Late Cretaceous temperature evolution of the southern high latitudes: a TEX$_{86}$ perspective

Lauren K. O’Connor$^1$, Stuart A. Robinson$^1$, B. David A. Naafs$^2$, Hugh C. Jenkyns$^1$, Sam Henson$^1$, Madeleine Clarke$^3$, Richard D. Pancost$^2$

$^1$Department of Earth Sciences, University of Oxford, South Parks Road, Oxford, OX1 3AN, UK. $^2$Organic Geochemistry Unit, School of Earth Sciences and School of Chemistry, University of Bristol, Bristol BS8 1RJ, UK. $^3$Department of Earth Sciences, University College London, Gower Street, London, WC1E 6BT, UK.

Corresponding author: Lauren O’Connor (lauren.oconnor@earth.ox.ac.uk)

Key Points:

- New TEX$_{86}$ data show extreme warmth in the southern high-latitudes during the Late Cretaceous
- These data indicate slow, steady cooling from the Turonian–mid-Campanian
Abstract

The Late Cretaceous was a greenhouse world, characterized by elevated temperatures and high atmospheric $p$CO$_2$. However, even in the context of an extreme greenhouse climate, existing planktic foraminiferal $\delta^{18}$O data from the Falkland Plateau (paleolatitude of ~55°S) suggest anomalous warmth, with sea-surface temperatures (SSTs) >30°C for much of the Late Cretaceous, followed by cooling. Over the last two decades there has been discussion as to whether these high $\delta^{18}$O-based SSTs reflect a genuine temperature signal and, if so, whether there was a local temperature anomaly in the South Atlantic or whether the data are representative of zonal paleotemperatures at 55°S. To provide new insights into the degree of ocean warming in the southern high-latitudes during the Late Cretaceous (Cenomanian to Campanian), new SST records from the Falkland and Kerguelen Plateaus are presented here using the organic geochemical paleothermometer TEX$_{86}$. Overall, the TEX$_{86}$ data support the $\delta^{18}$O data, indicating extreme and widespread warmth in the mid- to high southern latitudes in the Late Cretaceous, with SSTs from 27–37°C. Crucially, the TEX$_{86}$ data show slow, steady cooling from the Turonian to the Campanian, and suggest that temperature gradients during the Campanian did not become as steep as suggested by some planktic foraminiferal data.

1 Introduction

The Late Cretaceous (100.5-66.0 Ma) was a greenhouse world, characterized by high temperatures and reduced latitudinal temperature gradients (Sinninghe Damsté et al., 2010; O’Brien et al., 2017; Robinson et al., 2019). Peak warmth was attained in the Cenomanian–Turonian (~100-90 Ma) (Jenkyns et al., 1994; Norris & Wilson, 1998; Clarke & Jenkyns, 1999;
Wilson & Norris, 2001; Norris et al., 2002; Wilson et al., 2002; Voigt et al., 2004; Forster et al., 2007) with sea-surface temperatures (SSTs) reaching ≥ 30°C in the tropics and the southern mid-to high-latitudes (Huber et al., 1995, 2002, 2018; Bice et al., 2003, 2006; MacLeod et al., 2013; O’Brien et al., 2017; Robinson et al., 2019). After this interval, sea-surface and deep-water temperatures decreased to the cooler greenhouse of the Campanian–Maastrichtian (80–66 Ma) (Huber et al., 1995, 2002, 2018; MacLeod et al., 2005; Friedrich et al., 2012; Linnert et al., 2014; O’Brien et al., 2017).

However, even in the context of an extreme greenhouse climate, δ¹⁸O data from some parts of the Southern Hemisphere show anomalous warmth: in particular, estimates from Sites 327 and 511 at the Falkland Plateau (Figure 1) suggest that, at paleolatitudes of ~55–60°S, SSTs exceeded 30°C for much of the Late Cretaceous, before cooling rapidly in the mid-Campanian (Huber et al., 1995, 2018; Bice et al., 2003). By contrast, modern mean annual SSTs at ~60°S are ~0°C (NOAA 2013). If these records from Falkland Plateau are indicative of widespread and prolonged warmth in the mid- to high-latitudes, there are major implications for polar climate and latitudinal temperature gradients in a greenhouse world. These high temperatures within 15 to 20° latitude of the coast of Antarctica question the feasibility of polar ice caps and sea-ice during the Late Cretaceous (cf. Price, 1999; Li et al., 2000; Gale et al., 2002; Miller et al., 2003, 2005; Bornemann et al., 2008; Davies et al., 2009; Galeotti et al., 2009; Bowman et al., 2013). In addition, the high δ¹⁸O-based temperatures are not easily reconciled with climate models without extremely high atmospheric pCO₂ (≥3000 ppm; e.g. Poulsen et al., 1999; Bice et al., 2003, 2006). Such high pCO₂ values are not corroborated by most proxy or carbon-cycle models for the Late Cretaceous (reviewed in Foster et al., 2017).
Considering these issues, the δ¹⁸O values might not purely reflect temperature but could be, at least in part, an artefact of diagenetic alteration or depletion of local δ¹⁸Oₗ. Diagenetic alteration at Sites 327 and 511 was deemed unlikely based on the high quality of preservation of the foraminifera and the separation between planktic and benthic values, which suggests no post-mortem calcification (Huber et al., 1995, 2018; Bice et al., 2003). Factors leading to the isotopic depletion of local seawater have been similarly eliminated: there is no evidence for a drastic increase in freshwater flux to the surface ocean nor any means of sufficiently depleting the surface waters to cause such light δ¹⁸O values (Bice et al., 2003). Further, new foraminiferal δ¹⁸O data from the southern Indian Ocean (DSDP Site 258) spanning the Cenomanian–Campanian interval corroborate the isotopically light values reported for the Falkland Plateau (Huber et al., 2018). Nonetheless, a long-standing question in paleoceanography is whether the
high δ\textsuperscript{18}O-based sea-surface temperatures from the Falkland Plateau in the Late Cretaceous reflect a genuine temperature signal and, if so, is this simply a local anomaly or a more widespread feature of the climate at this time?

Here, we explore this question using an independent SST proxy: the organic paleothermometer TEX\textsubscript{86} (TetraEther indeX of 86 carbon atoms) applied to sediments from sites at the Falkland and Kerguelen Plateaus (in the southern Atlantic and Indian Oceans, respectively) spanning the Late Cretaceous interval. Based on the relationship between the distribution of archaeal membrane lipids (glycerol dialkyl glycerol tetraethers—GDGTs) in marine core-top sediments and SST (Schouten et al., 2002, 2003, 2013), TEX\textsubscript{86} has been used extensively to reconstruct SSTs during the Cretaceous (e.g. Littler et al., 2011; Naafs & Pancost, 2016; O’Brien et al., 2017; Robinson et al, 2019). Through assessing potential influences on both TEX\textsubscript{86} and foraminiferal δ\textsuperscript{18}O values, and employing a multi-proxy approach, this study investigates the temperature evolution of the Falkland and Kerguelen Plateaus during the Late Cretaceous.

2 Study areas

2.1 Falkland Plateau

Samples are used from Deep Sea Drilling Project (DSDP) Sites 327 and 511 from the Falkland Plateau (Figure 1), which is a submarine projection of the South American continental margin, extending 1800 km east of the Falkland Islands. The Falkland Plateau broke away from southern Africa during the Aptian–Albian interval and, after final separation, subsided rapidly, reaching bathyal depths sometime during the late Albian–Cenomanian (Krasheninnikov & Basov, 1983). By the late Aptian, the Plateau was relatively isolated from the influence of
continental runoff from Africa (Thompson 1977), and terrigenous sediment supply from the Andean cordillera probably slowed after the mid-Cretaceous Andean orogeny (Barker et al., 1977a). Deposition of organic-rich black shales under anoxic conditions typified the mid- to Late Jurassic but more open-marine environments with well-oxygenated sea floors and pelagic sedimentation was established on the Plateau by the early to middle Albian (Thompson, 1977). A deep connection developed between the Atlantic and Indian Oceans in the mid-Cretaceous, with an increase in Atlantic deep-water formation driving improved circulation and ventilation during the Late Cretaceous as the plateau continued to subside (Barker et al., 1977b; Robinson et al., 2010; Robinson & Vance, 2012). Paleogeographic reconstructions indicate that Sites 327 and 511 lay between ~55 and ~60°S during the Late Cretaceous (e.g. Huber et al., 1995; Hay et al., 1999; Van Hinsbergen et al., 2015). The age models developed by Huber et al. (2018) are used for both sites.

Site 327 is currently located at 50.5°S, 46.5°W on the Maurice Ewing Bank in the Falkland Plateau Basin, ~10 km northeast of Site 511. Only the Campanian section is used in this study (90–142 mbsf), which comprises 52 m of greenish-grey calcareous ooze (Barker et al., 1977b, 1977c). Foraminifera are abundant in the Campanian section, with high planktic to benthic ratios and excellent preservation, indicating deposition above the foraminiferal lysocline in a lower bathyal environment at that time (Basov and Krasheninnikov, 1983).

Site 511 is currently located at 51.0 °S, 47.6 °W in the basin province of the Falkland Plateau. The Turonian to Campanian section (~200–430 mbsf) is used in this study and comprises pale grey calcareous ooze and zeolitic foraminiferal ooze, and grey zeolitic clay and claystone (Ludwig et al., 1983). In general, the occurrence of planktic foraminifers in the Upper Cretaceous sediments is quite sporadic but, in sections with abundant foraminifera, preservation
is described as “excellent” (Huber et al., 1995), possibly owing to shallow burial depths and high clay content. Sedimentation at this site was highly episodic, with episodes of exceptionally high sedimentation rates separated by condensed intervals or hiatuses (Ludwig et al., 1983).

2.2 Kerguelen Plateau

In addition to samples from the Falkland Plateau, samples used in this study derive from Ocean Drilling Program (ODP) Sites 1138 and 1135 located on the Kerguelen Plateau (Figure 1). This plateau is a submarine Large Igneous Province, generated by the Kerguelen hot spot in the Early Cretaceous (Frey et al., 2000). Formed ~100 Mya on the Antarctic Plate, the Kerguelen Plateau has remained at a relatively constant paleolatitude of ~50–60°S since its formation, recording a transition from shallow-marine to deeper marine deposition throughout the Late Cretaceous (Meyers et al., 2009; Dickson et al., 2017).

Site 1135 is located at 59.4°S, 84.2°E on the Southern Kerguelen Province. Samples used here are of Cenomanian–Maastrichtian age (~260–530 mbsf) and comprise a sequence of white to light greenish-grey calcareous ooze and chalk (Coffin et al., 2000; Petrizzo, 2001).

Site 1138 is located at 53.3°S, 75.6°E on the Central Kerguelen Province. The samples used here are of Cenomanian–Maastrichtian age (~480–660 mbsf), and the sequence consists of cyclic alternations of white foraminifer-bearing chalk and grey to greenish grey to black intervals of nannofossil claystone (Coffin et al., 2000; Mohr et al., 2002; Dickson et al., 2017). The base of the sequence contains an organic-rich claystone (<20% TOC), indicative of low-oxygen conditions, thought to represent OAE 2 and record the Cenomanian–Turonian Boundary (Meyers et al., 2009, Dickson et al., 2017). Some TEX86 data have been published by Robinson et al. (2019) from around the Cenomanian–Turonian boundary and these data are integrated with new data from the Turonian–Santonian sediments.
Detailed foraminiferal biostratigraphy for the Turonian–Santonian at both sites (Petrizzo, 2001) and nannofossil biostratigraphy for the Cenomanian–Campanian at Site 1138 (Russo, 2014) allows correlation of the Kerguelen Plateau section with other localities. Age models for both sites have been constructed based upon combined carbon-isotope and biostratigraphy (see supplementary information for details).

3. Methods

3.1 Organic geochemistry

All samples were oven-dried overnight at 40°C to remove residual water and the edges were scraped clean using a metal spatula to remove any loose rock debris or other detritus, before being finely ground and homogenized.

The Total Lipid Extract (TLE) was obtained via solvent extraction, using either an ultrasonic bath (method described in Linnert et al., 2014) at University College London (UCL) or using an Analytix Advanced Microwave Digestion System at the University of Oxford. For the latter method, lipids were solvent-extracted from 5 g of sediment using 20 mL of DCM/methanol (9:1, v/v). Temperature in the microwave was programmed to increase linearly from room temperature to 70 °C over 10 minutes, hold at 70 °C for 10 minutes, and then cool to 25°C over 20 minutes.

At UCL, the TLE was split into polar and apolar fractions before analysis of glycerol dialkyl glycerol tetraethers (GDGTs), following the method described in Linnert et al., (2014). However, due to the low concentrations of organic matter in many of the samples, the TLE extracted at Oxford was not split before GDGT analysis (some samples were split later for analysis of the apolar fractions). Prior to analysis, the TLE (Oxford) or polar (UCL) fraction was
re-dissolved in a hexane-isopropanol mixture (99:1, v/v) and passed through a 0.45 μm polytetrafluoroethylene filter.

Analysis of GDGTs at UCL and Oxford (see supplementary information) followed the methods described in Littler et al. (2011) and Linnert et al. (2014) using an Agilent 1200 series HPLC attached to a G6130A single-quadrupole mass spectrometer (note that the same instrument was used as it was moved from UCL to Oxford). The analytical protocol followed was as described in Schouten et al. (2007). For samples analyzed in Bristol (see supplementary information), GDGTs were analysed using a ThermoFisher Scientific Accela Quantum Access triple quadrupole mass spectrometer at the Organic Geochemistry Unit (University of Bristol). Normal phase separation was achieved using two ultra-high performance liquid chromatography silica columns, following Hopmans et al., (2016). For both sample sets, the isoprenoid and branched GDGT (isoGDGTs and brGDGTs, respectively) abundances were measured in selective ion monitoring (SIM) mode ($m/z$ 1302, 1300, 1298, 1296, 1294, 1292, 1050, 1048, 1046, 1036, 1034, 1032, 1022, 1020, 1018, 744, and 653). Ion peaks of the GDGTs were integrated to determine the relative abundance of each compound. The relative abundance of some of the isoGDGTs was used to determine the TEX$_{86}$ (Schouten et al., 2002) value for each sample using the following equation:

$$TEX_{86} = \frac{isoGDGT_2 + isoGDGT_3 + cren.isomer}{isoGDGT_1 + isoGDGT_2 + isoGDGT_3 + cren.isomer}$$

In both analytical laboratories (UCL and Bristol), repeated analysis of in-house standards alongside the samples reported here suggests long-term reproducibility of TEX$_{86}$ values of ±0.05 (1σ). Comparison of TEX$_{86}$ values measured in the different laboratories suggests no significant difference between them. Other GDGT-based indices used to determine the reliability of the
TEX$_{86}$ data, e.g. BIT index (Hopmans et al., 2004), Methane Index (MI) (Zhang et al., 2011), %GDGT-0 (Sinninghe Damsté et al., 2012), and the Ring Index (ΔRI) (Zhang et al., 2011) were also calculated where possible using the relative abundances of isoGDGTs and brGDGTs. The BAYSPAR (Tierney & Tingley, 2014, 2015), TEX$_{86}$ (Kim et al., 2010) and linear (O’Brien et al., 2017) calibrations were used to convert TEX$_{86}$ to SSTs. The relative strengths and weaknesses of these calibrations for Cretaceous SSTs are discussed in O’Brien et al. (2017); we have no preference. Where TEX$_{86}$- and δ$^{18}$O-SSTs are compared, the TEX$_{86}$ calibration with the minimum offset is used.

To determine the thermal maturity of the organic matter, the C$_{31}$-homohopane distribution in representative samples was analyzed (Mackenzie et al., 1980). For this purpose, elemental sulphur was removed from the apolar fractions using activated copper. The apolar fraction was dissolved in 1 mL hexane, to which the activated copper was added until the reaction ceased, then left for 24 hours. These samples were then analyzed using a Finnigan Trace gas chromatograph-mass spectrometer (GC–MS) in the Organic Geochemistry Unit (University of Bristol). The GC oven program was: 70°C (1 min hold) to 130°C at 20°C/min, then to 300°C (held for 24 min) at 4°C/min. Separation was achieved using a Zebron non-polar column (50 m x 0.32 mm, 0.10 µm film thickness). The injection volume was 1 µl. The mass spectrometer continuously scanned between m/z 50 and 650. Using the m/z 191 trace that is characteristic for hopanes, the relative abundance of the different homologues of the C$_{31}$-homohopane was quantified as a proxy for thermal maturity (Mackenzie et al., 1980) and constrain its influence on TEX$_{86}$ values (e.g. Schouten et al., 2004).

$$C_{31} \text{ hopane ratio} = \frac{\beta \beta}{(\alpha \beta + \beta \alpha + \beta \beta)}$$
3.2 Stable isotope analysis

Samples from DSDP Sites 327 and 511 on the Falkland Plateau were also used for foraminiferal δ\(^{18}\)O analysis. Sediments were left in a beaker of deionised water for several hours until they disaggregated and were then wet-sieved, keeping the ≥63 μm fraction, which was oven-dried at 40°C overnight. For all samples, the best-preserved (glassy) specimens were selected to eliminate any obvious diagenetic influence on isotopic values. Also, the largest specimens were selected and, where possible, all from the same size fraction to minimize artefacts of size- and age-dependent isotopic values. ~20–50 μg of a single species were picked for each analysis (see supplementary information). Samples were crushed prior to analysis, but due to the small size of the samples and the risk of losing sample material, further cleaning was not conducted. Isotope measurements were performed on a Delta V Advantage isotope mass spectrometer fitted with a Kiel IV carbonate device. Oxygen isotope values are reported using the standard delta notation in parts per mil relative to VPDB. Three standards were used: NOCZ, NSB-18, and NSB-19; reproducibility (±1σ) of δ\(^{18}\)O from these standards was 0.08%, 0.03%, and 0.06%, respectively. δ\(^{18}\)O values are expressed in per mil variations relative to VPDB.

Samples from ODP Sites 1135 and 1138 were analyzed for bulk carbonate stable isotopes. Samples were oxidized to remove organic matter; H\(_2\)O\(_2\) (15%, pH 8,) was added to each sample and allowed to react for 30 minutes, then oven-dried at 40°C. Measurements were performed on a Delta V Advantage isotope mass spectrometer fitted with a Gas Bench II in the Department of Earth Sciences (University of Oxford); the carbonates were converted to CO\(_2\) with 100% H\(_3\)PO\(_4\). Two standards were used: NBS-18 and NBS-19. Repeat analysis of standards gives a standard deviation of 0.03 for δ\(^{18}\)O and 0.03 δ\(^{13}\)C, (n=15). δ\(^{18}\)O and δ\(^{13}\)C values are expressed in per mil variations relative to VPDB.
3.3 Rock-Eval

TOC content and %Mineral Carbon data were obtained using the Rock-Eval 6 Standard Analyzer at the Department of Earth Sciences (University of Oxford). ~60 mg of sediment was analyzed in the oxidation and pyrolysis ovens, by heating with incremental temperature increases from room temperature to 850°C. Mineral carbon content was calculated from the S3MINC and S5 peaks, produced from the CO and CO2 flux from the sample and analysed by infrared detector (Behar et al., 2001). The TOC content was simultaneously obtained from the combined CO and CO2 fluxes representing the Pyrolyzable Carbon (S1 + S2 + S3CO + S3CO2) and the Residual Carbon (S4CO + S4CO2). An in-house standard was regularly measured every ~10 samples. The reproducibility (±1σ) for TOC and %Mineral Carbon was 0.43% and 0.01% respectively.

4 Results and data interpretation

4.1 Falkland Plateau

4.1.1 DSDP Site 327

16 samples of Campanian age contained quantifiable GDGTs, yielding TEX86 values ranging from 0.47–0.78 (Figure 2), corresponding to TEX86-SSTs of 13–34.3°C (BAYSPAR; supplementary information). BIT values range from 0.00 to 0.43, %GDGT-0 from 17 to 82 %, Methane Index from 0.21 to 0.84, and ΔRIs from |0.03| to |1.57| (see supplementary information for methods and data). The samples with high |ΔRI| values also have high MIs and %GDGT-0 values, implying that methanogenesis was the likely cause of anomalous GDGT distributions. Samples with high BIT values may have been influenced by terrestrial-derived GDGTs. After removing samples with unacceptable BIT (≥0.4), %GDGT-0 (≥67%), Methane Index (≥0.5), or Ring Index (≥|0.3|) values, 6 samples remained with TEX86 values from 0.47–0.78. These
samples show no discernible stratigraphic trend. $T_{\text{max}}$ values were $\leq 463^\circ\text{C}$ (supplementary information). A C$_{31}$-hopane ratio was only obtained for one sample, which gave a value of 0.50.

15 samples contained sufficient foraminifera for analysis, 13 of which contained enough for the analysis of two different species. These analyses produced $\delta^{18}\text{O}_{pf}$ values of -1.12 to 0.05‰ (Figure 2). $\delta^{18}\text{O}_{pf}$ values were converted to paleotemperatures using the equation of Bemis et al., (1998). Following the methods outlined in O’Brien et al. (2017), a value of -1.27‰ VPDB was used for global $\delta^{18}\text{O}_{SW}$. A paleolatitudinal correction was also applied to the global $\delta^{18}\text{O}_{SW}$ value (Zachos et al., 1994), using a paleolatitude of 55°S. This approach yields SST estimates of 8–16°C (supplementary information). As with the TEX$_{86}$ values, these data show no discernible stratigraphic trend.

4.1.2 DSDP Site 511

From the TLEs it was possible to quantify GDGT abundances in 79 samples spanning the Cenomanian–Campanian interval. The TEX$_{86}$ values range from 0.53–0.95 (Figure 2); reconstructed paleotemperatures are generally relatively high with values from 17–44.8°C (BAYSPAR; supplementary information). BIT indices range from 0.09 to 0.76, %GDGT-0 from 0 to 96%, Methane Index from 0.18 to 0.88, and ΔRIs from |0.00| to |2.38| (supplementary information). High |ΔRI| values commonly correspond to high BIT, suggesting the influence of terrestrial GDGTs in those samples. After removing samples with unacceptable BIT, %GDGT-0, Methane Index or Ring Index values, 34 samples remained, with TEX$_{86}$ values from 0.69–0.80. These data show a steady decrease in TEX$_{86}$ from the Turonian to the Campanian. There are no trends in the other GDGT indices. $T_{\text{max}}$ values were $\leq 606^\circ\text{C}$ and C$_{31}$-hopane ratios were 0.22–0.43 (supplementary information).
15 samples contained enough planktic foraminifera for analysis and only 7 of these had enough for multiple analyses of different species. Foraminifera were sparse through large sections of this core, as reported previously (e.g. Huber et al., 1995), which limited sampling. The low abundances of foraminifera (and to some extent also GDGTs) can be attributed to the high clay sedimentation rate through sections of this core (Ludwig et al., 1983). Foraminiferal δ¹⁸O values for these samples range from -3.10 to 0.07 ‰ (Figure 2), corresponding to SSTs of 10–25.5°C. These data show no discernible trend due to the scatter in the values. The δ¹³C values exhibit no stratigraphic trend.

**Figure 2.** New TEX₈⁶ and stable-isotope data generated from DSDP Sites 511 and 327, plotted against depth. For TEX₈⁶ values, closed circles indicate reliable data and open circles indicate samples with GDGT indices exceeding acceptable thresholds (see supplementary information for details).
4.2 Kerguelen Plateau

4.2.1 ODP Site 1135

No GDGTs were detected in the Santonian-Maastrichtian, and in older sediments only 3 samples contained quantifiable GDGTs. These yield TEX\textsubscript{86} values from 0.74–0.83 (Figure 3), corresponding to SSTs of 29–35.0°C (BAYSPAR; supplementary information). These values are interpreted to reliably reflect SST, as all other GDGT indices have acceptable values (see supplementary information). Only one sample yielded quantifiable C\textsubscript{31}-hopanes and this sample had a hopane ratio of 0.53 (supplementary information).

Bulk $\delta^{18}$O values range from -3.56–0.70‰ (Figure 3), increasing steadily from the Cenomanian to the Campanian. $\delta^{13}$C\textsubscript{carb} values range from 0.82–3.29‰, showing no stratigraphic trend but two negative excursions at ~308 and 373 m.

4.2.2 ODP Site 1138

15 samples from the Cenomanian and lowest Turonian contained quantifiable GDGTs, yielding TEX\textsubscript{86} values from 0.83–0.90 (Figure 3), corresponding to SSTs of 27.2–32.8°C (BAYSPAR; supplementary information). These values are interpreted to reliably reflect SSTs, as all other GDGT indices have acceptable values (see supplementary information). TEX\textsubscript{86} decreases through time.

Bulk carbonate oxygen-isotope values range from -4.73–0.48‰ (Figure 3), increasing steadily from the Turonian to the Maastrichtian. $\delta^{13}$C\textsubscript{carb} ranges from 1.32–3.46‰. Negative $\delta^{13}$C
excursions occur at ~640 m, and ~655 m, and positive excursions occur at ~500, ~630 m, and ~650 m.

Figure 3. New TEX$_{86}$ and stable-isotope data generated from ODP Sites 1135 and 1138, plotted against depth.

5 Discussion

5.1 TEX$_{86}$ from Falkland Plateau and Kerguelen Plateau

5.1.1 Veracity

A number of samples from Sites 327 and 511 at the Falkland Plateau yielded values for various quality-control indices that suggest their TEX$_{86}$ values might not reflect temperature.
This phenomenon could be due to a range of processes, such as input of terrestrial GDGTs from soils or in situ production in the sediments by benthic (methanogenic) archaea. After removing these data points, the remaining TEX$_{86}$ data clearly indicate very high paleotemperatures at these sites during the Late Cretaceous. These data are supported by similar TEX$_{86}$ values from Sites 1135 and 1138 at the Kerguelen Plateau. In isothermal culture studies, the degree of cyclisation decreased with higher rates of ammonia oxidation; higher ammonia oxidation rates (i.e., faster growth rates) produced cooler TEX$_{86}$ temperature estimates, and low ammonia oxidation rates gave warmer temperatures. The authors attribute the correlation between TEX$_{86}$ and temperature to be actually reflecting depth in the water column, which also corresponds with ammonia oxidation. This mechanism is invoked to explain TEX$_{86}$ values indicating warmth in oxygen minimum zones (Basse et al. 2014; Hernández-Sánchez et al. 2014), where ammonia oxidation rates are low, and cold SSTs in high productivity sites such as upwelling zones (Huguet et al. 2007; Lee et al. 2008). Hurley et al. (2016) state that it is possible that the ammonia oxidation rate itself depends on temperature, making the two inseparable variables. However, there is no evidence from the lithology or GDGT ratios to indicate the presence or influence of an oxygen minimum zone on the samples used in this study.

Given the age of these sediments, it is possible that the TEX$_{86}$ values have been altered during burial and thermal degradation. However, with increasing thermal maturity TEX$_{86}$ values generally appear to be biased towards lower SST reconstructions (Schouten et al., 2004). Thermal maturity can be assessed by measuring the abundance of hopanes possessing the biological 17\(\beta\),21\(\beta\) stereochemical configuration relative to the more thermally stable 17\(\alpha\),21\(\beta\) isomers (Mackenzie et al., 1980). Hopane ratios ($\beta\beta/(\beta\beta+\alpha\beta+\beta\alpha)< 0.5$) could be associated with sufficiently high thermal maturities that TEX$_{86}$ has been biased (Schouten et al.,
Only one sample from Site 1135 yielded quantifiable C$_{31}$ hopanes, with a ratio of 0.53, indicating thermal maturity too low to impact TEX$_{86}$. Older, deeper samples from Site 1138 have previously yielded high hopane ratios (~0.5–0.8; Dickson et al., 2017). At both Sites 1135 and 1138, T$_{\text{max}}$ values of Upper Cretaceous sediments are low (≤431 °C), supporting the evidence from hopanes for low thermal maturity and no impact on TEX$_{86}$ values.

At Site 511, mid-Jurassic–Upper Cretaceous samples have previously yielded GDGTs and high hopane ratios, suggesting a generally low thermal maturity (Jenkyns et al., 2012). Intriguingly, the samples analyzed in this study from Sites 511 and Site 327 yielded hopane ratios between 0.3 and 0.5, suggesting elevated thermal maturity, which might have influenced TEX$_{86}$. However, the hopanes are present in very low abundances, and T$_{\text{max}}$ values from samples with reliable S2 values (>0.2 mg HC/g TOC; Hart & Steen, 2015; Carvajal-Ortiz & Gentzis, 2015) yield a range of values consistent with both thermal immaturity (< 400 °C) and high maturity (< 463 °C). At both sites, the available T$_{\text{max}}$ data show no discernible stratigraphic trends.

The conflicting thermal maturity indicators recorded in both Falkland Plateau sites and older sediments at Site 511 (Jenkyns et al., 2012) are difficult to reconcile, suggesting that the maturity recorded by the low-concentrations of organic-matter in the Late Cretaceous sediments may not reflect a primary burial signal, i.e. the hopanes could be associated with reworked kerogen (Handley et al., 2012; Hefter et al., 2017), though previous work suggests no evidence for organic diagenesis (Robert & Malliot, 1983). Irrespective of these different perspectives on the thermal maturity, if in situ maturity has affected the TEX$_{86}$ values, then the values measured likely represent minimum estimates of SSTs. As such, thermal diagenesis of the organic matter cannot explain the high temperatures at the Falkland and Kerguelen Plateaus. Therefore, after the
filtering for samples with unreliable index values (i.e. BIT, MI, %GDGT-0, and ΔRI), we consider the remaining TEX\textsubscript{86} values to be robust indicators of paleotemperature trends.

5.1.2 Temperature trends in TEX\textsubscript{86}

The similarity of TEX\textsubscript{86} values during the Turonian at Sites 511 and 1138 and the Santonian at Sites 511 and 1135, suggests that both the Falkland and the Kerguelen Plateau experienced broadly similar long-term trends in SST. Taken together, the data suggest that peak warmth occurred in the early Turonian, followed by a long-term cooling trend into the Santonian and Campanian (Figure 4). This trend is comparable to low- and mid-latitude TEX\textsubscript{86} records that also show a slow decline in SSTs through to the late Campanian (Forster et al., 2007; Alsenz et al., 2013; Linnert et al., 2014). Around the Cenomanian–Turonian boundary, many sites yield very similar TEX\textsubscript{86} values, as the proxy is close to the upper limit (TEX\textsubscript{86} = 1) in the low-latitudes (as also noted by Sinninghe Damsté et al., 2010; O’Brien et al., 2017; Robinson et al., 2019). For the upper Turonian to lower Campanian interval, the TEX\textsubscript{86} data from Sites 511, 1135 and 1138 are offset towards lower values compared with those from lower paleolatitude sites. By contrast, the six TEX\textsubscript{86} data points from the upper Campanian at Site 327 do not present a consistent pattern. To explore this inconsistency further, it is useful to contextualize these data by comparison with other late Campanian TEX\textsubscript{86} records. One data point from Site 327 is almost as high as data from Shuqualak, which lies at \(\sim35^\circ\)N (Linnert et al., 2014), whereas other samples from Site 327 have TEX\textsubscript{86} values as low as those from the PAMA Quarry in Israel (0° latitude; Alsenz et al., 2013). It has been suggested that the depositional environment at the PAMA Quarry was influenced by upwelling (Alsenz et al., 2013). Such an effect may explain why the PAMA Quarry yields lower TEX\textsubscript{86} values than Shuqualak (Linnert et al., 2014), that are, perhaps coincidentally the same as those recorded at Site 327. The one sample from Site 327 with a
TEX\textsubscript{86} value similar to Shuqualak is not easily explained and the limited size of the dataset from Site 327 prohibits a confident interpretation of late Campanian SSTs at the Falkland Plateau.

**Figure 4.** Compilation of benthic foraminiferal $\delta^{18}O$, bulk carbonate $\delta^{18}O$, planktic foraminiferal $\delta^{18}O$, and TEX\textsubscript{86} data. All oxygen isotope values are reported VPDB. $^1$Huber et al., 2018; $^2$Huber
et al., 1995; Barrera & Savin, 1999; Friedrich et al., 2009; Barrera & Huber, 1990; Falzoni et al., 2016; Clarke and Jenkyns, 1999; Ando et al., 2013; Woelders et al., 2018; Bice et al., 2003; Forster et al., 2007; Schouten et al., 2003; Bornemann et al., 2008; Sinninghe Damsté et al., 2010; Robinson et al., 2019; Alsenz et al., 2013; van Helmond et al., 2014; Linnert et al., 2014; van Helmond et al., 2015; Woelders et al., 2017. All age models from original sources, plotted on the 2016 timescale. Bulk carbonate and fine-fraction δ¹⁸O values are plotted with the benthic foraminifera δ¹⁸O values as the carbonate is assumed to have either a bottom-water diagenetic overprint and/or offsets related to vital effects in the calcareous nannoplankton that contribute much of carbonate to the sediment (e.g. Falzoni et al., 2016).

5.2 δ¹⁸O values from Falkland Plateau and Kerguelen Plateau

The new Falkland Plateau δ¹⁸Oₚf data presented here are within the range of previous values (Huber et al., 1995; Barrera & Savin, 1999; Fassell & Bralower, 1999; Bice et al., 2003), albeit with slightly lower maximum temperatures. The most negative δ¹⁸Oₚf values from Site 511 are as high as those at the Exmouth Plateau, ~10° further north, and the most positive values are comparable to benthic values from sites in both the Atlantic and Tethys (Figure 4). At Site 511, there is an apparent warming event in the Turonian (~91.5 Ma in Figure 4), which has previously been the focus of discussion by Bice et al., 2003. The absence of a similar event in other records (Figure 4) suggests that this might be a consequence of local processes (e.g. local warming, variable preservation and/or species effects). Unfortunately, insufficient GDGTs are preserved in the same interval to test whether SSTs increased locally at that time.

The combined δ¹⁸Oₚf data from DSDP Sites 327, 511, and 690 (Weddell Sea) show an apparent drastic cooling at ~76 Ma—a trend not seen at other locations. However, this shift
arises largely from a change from samples from the Falkland Plateau to the Weddell Sea, which lies ~10° to the south. The individual sites only show long-term steady cooling.

At Site 1138, bulk-carbonate δ¹⁸O exhibits an apparently sharp increase of about 2‰ in the earliest Turonian following OAE 2, and then a steady increase through to the Campanian (Figure 4). The long-term trend is also observed at Site 1135. The most negative δ¹⁸O at Site 1138 could either be a response to peak warmth in the Early Turonian (e.g. Jenkyns et al., 1994; Robinson et al., 2019) or a consequence of the transition of the seafloor (and hence the diagentic environment) from relatively warm, shallow-water depths in the earliest Turonian to cooler, deeper water environments as the Plateau subsided. This event is not seen in other records (including in the TEX₈₆ record from the same site; Robinson et al., 2019) and suggests that it might be predominantly a local environmental signal. However, the gentle increase in δ¹⁸O values from the Turonian to Campanian at Sites 1135 and 1138 shows a strong similarity to planktic foraminiferal, benthic foraminiferal, and bulk carbonate δ¹⁸O values from the Exmouth Plateau, approximately 10° further north (Clarke and Jenkyns, 1999; Falzoni et al., 2016) and is thus interpreted to reflect dominantly bottom-water temperature trends. These data further suggest slow, steady cooling through the Late Cretaceous.

5.3 Comparison of trends in δ¹⁸O with TEX₈₆

The overall Late Cretaceous TEX₈₆ trend from the Falkland and Kerguelen Plateaus is consistent with the slow, steady cooling inferred from benthic foraminiferal δ¹⁸O and most planktic foraminiferal and bulk carbonate δ¹⁸O records from the South Atlantic and Indian Oceans (Figure 4). However, the evolution of surface-water temperatures through time does show some important differences between both proxies and localities.
For much of the Turonian–mid-Campanian, the benthic compilation, the Indian Ocean planktic foraminiferal $\delta^{18}O$ record from Site 762, and the combined TEX$_{86}$ record from Sites 511, 1135 and 1138 suggests a broadly similar steady trend towards a cooler climate (Figure 4). This trend broadly follows the proxy and model reconstructions of long-term atmospheric CO$_2$ evolution (e.g. Berner 2006; Foster et al., 2017), which indicate a decrease in its concentration from the mid- to latest Cretaceous. During this same interval, the planktic foraminiferal $\delta^{18}O$ record from Site 511 shows, on average, a similar pattern of cooling but displays considerable scatter and an interval (e.g. ~81 to 78 Ma) in which the lightest $\delta^{18}O$ values are a few per mil greater than those before or after. The scatter in $\delta^{18}O_{pf}$ at Site 511 is largely due to the inclusion here of all planktic species, irrespective of depth habitat. Although the depth habitats of some common taxa have been suggested by cross-plotting carbon and oxygen isotopes, others, most notably Heterohelix, exhibit inconsistent behavior (Huber et al., 1995). The apparent cooling in the Campanian (at ~76 Ma) is manifested within the shallow dwelling planktics, which would suggest that this event is not an artifact of changing depth habitats, yet the absence of such an event elsewhere (or in the, albeit limited, TEX$_{86}$ data from Site 511), suggests another explanation is required. One possibility is that the relative paucity of $\delta^{18}O_{pf}$ data through this interval, or subtle differences in carbonate preservation, may be impacting the apparent temperature signal from this site. If this interval is excluded, then the lightest $\delta^{18}O$ values from Site 511 support a broadly smooth cooling trend from the Turonian to the mid-Campanian.

From the mid-Campanian onwards, commonality in trends becomes harder to discern. In the benthic and bulk $\delta^{18}O$ records there is a slight trend towards heavier values, which is also seen in the planktic record from Site 690 in the South Atlantic. In the Indian Ocean at Site 762, planktic $\delta^{18}O$ values appear to show some minor variability over this time. The TEX$_{86}$ records
from the Southern Hemisphere do not, unfortunately, capture this time interval well and the record from Site 327 is, as discussed previously, difficult to interpret. Intriguingly, the planktic foraminifera from Site 327 of late Campanian age suggest cooler conditions on the Falkland Plateau compared to the immediately older samples from Site 511 at ~77 Ma. A similar drop in temperatures is not observed in any other records from the Southern Hemisphere. The range of $\delta^{18}O$ planktic values from Site 327 is less than that from Site 511 and, counter-intuitively, the absolute $\delta^{18}O$ values are as positive as, or more positive than, those observed at Site 690, ~10° further south. Surprisingly, our observations indicate that the preservation of the planktic foraminifera from Site 327 is actually better than that at Site 511 (Figure 5), suggesting that diagenesis cannot explain the positive $\delta^{18}O$ values. With all of these uncertainties regarding Site 327, we suggest that the evolution of temperature in the late Campanian was likely not marked by a step-changes in temperature and that for the Falkland Plateau the evolution of temperature from the late Campanian–Maastrichtian remains unresolved.

Figure 5. Photographs of planktic foraminifera showing excellent preservation. Scale = 0.1 mm. a) *A. australis* from DSDP Site 327 Core 11 Section 1 50–53 cm; b) *H. planata* from DSDP Site 327 Core 11 Section 1 50–53 cm; c) *M. marginata* from DSDP Site 511 Core 44 Section 5 50–53 cm; d) *H. globulosa* from DSDP Site 511 Core 44 Section 5 50–53 cm.
5.4 Absolute SST value comparison from Site 511 and implications

The absolute values of TEX\textsubscript{86}-SSTs from the Falkland Plateau (using all three calibrations) are close to the maximum SSTs suggested by planktic foraminiferal $\delta^{18}$O-SST records from the Southern Atlantic and Southern Tethys (e.g. Huber et al., 1995, 2018; Falzoni et al., 2016), and provide further evidence supporting the existence of extremely warm temperatures at southern paleolatitudes of ~55 to 60°S in the Late Cretaceous (e.g. Huber et al., 1995, 2018; Falzoni et al., 2016). Nonetheless, the SSTs calculated from TEX\textsubscript{86} at Site 511 are slightly warmer than the corresponding $\delta^{18}$O values (Figure 6). Unfortunately, only a few samples (n=15) yielded both sufficient GDGTs and planktic foraminifera to directly compare absolute temperatures, but these also indicate that TEX\textsubscript{86}-derived SSTs are warmer than those derived from $\delta^{18}$O values. However, such an offset is not unprecedented. In the Cenozoic, TEX\textsubscript{86} generally agrees well with the trends and absolute temperatures reconstructed from alkenones (e.g. $U^{15}_3$), foraminiferal $\delta^{18}$O and Mg/Ca ratios, and clumped isotopes ($\Delta_{47}$) of carbonates, though commonly showing a slight offset to warmer SSTs (Zachos et al., 2006; Huguet et al., 2006; Menzel et al., 2006; Pearson et al., 2007; Bijl et al., 2009, 2010; Hollis et al., 2009, 2012; Casteñada et al., 2010; Richey et al., 2010; Keating-Bitonti et al., 2011; Shintani et al., 2011; Lopes dos Santos et al., 2013; Douglas et al., 2014; Meyer et al., 2018; Naafs et al., 2018; Woelders et al., 2018). The causality of these higher temperatures remains unresolved, but has been attributed by many authors to a summer bias of the Thaumarchaeota or a winter bias of the other proxies.
Figure 6. A comparison of benthic and upper ocean temperatures at DSDP Site 511, calculated from $\delta^{18}O$ and TEX$_{86}$. Planktonic foraminifera include all taxa, irrespective of depth habitat. See 4.1.1. for description of approach used to calculate temperatures from oxygen-isotopes. SSTs were calculated from TEX$_{86}$ using the following empirical calibrations: TEX$_{86}^H$ (Kim et al., 2010), TEX$_{86}$-linear (O’Brien et al., 2017) and the deep-time version of BAYSPAR (Tierney & Tingley, 2014, 2015). Age model from Huber et al. (2018).
TEX$_{86}$ is assumed to reflect mean annual SST, but as nutrient availability, food-web activity, and temperatures vary throughout the year, seasonality in Thaumarchaeotal abundance and export can result in bias. Thaumarchaeota and their GDGTs have been found to exhibit pronounced seasonal variations in the modern ocean, although the timing of peak abundance varies significantly both geographically and environmentally. Observations from sediment traps in the Arabian Sea (Wuchter et al., 2005) and the Mediterranean Sea (Castañeda et al., 2010) indicate a warm (summer) bias in TEX$_{86}$, whereas data from the North Sea (Wuchter et al., 2005; Pitcher et al., 2011) and Santa Barbara Basin (Huguet et al., 2007) indicate a cold (winter) bias. This effect is exacerbated by the fact that GDGTs almost certainly are produced in the shallow subsurface associated with the zone of maximum ammonia oxidation (Hurley et al., 2018). The strong seasonality in sunlight at high latitudes may further exacerbate any seasonality in nutrient regimes, and thus TEX$_{86}$ in the water column. Nonetheless, despite the indications for highly seasonal GDGT production, sedimentary TEX$_{86}$ values both at a regional (e.g. North Sea; Pitcher et al., 2011) and global scale suggest that the best correlation is obtained between TEX$_{86}$ and mean annual SSTs (Kim et al., 2010).

Foraminiferal tests can undergo alteration via dissolution, recrystallization, and/or the addition of diagenetic calcite overgrowths, all of which have the potential to compromise the $\delta^{18}$O signal. Calcite precipitated from pore waters during water column, sea-floor and shallow burial recrystallization are generally more enriched in $^{18}$O than the original planktonic tests (e.g. Pearson et al., 2001; Pearson 2012), equating to lower paleotemperatures. Despite efforts to select only the best-preserved (glassy) specimens for isotopic analysis, there is a possibility that there is some degree of diagenetic influence on some tests.
Planktic foraminifera can live over a range of depths—from the surface mixed layer to the (sub)thermocline—and therefore over a range of temperatures. The analysis of species from only a certain depth habitat may bias temperature estimates. Foraminifera can change depth habitat over their life cycle; in the modern ocean, many planktic species descend through the water column at maturity to reproduce at greater depths, typically in the upper thermocline (e.g. Hemleben et al., 1989). This behavior can present problems when using foraminiferal $\delta^{18}O$ for SST reconstructions as the descent through the water column can bias temperature estimates to the colder values at depth. Ontogeny can also bias $\delta^{18}O$ values, as the metabolic rates of foraminifera were found to change throughout their life cycle (Spero & Lea, 1996), with decreasing $^{18}O$ depletion with age. If fractionation and depth habitat change with age, the $\delta^{18}O$ signal can be offset from SST. As such, this ontogenetic effect can skew temperature estimates towards the more abundant size fraction/age.

While there are many caveats to the use of both TEX$_{86}$ and foraminiferal $\delta^{18}O$, with appropriate screening both are valuable tools for paleoclimate reconstruction. However, the contrast in reconstructed temperatures and trends at Site 511 suggests that (at least) one of the proxies at this location does not record mean annual SST. The Site 511 TEX$_{86}$ values also exhibit a strong similarity to the Site 511 benthic record as well as the global compilation of TEX$_{86}$, planktic $\delta^{18}O$, and benthic $\delta^{18}O$ records, suggesting that the TEX$_{86}$ record accurately reflects long-term regional and global climatic trends. As noted previously, although isoprenoid GDGTs are produced throughout the water column (Shah et al., 2008), modern sedimentary TEX$_{86}$ values correlate best with SSTs suggesting that the dominant signal in sediments is exported from the surface ocean (e.g. Schouten et al., 2013). Thus the similarity in trends at Site 511 between benthic temperatures and TEX$_{86}$ reflects temperature throughout the water column. In contrast to
the TEX$_{86}$ and benthic records, the planktic $\delta^{18}$O values show much more temporal variability, likely reflecting short-term changes in climate and environment, and/or the effects of subtle diagenesis and under sampling.

In general, all proxy data indicate very high SSTs in the southern mid/high latitudes during the Late Cretaceous. If correct, then why do models struggle to reproduce such warmth and reduced latitudinal gradients? A long-standing problem in paleoclimatology is the discrepancy between proxy data and model simulations for high latitudes in a greenhouse world. Proxy records indicate equable conditions and reduced latitudinal temperature gradients (e.g. Huber et al. 1995; Huber et al. 2002; Amiot et al. 2004; Pearson et al. 2007; Hollis et al. 2009; Sinninghe Damsté et al. 2010; Keating-Bitonti et al. 2011; Naafs et al., 2018), whereas models indicate cooler polar conditions and steeper latitudinal gradients (Barron 1987; Huber & Sloan 1999, 2001; Bice et al. 2006; Huber & Caballero 2011; Lunt et al. 2012; Upchurch et al. 2015; Lunt et al. 2016; Tabor et al. 2016).

Warming appears to be amplified in the polar regions, particularly in periods of extreme warmth (Huber et al. 2000; Johannessen et al. 2004; Sluijs et al. 2006; Miller et al. 2010; Lee 2014), leading to reduced latitudinal temperature gradients. Climate modelling demonstrates that the contributions to polar amplification include the greenhouse effect of CO$_2$ (Manabe & Wetherald 1980), additional water vapor (Graversen & Wang 2009), cloud feedbacks (Sloan & Pollard 1998; Vavrus 2004; Kump & Pollard 2008; Kiehl & Shields, 2013; Sagoo et al., 2013), methane concentration (Kirk-Davidoff et al. 2002; Schmidt & Shindell 2003; Bice et al. 2006), vegetation–climate interactions (Zhou et al. 2012), and/or changes in atmospheric (Manabe & Wetherald 1980) and oceanic heat transport (Holland & Bitz 2003; Khodri et al. 2001; Spielhagen et al. 2001). One study found that at high temperatures the greatest contribution to
polar amplification derives from temperature feedbacks wherein more energy is radiated back into space from the tropics than the poles (Pithan & Mauritsen 2014). However, the relative influence of these drivers is debated, and these studies suggest that no single mechanism can produce the temperatures and latitudinal gradients such as those suggested for the Cretaceous and early Paleogene, but rather a complex interplay of multiple drivers is required. For example, through the addition of CH$_4$ and stratospheric water vapor, Schmidt & Shindell (2003) were able to reconcile global warmth of the PETM without a significant CO$_2$ increase. Other studies maintain that methane production must be consistently high for stratospheric water vapor to drive warming (Sloan & Pollard 1998; Kirk-Davidoff et al. 2002).

It appears that previous models of greenhouse climate lack the physio-chemical inputs that allow for the magnitude of polar amplification recorded by a range of temperature proxies. With sufficiently high CO$_2$, models are increasingly showing agreement with proxy records, especially terrestrial data (e.g. Huber et al., 2011; Lunt et al. 2012), but the mechanisms vary between different models. Comparisons with SST data are also improving, though problems remain with high latitudes; for example, $\delta^{18}$O- and TEX$_{86}$-SST uncertainties only just overlap with that of the model for the Eocene (Lunt et al. 2012). However, if seasonality is an important component of some Late Cretaceous proxy reconstructions then this could lead to discrepancies in model-data comparisons that cannot be easily accounted for without better approaches to determining the seasonal influence on planktic organisms in the geological record.

6. Conclusions

New TEX$_{86}$- and $\delta^{18}$O-SST estimates from the Falkland Plateau confirm previous studies indicating extreme warmth through the Late Cretaceous at the high southern latitudes, suggesting
that the depleted δ¹⁸O values cannot simply be ascribed to diagenesis or anomalously light δ¹⁸Osw. The compilation of data from the Falkland and Kerguelen Plateaus indicates high SSTs across the southern high-latitudes, and steady cooling from the Turonian to Campanian. The extreme warmth in the southern high latitudes suggests that latitudinal gradients remained low throughout the Late Cretaceous, and that polar ice caps at low altitudes are hence deemed unlikely.

Because both the TEX₈⁶ and δ¹⁸Oₚf data from Site 327 on the Falkland Plateau are interpreted to be problematic and are therefore excluded at this point, it is suggested that the evolution of temperature in the late Campanian was likely not marked by any step-changes in temperature and that, for the Falkland Plateau, the exact pattern of climate change from the late Campanian–Maastrichtian remains unresolved. However, with so little data, and from so few sites, it is difficult to determine precise climate dynamics in the Late Cretaceous southern high-latitudes, impeding the accurate modelling of global climate at this time.

When compared to other data from a similar paleolatitude, it appears that the Falkland Plateau SSTs are not especially anomalous, indicating a global forcing mechanism, probably a combination of elevated atmospheric CO₂, CH₄, and water vapor. However, the precise tectonic and volcanic mechanisms by which CO₂ remained high, the sources and amount of atmospheric methane, and the role of the hydrological cycle in greenhouse climates, remain issues that, although debated over decades, are still to be fully resolved. This problem highlights the fact that major uncertainties exist in our understanding of Late Cretaceous carbon cycling and climate and the operation of the Earth system in a greenhouse world.
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**Figure 1.** DSDP/ODP Sites used in this study (stars) and previously published data (circles). Adapted from Blakey 2016.

**Figure 2.** New TEX$_{86}$ and stable-isotope data generated from DSDP Sites 511 and 327, plotted against depth. For TEX$_{86}$ values, closed circles indicate reliable data and open circles indicate samples with GDGT indices exceeding acceptable thresholds (see supplementary information for details).

**Figure 3.** New TEX$_{86}$ and stable-isotope data generated from DSDP Sites 511 and 327, plotted against depth.

**Figure 4.** Compilation of benthic foraminiferal $\delta^{18}$O, bulk carbonate $\delta^{18}$O, planktic foraminiferal $\delta^{18}$O, and TEX$_{86}$ data. All oxygen isotope values are reported VPDB. $^1$Huber et al., 2018; $^2$Huber et al., 1995; $^3$Barrera & Savin, 1999; $^4$Friedrich et al., 2009; $^5$Barrera & Huber, 1990; $^6$Falzoni et al., 2016; $^7$Clarke and Jenkyns, 1999; $^8$Ando et al., 2013; $^9$Woelders et al., 2018; $^{10}$Bice et al., 2003; $^{11}$Forster et al., 2007; $^{12}$Schouten et al., 2003; $^{13}$Bornemann et al., 2008; $^{14}$Sinninghe Damsté et al., 2010; $^{15}$Robinson et al., 2019; $^{16}$Alsenz et al., 2013; $^{17}$van Helmond et al., 2014; $^{18}$Linnert et al., 2014; $^{19}$van Helmond et al., 2015; $^{20}$Woelders et al., 2017. All age models from original sources, plotted on the 2016 timescale. Bulk carbonate and fine-fraction $\delta^{18}$O values are plotted with the benthic foraminifera d18O values as the carbonate is assumed to have either a bottom-water
diagenetic overprint and/or offsets related to vital effects in the calcareous nannoplankton that contribute much of carbonate to the sediment (e.g. Falzoni et al., 2016).

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**Figure 6.** A comparison of benthic and upper ocean temperatures at DSDP Site 511, calculated from $\delta^{18}O$ and TEX$_{86}$. Planktonic foraminifera includes all taxa, irrespective of depth habitat. See text for description of approach used to calculate temperatures from oxygen-isotopes. SSTs were calculated from TEX$_{86}$ using the following empirical calibrations: TEX$_{86}^H$ (Kim et al., 2010), TEX$_{86}$-linear (O’Brien et al., 2017) and the deep-time version of BAYSPAR (Tierney and Tingley, 2014, 2015). Age model from Huber et al. (2018).