 of planktic foraminifera\(^8\).

Finally, TEX\textsubscript{86} has been suggested to be affected by ammonium oxidation rates and/or oxygen depletion, i.e. increasing values with decreasing oxygen concentrations and ammonium oxidation rates\(^9\),\(^9\). However, since we do not find large changes in productivity and redox condition based on dinocyst assemblages, palynofacies and ichnofabric, we assume these factors did not have a large impact on our temperature trends.
Inorganic proxies. Bottom water and thermocline temperatures derived from \( \delta^{18} \)O data. Planktic foraminiferal \( \delta^{18} \)O and \( \delta^{13} \)C composition was measured on *Subbotina* sp., while the benthic values were measured on *Cibicoides eocaenicus*. *Subbotina* sp. constitutes mostly 80–100% of the planktic fauna and is the only planktic taxon which is continuously present. The second-most important planktic taxon, *Acarrinita*, is only present in three short intervals. The foraminifera were picked from the 100–500 \( \mu \)m size fractions in 52 samples of planktic foraminifera and 42 samples of benthic foraminifera. The generally well-preserved tests were crushed and ultrasonically washed in distilled water. The measurements were performed on a Finnigan MAT 253 mass spectrometer versus VPDB. The temperature reconstructions were calculated applying the equation of Shackleton:

\[
T = 16.9 - 4.38(\delta^{18}O_{\text{al}i\text{c}e} - \delta^{18}O_{\text{water}}) + 0.10(\delta^{18}O_{\text{al}i\text{c}e} - \delta^{18}O_{\text{water}})^2
\]

We estimated a \( \delta^{18}O_{\text{water}} \) value of ca. 0.3‰ based on the modern value for the study area and corrected for changes in continental ice volume. We applied a correction factor of 0.011‰ per meter sea-level change. Finally, we added 0.27‰ for conversion from the VSMOW scale to the VPDB scale.

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