

Extreme polar warmth during the Cretaceous greenhouse? Paradox of the late Turonian $\delta^{18}\text{O}$ record at Deep Sea Drilling Project Site 511

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[1] Oxygen isotope data for upper Turonian planktonic foraminifera at Deep Sea Drilling Project Site 511 (Falkland Plateau, 60°S paleolatitude) exhibit an $\sim 2\text{‰}$ excursion to values as low as -4.66‰ (Vienna Pee Dee belemnite standard; PDB) coincident with the warmest tropical temperature estimates yet obtained for the open ocean. The lowest planktonic foraminifer $\delta^{18}\text{O}$ values suggest that the upper ocean was as warm as 30–32°C. This is an extraordinary temperature for 60°S latitude but is consistent with temperatures estimated from apparently coeval mollusc $\delta^{18}\text{O}$ from nearby James Ross Island (65°S paleolatitude). Glassy textural preservation, a well-defined depth distribution in Site 511 planktonics, low sediment burial temperature ($\sim 32^\circ\text{C}$), and lack of evidence of highly depleted pore waters argue against diagenesis (even solid state diffusion) as the cause of the very depleted planktonic values. The lack of change in benthic foraminifer $\delta^{18}\text{O}$ suggests brackish water capping as the mechanism for the low planktonic $\delta^{18}\text{O}$ values. However, mixing ratio calculations show that the amount of freshwater required to produce a 2‰ shift in ambient water would drive a 7 psu decrease in salinity. The abundance and diversity of planktonic foraminifera and nannofossils, high planktonic:benthic ratios, and the appearance of keeled foraminifera argue against lower-than-normal marine salinities. Isotope calculations and climate models indicate that we cannot call upon more depleted freshwater $\delta^{18}\text{O}$ to explain this record. Without more late Turonian data, especially from outside the South Atlantic basin, we can currently only speculate on possible causes of this paradoxical record from the core of the Cretaceous greenhouse. **INDEX TERMS:** 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation; 4267 Oceanography: General: Paleoceanography; **KEYWORDS:** carbon dioxide, Cretaceous, Turonian, oxygen isotopes, modeling, salinity

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1. Introduction

[2] Mid-Cretaceous upper ocean temperature estimates from oxygen stable isotope ($\delta^{18}\text{O}$) analyses of planktonic foraminifera are typically 8–20°C warmer than modern sea surface temperatures (SST) [Bice and Norris, 2002]. Terrestrial plant assemblages from the high latitudes of both hemispheres also indicate warmer-than-modern mean annual and winter minimum temperatures for this interval [Herman and Spicer, 1996]. Fossil vertebrate assemblages from $\sim 72^\circ\text{N}$ paleolatitude are consistent with mean annual temperatures $>14^\circ\text{C}$ in the Turonian-Coniacian [Tarduno *et al.*, 1998]. Yet, among data supporting a warm mid-Cretaceous earth, paleotemperature estimates from the high-

latitude late Turonian at Deep Sea Drilling Program (DSDP) Site 511 (Falkland Plateau, Figure 1) stand out as problematic. Huber *et al.* [1995] reported $\delta^{18}\text{O}$ values as low as -4.39‰ relative to Vienna Pee Dee belemnite (PDB) from exquisitely preserved planktonic foraminifera. The data reported by Huber *et al.* indicate upper ocean waters as warm as 28–30°C at $\sim 60^\circ\text{S}$ paleolatitude in the late Turonian. Today, such temperatures are typical of the tropical western Pacific warm pool and so seem extraordinary for high latitudes. These estimates might therefore be written off as an artifact of unrecognized diagenetic alteration or unexplained extreme isotopic depletion of local water. However, Wilson *et al.* [2002] recently presented tropical planktonic $\delta^{18}\text{O}$ values consistent with temperatures 6–8°C warmer than modern tropical SSTs from DSDP Site 144 on Demerara Rise ($\sim 2\text{--}5^\circ\text{N}$ paleolatitude, Figure 1) in a late Turonian sequence that is broadly coeval with the



Figure 1. South Atlantic Late Cretaceous paleogeographic reconstruction showing the sites discussed in the text. The 92 Ma reconstruction is from the PLATES Project of the University of Texas Institute for Geophysics [Lawver *et al.*, 2001]. KP, Kerguelen Plateau; EANT, East Antarctica; DSDP, Deep Sea Drilling Project.

anomalous Site 511 record. Recognition of a late Turonian hyperthermal period in the tropics provides motivation to re-examine the late Turonian sequence on Falkland Plateau. Here we present new Site 511 isotope data from planktonic and benthic foraminifera and the possible implications of the record.

2. Material and Methods

[3] DSDP Site 511 (51°00.28'S, 46°58.30'W) is located on continental crust on Falkland Plateau at 2600 m water depth. To obtain a detailed record across the negative $\delta^{18}\text{O}$ shift reported for Turonian sediments by Huber *et al.* [1995], we sampled at 5–75 cm intervals in core 47 and the upper part of core 48, which were recovered from between 404.31 and 413.83 m below seafloor (mbsf). The sediments are unconsolidated, primarily grey to dark grey clay stone with a calcium carbonate content ranging from 10 to 23% [Ludwig *et al.*, 1983; Huber *et al.*, 1995]. Sediments in core 47-1 and upper 47-2 are carbonaceous and tended to yield no foraminifera or too few individuals for stable isotope measurements. Below 47-2, 33 cm, the planktonic foraminiferal and calcareous nannofossil assemblages include cosmopolitan species that are associated with normal, open marine depositional settings, an environmental interpretation supported by the sedimentary trace element geochemistry [Varentsov, 1983].

[4] A late Turonian age is assigned to these samples on the basis of co-occurrence of the calcareous nannofossil *Kamptnerius magnificus* and the planktonic foraminifera

Marginotruncana marginata (= *Globotruncana bulloides* of Huber *et al.* [1995]) and *Praeglobotruncana stephani* in the upper 40 cm of section 511-48-1 [Wise, 1983; this study]. In Tethyan sections the first occurrences of *K. magnificus* and *M. marginata* are recorded in the late Turonian [Roth, 1978; Premoli Silva and Sliter, 1994] and the last occurrence of *P. stephani* in the late middle Turonian [Premoli Silva and Sliter, 1994]. Strontium isotope ratios are also consistent with a late Turonian age (below, section 4).

[5] The location of the Cenomanian/Turonian boundary below late Turonian sample 48-1, 76 cm (414.26 mbsf) is uncertain because of an 11-m sequence of red clay stone that contains non-diagnostic agglutinated foraminifera and mollusk and echinoderm fragments and is barren of calcareous foraminifera and nannofossils. The absence of calcareous microfossils may be due to a shallow calcite compensation depth [Wise, 1983]. Planktonic foraminifera reappear farther downcore, in sample 49-2, 25 cm, where they are rare relative to benthic foraminifera, but their abundance increases to 95% of the foraminiferal assemblage by sample 49-5, 25 cm. Planktonic foraminifera diagnostic of the late Albian, including *Rotalipora ticensis* and *Rotalipora* cf. *appenninica*, occur 40 cm below this sample. No biomarker species have been recorded between this level and the upper Turonian identified in core 48-1.

[6] Samples were dried, soaked in distilled water with Calgon, and washed through a 38 μm sieve. Disaggregation of clay required drying/washing the samples twice. The samples below 405 mbsf contain abundant, small planktonic foraminifera and rare benthic foraminifera. Carbon and oxygen stable isotope measurements were made on monospecific samples of the planktonic foraminifera *Globigerinelloides volutus*, *Heterohelix moremani*, and *Hedbergella* sp. from the 106–180 μm fraction, and *Whiteinella baltica*, *Archaeoglobigerina bosquensis*, *Marginotruncana marginata*, and *Praeglobotruncana stephani* from the 212–300 μm fraction. Each sample contained 30–100 individual planktonic tests in order to assure adequate sample mass, with the exception of *M. marginata* and *P. stephani* samples, which contained 3–8 of these larger individuals. Measurements were also made of the benthic foraminifera *Gyroidinoides globosus*, *Berthelina* spp. and *Dentalina* sp., as well as groups of mixed benthics and shallow infaunal benthic specimens. Stable isotope analyses were performed at Woods Hole Oceanographic Institution on a Micromass PRISM-II spectrometer with common acid bath. Precision of the measurements is $\pm 0.03\%$ for carbon and $\pm 0.05\%$ for oxygen. All carbonate isotope data here are reported relative to the Vienna Pee Dee belemnite (PDB) isotope standard.

3. Results

3.1. Foraminiferal Preservation and Calcareous Plankton Assemblages

[7] Calcareous benthic foraminifera typically have glassy, translucent tests throughout core 47. Glassy specimens of planktic foraminifera can also be found throughout the core when viewed as dry specimens, but constitute the majority

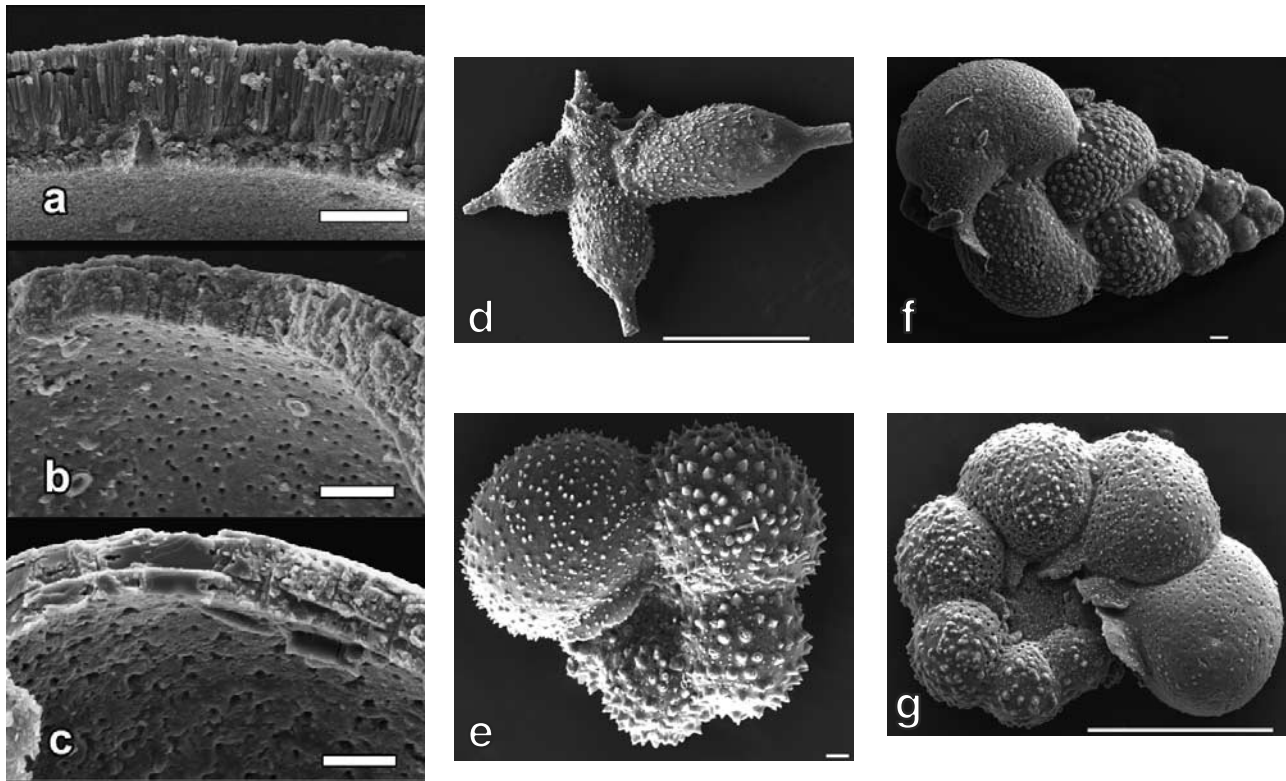


Figure 2. Scanning electron micrographs of foraminifera from Deep Sea Drilling Project Site 511. (a)–(c) Wall microstructure, note primary microlayering and absence of secondary calcite on inner and outer shell wall. Scale bars = 5 μm . (a) Planktonic *Whiteinella baltica*, 511-47-3, 23–25 cm. (b) Benthic *Gyroidinoides globosus*, 511-47-5-13.5–15.5 cm. (c) Benthic *Gyroidinoides globosus*, 511-48-1-24–27 cm. (d)–(g) Whole specimens from 511-47-5, 23–25 cm. (d) *Schackoina cenomana*, scale = 100 μm . (e) *Whiteinella baltica*, scale = 10 μm . (f) *Laeviheterohelix*, scale = 10 μm . (g) *Globigerinelloides* sp., scale = 100 μm .

(>85%) of specimens only between 47-6, 72 cm and 47-3, 23 cm. Above and below this interval in core 47, specimens display excellent preservation, but have mostly opaque skeletons. In section 48-1, foraminifera are texturally well preserved but are typically opaque and in many cases fragmentary. Preservation of glassy tests, which is very rare for foraminifera of this age, can be attributed to the moderate burial depth and low porosity and permeability of the enclosing clay-rich sediments. Primary foraminifer wall microstructure preservation is excellent throughout (Figure 2), with distinct shell layering and wall pores clearly visible even in the material that could not be described as “glassy” in dry specimens. There is no visible evidence of recrystallization or addition of secondary calcite inside or outside the shell walls in any foraminifera samples. Calcareous nannofossils from all samples in the late Turonian interval were rated as moderately preserved by Wise [1983] because of etching as a result of minor dissolution.

[8] The foraminiferal abundance and species diversity follows similar trends to skeletal transparency, with the greatest species richness (up to 16 species) of the entire Cretaceous sequence, the greatest individual abundance, and the highest percentage of keeled planktonic species in sections 47-4 to the top of 47-6. Such dominance of the

foraminifer assemblages by planktonic specimens suggests the sediments were deposited below 1000 m water depth on the basis of analogy with modern planktonic:benthic foraminiferal ratios [Gibson, 1989]. The increased species diversity and presence of the keeled species in the late Turonian are consistent with warming and/or increased vertical stratification of a normal marine, oligotrophic surface mixed layer. Calcareous nannofossils from this interval similarly indicate their growth in relatively warm, normal marine and oligotrophic surface waters during the late Turonian (S. Wise, personal communication, 2002).

[9] Planktic foraminifera are almost entirely absent above sample 47-1, 106 cm and are present almost exclusively as atypically small specimens in the lower part of section 47-1 and throughout 47-2. In section 48-1, planktic foraminifera vary between 50 and 95% of the total assemblage and are lower in diversity compared with those in much of core 47.

3.2. Stable Isotopes

[10] The new isotope data are consistent with values reported by Huber *et al.* [1995]. Both sets of data are shown in Figure 3 and our new data are listed in Table 1. There is an abrupt decrease upsection in planktonic $\delta^{18}\text{O}$ between 412.22 and 412.14 mbsf. Below this level,

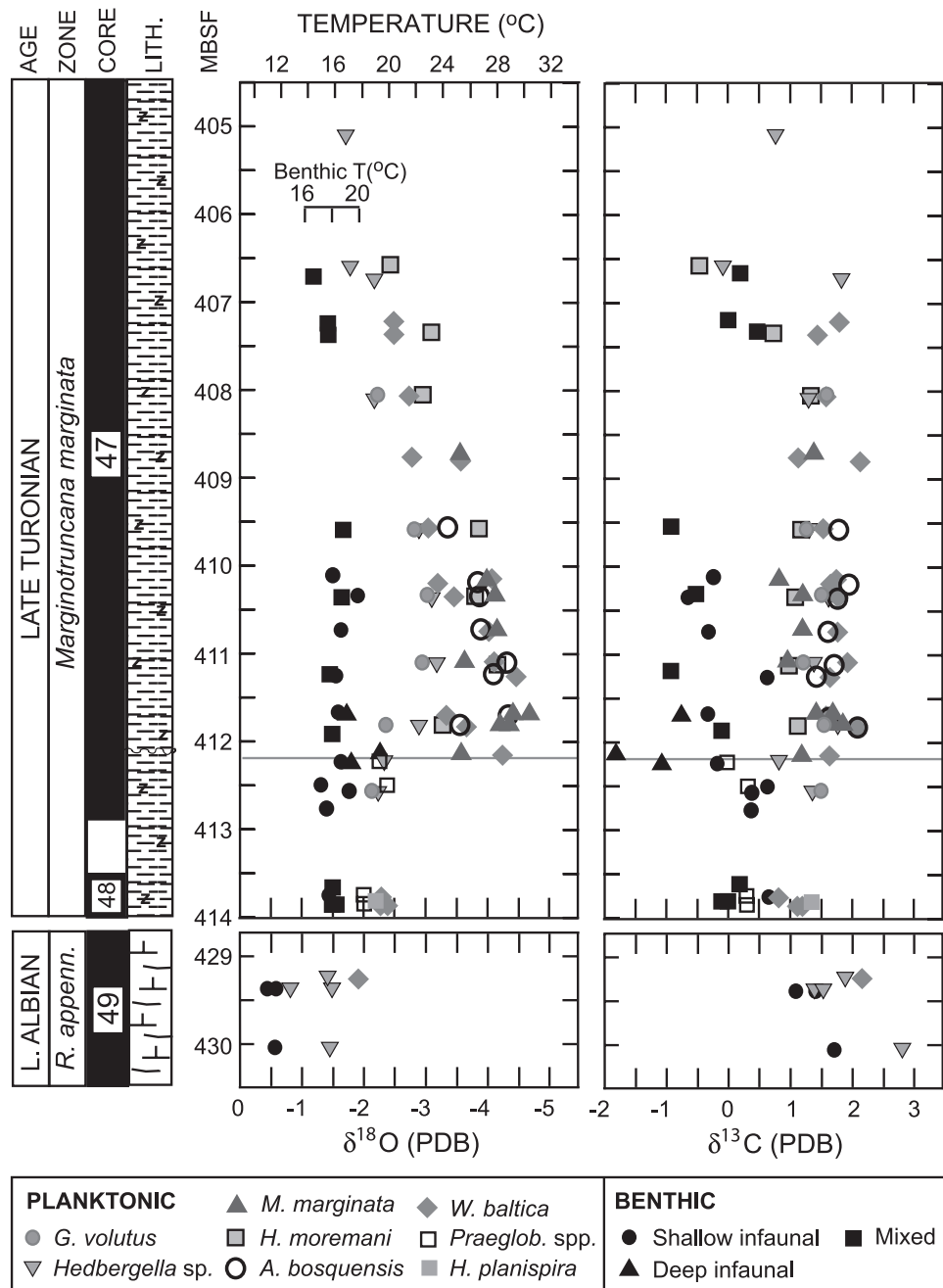


Figure 3. Oxygen and carbon isotope data for DSDP Site 511, cores 47-1 through 49-5. Planktonic foraminifer temperature calculations assume $\delta_w = -1.5\text{‰}$ SMOW. Benthic foraminifer temperature calculations assume δ_w at 1000 m water depth = -1.0‰ . The horizontal line indicates the position of a subtle lithologic change observed in the archive half of core 511-47-6 at 412.18 mbsf (see text). Other such lithologic changes occur in core 47 but their positions are not noted here. The lithologies are clay stone with rare zeolites (Turonian) and calcareous ooze (Albian). Black fill represents core recovered. The core number is shown in the core column. See color version of this figure at back of this issue.

Turonian $\delta^{18}\text{O}$ averages $\sim -2.2\text{‰}$ (PDB) with low ($<0.4\text{‰}$) variance and late Albian $\delta^{18}\text{O}$ values average -1.4‰ (PDB). Above this level, planktonic oxygen values decrease abruptly by $\sim 2\text{‰}$ and show good separation. *Marginotruncana marginata*, *H. moremani* and *A. bosquensis* generally exhibit the lowest values and *Hedbergella* and *G.*

volutus have the highest values. Two samples of *W. baltica* exhibit values below -4.00‰ , including the second lowest $\delta^{18}\text{O}$ value (-4.47‰) we measured, but otherwise *W. baltica* varies within the spread of planktonic $\delta^{18}\text{O}$ values. *M. marginata* yielded the lowest $\delta^{18}\text{O}$ measurement (-4.66‰). With the exception of an anomalous (in $\delta^{18}\text{O}$

Table 1. Oxygen and Carbon Isotopic Values for Planktonic and Benthic Foraminifera From DSDP Site 511

Core	Interval, cm	Depth, mbsf	Taxon	$\delta^{13}\text{C}$, ‰	$\delta^{18}\text{O}$, ‰
47-1	106–108	405.06	<i>Hedbergella</i> sp.	0.746	–1.721
47-2	106–108	406.56	<i>Hedbergella</i> sp.	–0.048	–1.768
47-2	106–108	406.56	<i>Heterohelix moremani</i>	–0.483	–2.467
47-2	120–122	406.70	<i>Hedbergella</i> sp.	1.826	–2.169
47-2	120–122	406.70	mixed benthics	0.176	–1.222
47-3	34–36	407.34	<i>Heterohelix moremani</i>	0.758	–3.106
47-3	34–36	407.34	<i>Whiteinella baltica</i>	1.453	–2.485
47-3	34–36	407.34	mixed benthics	0.457	–1.459
47-3	105–107	408.05	<i>Globigerinelloides volutus</i>	1.592	–2.224
47-3	105–107	408.05	<i>Hedbergella</i> sp.	1.294	–2.205
47-3	105–107	408.05	<i>Whiteinella baltica</i>	1.586	–2.763
47-3	105–107	408.05	<i>Heterohelix moremani</i>	1.312	–2.945
47-4	30–32	408.80	<i>Whiteinella baltica</i>	2.104	–3.578
47-4	106–108	409.56	<i>Globigerinelloides volutus</i>	1.276	–2.816
47-4	106–108	409.56	<i>Hedbergella</i> sp.	1.296	–2.906
47-4	106–108	409.56	<i>Whiteinella baltica</i>	1.528	–3.073
47-4	106–108	409.56	<i>Heterohelix moremani</i>	1.161	–3.862
47-4	106–108	409.56	<i>Archaeoglobigerina bosquensis</i>	1.757	–3.374
47-4	106–108	409.56	mixed benthics	–0.904	–1.682
47-5	33–36	410.33	<i>Globigerinelloides volutus</i>	1.499	–3.031
47-5	33–36	410.33	<i>Hedbergella</i> sp.	1.632	–3.120
47-5	33–36	410.33	<i>Whiteinella baltica</i>	1.762	–3.466
47-5	33–36	410.33	<i>Heterohelix moremani</i>	1.062	–3.802
47-5	33–36	410.33	<i>Archaeoglobigerina bosquensis</i>	1.737	–3.921
47-5	33–36	410.33	<i>Marginotruncana marginata</i>	1.219	–4.115
47-5	33–36	410.33	<i>Gyroidinoides globosus</i>	–0.655	–1.932
47-5	33–36	410.33	mixed benthics	–0.533	–1.675
47-5	13.5–15.5	410.14	<i>Whiteinella baltica</i>	1.706	–4.078
47-5	13.5–15.5	410.14	<i>Marginotruncana marginata</i>	0.849	–3.980
47-5	13.5–15.5	410.14	<i>Gyroidinoides globosus</i>	–0.256	–1.477
47-5	72–74	410.72	<i>Whiteinella baltica</i>	1.779	–4.014
47-5	72–74	410.72	<i>Archaeoglobigerina bosquensis</i>	1.627	–3.876
47-5	72–74	410.72	<i>Marginotruncana marginata</i>	1.221	–4.154
47-5	72–74	410.72	<i>Gyroidinoides globosus</i>	–0.338	–1.653
47-5	109–111	411.09	<i>Globigerinelloides volutus</i>	1.218	–2.938
47-5	109–111	411.09	<i>Hedbergella</i> sp.	1.369	–3.164
47-5	109–111	411.09	<i>Whiteinella baltica</i>	1.930	–4.085
47-5	109–111	411.09	<i>Heterohelix moremani</i>	0.995	–4.164
47-5	109–111	411.09	<i>Archaeoglobigerina bosquensis</i>	1.707	–4.334
47-5	109–111	411.09	<i>Marginotruncana marginata</i>	0.941	–3.610
47-5	124–126	411.24	<i>Whiteinella baltica</i>	1.635	–4.473
47-5	124–126	411.24	<i>Archaeoglobigerina bosquensis</i>	1.429	–4.089
47-5	124–126	411.24	<i>Berthelina</i> sp.	0.598	–1.585
47-5	124–126	411.24	mixed benthics	–0.920	–1.476
47-6	18–20	411.68	<i>Marginotruncana marginata</i>	1.673	–4.664
47-6	18–20	411.68	<i>Dentalina</i> sp.	–0.789	–1.753
47-6	18–20	411.68	shallow infaunal benthics	–0.318	–1.612
47-6	31–33	411.81	<i>Globigerinelloides volutus</i>	1.522	–2.369
47-6	31–33	411.81	<i>Hedbergella</i> sp.	1.764	–2.908
47-6	31–33	411.81	<i>Whiteinella baltica</i>	2.061	–3.677
47-6	31–33	411.81	<i>Heterohelix moremani</i>	1.107	–3.277
47-6	31–33	411.81	<i>Archaeoglobigerina bosquensis</i>	2.081	–3.589
47-6	31–33	411.81	<i>Marginotruncana marginata</i>	1.690	–4.240
47-6	31–33	411.81	<i>Marginotruncana marginata</i>	1.849	–4.191
47-6	38–40	411.88	mixed benthics	–0.111	–1.542
47-6	64.5–66.5	412.14	<i>Whiteinella baltica</i>	1.647	–4.282
47-6	64.5–66.5	412.14	<i>Marginotruncana marginata</i>	1.172	–3.584
47-6	64.5–66.5	412.14	<i>Dentalina</i> sp.	–1.807	–2.246
47-6	72–74	412.22	<i>Hedbergella</i> sp.	0.835	–2.348
47-6	72–74	412.22	<i>Praeglobotruncana delrioensis</i>	–0.006	–2.236
47-6	72–74	412.22	trochospiral benthics	–0.206	–1.664
47-6	72–74	412.22	deeper infaunal benthics	–1.129	–1.830
47-6	98–100	412.48	<i>Praeglobotruncana stephani</i>	0.275	–2.390
47-6	98–100	412.48	<i>Berthelina</i> sp.	0.613	–1.334
47-6	107.5–109.5	412.58	shallow infaunal benthics	0.400	–1.742
47-6	105–107	412.55	<i>Globigerinelloides volutus</i>	1.501	–2.146
47-6	105–107	412.55	<i>Hedbergella</i> sp.	1.358	–2.244
47-6	126–128	412.76	<i>Berthelina</i> sp.	0.333	–1.381
48-1	24–27	413.74	<i>Berthelina</i> sp.	0.631	–1.475
48-1	33–35	413.83	<i>Whiteinella baltica</i>	1.154	–2.327
48-1	33–35	413.83	<i>Whiteinella baltica</i>	1.238	–2.371

Table 1. (continued)

Core	Interval, cm	Depth, mbsf	Taxon	$\delta^{13}\text{C}$, ‰	$\delta^{18}\text{O}$, ‰
48-1	33–35	413.83	<i>Hedbergella</i> sp. <i>planispira</i>	1.324	−2.194
48-1	33–35	413.83	<i>Praeglobotruncana stephani</i>	0.339	−2.075
48-1	33–35	413.83	<i>Praeglobotruncana stephani</i>	0.295	−2.028
48-1	33–35	413.83	mixed benthics	−0.087	−1.582
48-1	33–35	413.83	deeper infaunal benthics	0.013	−1.542
49-5	65.5–67.5	429.36	<i>Hedbergella</i> sp. <i>planispira</i>	1.511	−0.811
49-5	65.5–67.5	429.36	<i>Hedbergella</i> sp. <i>portsdownensis</i>	1.370	−1.476
49-5	65.5–67.5	429.36	<i>Berthelina</i> sp.	1.086	−0.558
49-5	65.5–67.5	429.36	<i>Berthelina</i> sp.	1.376	−0.448
49-5	103–105	430.04	<i>Hedbergella yezoana</i>	2.784	−1.457
49-5	103–105	430.04	<i>Berthelina</i> sp.	1.704	−0.555

and $\delta^{13}\text{C}$) *Dentalina* measurement at 412.14 mbsf, benthic values show little change through the study interval.

[11] Cross plots of carbon and oxygen isotope data provide an indication of planktic foraminiferal depth ecology. Within core 47, *G. volutus*, *H. moremani*, and *Hedbergella* spp. consistently plot as lower mixed layer species. *A. bosquensis* and *W. baltica* appear to be upper mixed layer species, and *M. marginata* occupies a mid to upper mixed layer position. The separation of these species from co-occurring benthics is consistently large through the interval.

[12] The abrupt change in $\delta^{18}\text{O}$ that occurs between 412.22–412.14 mbsf suggests the possibility of an unconformity. There is a subtle lithologic change at 412.18 mbsf where sediment color changes from a medium dark gray below to lighter gray above. Carbonate content, foraminifer abundance and planktonic species diversity increase above this point. XRD analysis of six samples taken within 2 m above and below the $\delta^{18}\text{O}$ shift show no significant change in mineralogy of the bulk sediments. Sediment color alternates between darker and lighter gray throughout most of core 47-6, but the change is often more gradual than that at 412.18 mbsf. If the subtle lithologic change marks a hiatus, the amount of missing time is entirely unconstrained.

3.3. Temperature Estimates

[13] To calculate upper ocean temperatures from planktonic foraminiferal $\delta^{18}\text{O}$, we use the equation of Erez and Luz [1983]. Ambient water $\delta^{18}\text{O}$ value (δ_w) is estimated (relative to Standard Mean Ocean Water, SMOW) and is converted to the PDB scale by subtracting 0.27‰ [Hut, 1987]. We first use the standard approach to estimating δ_w , which assumes an “ice-free” mean ocean δ_w of −1.0‰ SMOW [Shackleton and Kennett, 1975] and includes an adjustment that takes into account average latitudinal variations in evaporation and precipitation controls on $\delta^{18}\text{O}$ by analogy with the modern ocean [e.g., Zachos et al., 1994]. At a latitude of 60°, this approach yields δ_w = −1.2‰ (SMOW) and paleotemperatures of 19–21°C below 412.18 mbsf. Above this depth, calculated temperatures are as warm as 30–32°C for shallower dwelling *M. marginata*, *W. baltica*, and *A. bosquensis*, and 23–25°C for deeper dwelling planktonic foraminifer species. More conservatively, we can assume that surface δ_w at 511 was as depleted relative to the global mean as the lowest water $\delta^{18}\text{O}$ in the *Geochemical Ocean Sections Study* (GEOSECS) [1987] data set poleward of 50°S latitude. This value (−0.48‰

SMOW) occurs at 1 m water depth at station 82 in the austral summer, near the South Sandwich Islands, at 56.3°S latitude. Assuming a local δ_w of −1.5‰ (\sim −0.48 + −1.0‰) decreases the temperature estimates given above by 1.4°C. The more conservative δ_w estimate results in maximum upper ocean temperatures of 28–30°C for shallower dwelling species and 22–24°C for deeper dwelling planktonic foraminifera (Figure 3).

[14] Assuming a range of possible bottom δ_w values between average “ice-free” seawater (−1.0‰ SMOW) and the conservative upper ocean water estimate (−1.5‰ SMOW), we estimate local bottom water temperatures (>1000 m) to have been 16–20°C throughout the study interval. The data indicate a local vertical temperature gradient of only \sim 4°C below 412.18 mbsf and as great as 12°C above this point.

4. Comparison With Other Late Turonian Data

[15] We compared our data from DSDP Site 511 with age equivalent records from other sites where synchronicity could be determined through a combination of biostratigraphy and strontium (Sr) isotope ratios. Unfortunately, few age equivalent records exist. Sr isotope data are available for foraminifera reported by Huber et al. [1995] for Site 511 and equatorial Atlantic Site 144 [Wilson et al., 2002], as well as new data for Site 511 (Table 2). After renormalization to NIST 987 = 0.710248 (as required to compare the data to the Sr reference curve of McArthur et al. [2001]), these records are age equivalent within an interval of <600 kyr in the late Turonian.

Table 2. $^{87}\text{Sr}/^{86}\text{Sr}$ Results for Late Albian and Late Turonian Samples for DSDP Site 511^a

Sample	Depth, mbsf	$^{87}\text{Sr}/^{86}\text{Sr}^b$	Standard Error, %	Type
47-5, 13.5–15.5 cm	410.14	0.707314	0.0006	Foraminifera
47-6, 18–20 cm	411.68	0.707323	0.0007	Foraminifera
48-1, 14–18 cm	413.21	0.707408	0.0006	Foraminifera
48-1, 33–35 cm	413.83	0.707377	0.0007	Foraminifera
48-2, 84–86 cm	415.84	0.707379	0.0006	Inoceramid prisms

^aSee Huber et al. [1995] and Bralower et al. [1997] for additional Cretaceous Sr values for this site.

^bAll samples analyzed by Paul D. Fullagar at the University of North Carolina. $^{87}\text{Sr}/^{86}\text{Sr}$ values are corrected using the difference between the laboratory average of 0.710270 for the NIST 987 carbonate standard and the accepted value of 0.710250 for NIST 987.

[16] *Ditchfield et al.* [1994] report oyster $\delta^{18}\text{O}$ values of -1.75 , -4.64 and -6.09 ‰ PDB from the mid Cretaceous on James Ross Island (JRI, paleolatitude $\sim 65^\circ\text{S}$, Figure 1). In paleotemperature calculations, *Ditchfield et al.* exclude the two lower values, apparently on the basis of textural preservation or the isotopic values themselves [*Ditchfield et al.*, 1994, p. 86]. However, based on iron and manganese concentrations reported by *McArthur et al.* [2000] for these samples and the criteria for unaltered material defined by *Ditchfield et al.* (<500 ppm Fe, <200 ppm Mn), only one of the oyster $\delta^{18}\text{O}$ values reported from JRI should have been used in paleotemperature calculations: that with an oxygen isotope value of -6.09 ‰ PDB. On the basis of SEM examination, *McArthur et al.* [2000] describe this sample (D.8228.113) as flaky, translucent calcite characteristic of pristine oyster carbonate.

[17] *McArthur et al.* [2000] showed that, based on $^{87}\text{Sr}/^{86}\text{Sr}$, JRI sample D.8228.113 dates to the late Turonian strontium minimum at ~ 89.45 – 89.74 Ma. Normalizing all the Sr data from Sites 144, Site 511, and James Ross Island to NIST 987 = 0.710248 [*McArthur et al.*, 2001] shows that the three data sets date to within a 1.5 myr interval of the late Turonian. We note, however, that high-resolution mid-Cretaceous records [e.g., *Wilson and Norris*, 2001] show substantial climate variability within such a time window. We are therefore severely limited in what we can say about changes in late Turonian global climate given the data available for comparison.

[18] *Wilson et al.* [2002] have inferred late Turonian sea surface temperatures of 30 – 33°C from their data suggesting that the tropical Atlantic had temperatures similar to or slightly warmer than maximum temperature estimates from Site 511 (Figure 4). Paleotemperatures implied by the oyster data on James Ross Island are hard to determine precisely since oysters can live in much lower-salinity waters (influenced by local runoff) than is generally believed to be true for planktonic foraminifera. Here we make the liberal ad hoc assumption that the oysters grew in waters 2.0 ‰ more depleted than open ocean waters at equivalent latitude. This is a generous allowance for depleted water given that the JRI faunal assemblage is interpreted as indicative of normal marine salinities [*Ditchfield et al.*, 1994]. If the mean ocean δ_w was -1.0 ‰ SMOW, our assumptions yield a δ_w of -3.5 ‰ for the oyster sample. The paleotemperature equation for mollusc carbonate [*Anderson and Arthur*, 1983] then indicates that shelf waters around James Ross Island were about 27°C in the late Turonian. (A smaller δ_w adjustment for runoff would result in a higher temperature estimate.) Hence the oyster data are in broad agreement with the temperature estimates derived above for DSDP Site 511 and reinforce an interpretation of a small latitudinal temperature gradient during the core of the Cretaceous greenhouse climate (Figure 4).

[19] Somewhat lower-temperature estimates result for all these sites if we take into account the long term trend in mean δ_w due to the imbalance between ^{18}O sources, primarily high-temperature seafloor alteration, and sinks such as continental weathering, low-temperature seafloor alteration, and water fixation in the upper oceanic crust and in clay minerals. *Wallmann* [2001] estimated that the mean

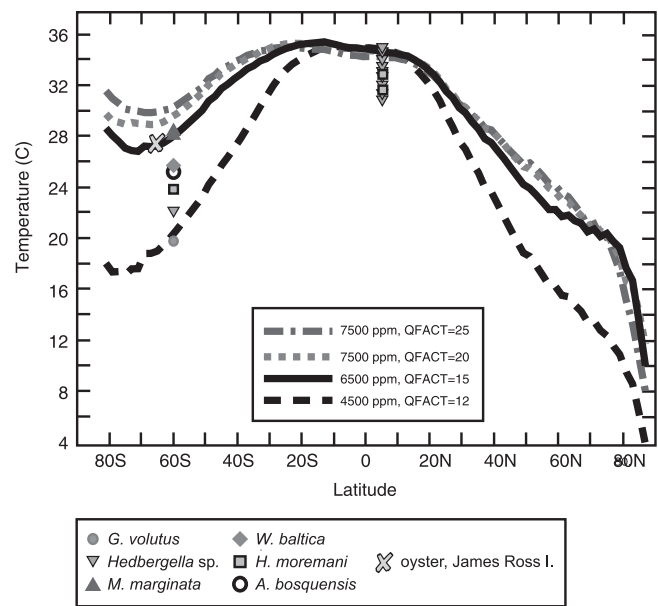


Figure 4. Curves show mean January (austral summer) sea surface temperature profiles from four GENESIS version 2.0 AGCM experiments. Symbols show estimated upper ocean temperatures from $\delta^{18}\text{O}$ of approximately coeval late Turonian samples. Isotopic temperature calculations assume the following δ_w values: Site 144 foraminifera (5°N lat.) $\delta_w = -0.25$ ‰ SMOW; Site 511 foraminifera (60°S lat.) $\delta_w = -1.5$ ‰; James Ross Island mollusc (65°S lat.) $\delta_w = -3.5$ ‰ (see text). Symbol shapes for foraminiferal species are the same as those used in Figure 3. All Site 144 estimates are from *H. moremani* (squares) and *Hedbergella* sp. (inverted triangles). Site 144 data are from *Wilson et al.* [2002]. The James Ross Island datum is from *Ditchfield et al.* [1994]. See color version of this figure at back of this issue.

ocean δ_w at 100 Ma would have been depleted ~ 0.25 ‰ beyond the ice-free value, yielding an ocean mean of -1.25 ‰ SMOW. This -0.25 ‰ adjustment would decrease all isotopic paleotemperature estimates shown in Figure 4 by $\sim 1.2^\circ\text{C}$.

[20] If the Site 511 late Turonian data record an interval of extreme polar warmth, the most likely mechanism is a substantial increase in atmospheric greenhouse gas concentrations and increased atmospheric poleward heat transport. Using annual mean conditions predicted from the GENESIS version 2.0 atmospheric general circulation model (AGCM), *Bice and Norris* [2002] determined that, if the anomalously high temperatures implied by the Site 511 late Turonian data are ignored, maximum late Albian-Turonian temperatures can be reproduced with 4500 ppm CO_2 and the consequent water vapor feedbacks. When the maximum temperatures estimated from the Site 511 data need to be matched, 6500 – 7500 ppm CO_2 is required in the AGCM. Poleward heat transport must be increased by at least 50% from the “background” warm mid-Cretaceous conditions. Mean zonal January (austral summer) temperature profiles from these models are shown in Figure 4.

[21] Any rise in high-latitude surface temperature occurring during a Turonian greenhouse climate should also affect temperatures at the sites of bottom water formation and increase temperatures estimated from benthic foraminifera. However, there is no significant change in bottom water temperature estimates at Site 511 associated with the interval of low planktonic $\delta^{18}\text{O}$ values (Figure 3). Indeed, when the GFDL Modular Ocean Model version 2 is forced with the mean annual atmospheric output from AGCM runs with 6500–7500 ppm CO_2 , the bottom water temperature predicted is $\sim 28^\circ\text{C}$, which is $\sim 10^\circ\text{C}$ warmer than bottom temperatures estimated from the benthic data. High-latitude winter SSTs predicted by the AGCM with 6500–7500 ppm are $23\text{--}25^\circ\text{C}$. Therefore, if the ocean model is forced by seasonally varying surface temperatures and bottom water formed only during winter months, bottom waters would still overpredict the isotopic paleotemperatures by $6\text{--}8^\circ\text{C}$.

[22] We can produce a reasonable match to Site 511 benthic paleotemperatures in model runs using 4500 ppm CO_2 . Although experiments with 4500 ppm CO_2 underpredict the warmest estimated surface ocean temperatures (Figure 4), the mismatch is not much greater than the error in the isotopic paleotemperature calculation. Assuming that the general circulation models are reliable tools for studying greenhouse climates, the models suggest that, if the Site 511 late Turonian data are a record of surface warming alone, then carbon dioxide concentrations would have likely been 4500 ppm or greater. Carbon dioxide concentrations of this magnitude are inferred from leaf stomatal index measurements of Cenomanian and Turonian *Ginkgo* fossils [Retallack, 2001] but are higher than values derived from other mid-Cretaceous proxies [Bice and Norris, 2002].

5. Alternative Explanations for the Site 511 $\delta^{18}\text{O}$ Data

5.1. Possible Diagenesis

[23] It appears highly unlikely that the low $\delta^{18}\text{O}$ planktonic foraminifer values result from diagenetic alteration. The significant separation among the planktonics and between planktonics and benthics argues against the low $\delta^{18}\text{O}$ being caused by postdepositional calcification or alteration. There is no textural evidence for infilling, encrustation or recrystallization. The high quality of planktonic foraminiferal preservation is also reflected by the abundant occurrence above 412.18 mbsf of delicate biserial foraminifera, which are rare lower in the section.

[24] Authigenic clay formation, which can promote low pore water $\delta^{18}\text{O}$ [Lawrence et al., 1975] appears to be an unlikely mechanism for alteration here. Robert and Maillot [1983] report no evidence for authigenic clays and the absence of evidence of burial, volcanic or organic diagenesis. The predominant clay mineral in the Turonian sequence is pedogenic smectite, interpreted as having been derived from deep soils in downstream, low-relief areas of a continental drainage basin in an area of constantly warm, wet climate [Robert and Maillot, 1983; Timofeev et al., 1983].

[25] Huelandite group zeolites are common in the Santonian through Maastrichtian section at Site 511, but are rare

below this level, including the interval we studied [Varentsov et al., 1983; Huber et al., 1995]. Varentsov et al. [1983] report that the highest content of fine basaltic volcanoclastic sediments that have been altered to huelandite occurs in the Campanian-Maastrichtian interval, yet foraminiferal $\delta^{18}\text{O}$ values for this interval are quite enriched [Huber et al., 1995]. This argues against diagenesis of volcanoclastic material as a source of light oxygen in the late Turonian interval.

[26] DSDP Site 511 sits on continental crust, so pore water interaction with oceanic crust basalts is an unlikely mechanism for the introduction of depleted oxygen. Site 511 sediment temperature measurements made at 0, 52 and 113 mbsf indicate a local temperature gradient of 74°C km^{-1} [Langseth and Ludwig, 1983] and indicate a sediment temperature at 412 mbsf of only $\sim 32^\circ\text{C}$. Cretaceous sediments crop out in various places on Falkland Plateau as a result of Maastrichtian and Eocene stripping [Ludwig, 1983]. It is therefore possible that the Turonian sequence was more deeply buried in the past. However, the relatively good preservation of foraminifera in the chalkier parts of the Cretaceous sequence suggests that they were never very deeply buried, and Robert and Maillot [1983] note the absence of any textural evidence for burial diagenesis in these sediments. Significant alteration of the primary $\delta^{18}\text{O}$ signal through solid state diffusion of oxygen isotopes at temperatures below 50°C is unlikely, even if we assume that pore waters had somehow become strongly ^{18}O -depleted at some point following burial [Anderson, 1969; Kronenberg et al., 1984].

5.2. Upper Ocean δ_w Decrease Through Increased Freshwater Flux

[27] The fact that benthic $\delta^{18}\text{O}$ is unchanged through the Site 511 Turonian interval suggests that the planktonic foraminifera record a period during which local water became more ^{18}O -depleted and that bottom waters were largely unaffected by any coincident climate or oceanographic change. Assuming no accompanying temperature change, the data indicate that waters over Falkland Plateau became $\sim 2\text{‰}$ more depleted relative to the global mean. Two broad mechanisms can decrease local water $\delta^{18}\text{O}$: an increased flux of ^{18}O -depleted freshwater and a decrease in the $\delta^{18}\text{O}$ of local runoff and precipitation. These two changes need not be mutually exclusive, but they can properly be considered separately.

[28] If we view the modern Southern Ocean surface water average δ_w ($\sim -0.3\text{‰}$) as resulting from a simple mixing of average seawater (0‰) and depleted precipitation and runoff ($\sim -10\text{‰}$) [Rozanski et al., 1993], then the fraction of upper ocean water contributed by precipitation and runoff today is $\sim 3\%$. In order to decrease local δ_w by 2‰ without invoking a change in the oxygen isotopic composition of precipitation/runoff, the freshwater contribution would have to increase to 23% (or to 20–22% if the Cretaceous mean ocean δ_w was -1.0 to -1.25‰).

[29] An increase in high-latitude precipitation rate is an expected consequence of tropical warming [Manabe, 1996] and a stronger hydrologic cycle. Hence some increase in high-latitude precipitation rate might be expected in

response to the late Turonian tropical warmth [Wilson *et al.*, 2002]. However, an increase to a 23‰ freshwater contribution implies a 6–7 psu decrease in local salinity, again, using a simple mixing ratio. The observed maximum in planktonic abundance and diversity in Site 511 in the interval with the lowest $\delta^{18}\text{O}$ argues against such a salinity decrease. Leckie [1987] showed that keeled foraminifera (e.g., *M. marginata*) are most abundant in open ocean settings, suggesting a preference for relatively normal salinities. While we cannot rule out some decrease in salinity, the increase in faunal diversity, high planktonic:benthic ratios and abundance of keeled forms where planktonic foraminiferal $\delta^{18}\text{O}$ becomes very low are more consistent with a change from abnormal salinity conditions below 412.18 mbsf to more normal marine salinity above. At 412.18 mbsf, we observe a subtle increase in carbonate content and consequent decrease in clay percentage upsection, into the interval of depleted $\delta^{18}\text{O}$. A substantial increase in continental runoff to the basin might be reflected in an increase in detrital clay flux, but the opposite change occurs. Given the lack of any change consistent with decreased salinity or increased runoff, we do not believe the Site 511 $\delta^{18}\text{O}$ record can be explained by a substantially increased freshwater flux.

5.3. Upper Ocean δ_w Decrease Through Decreased Freshwater $\delta^{18}\text{O}$

[30] Is it possible that surface waters could have become depleted by 2‰ relative to the global mean because of a decrease in the $\delta^{18}\text{O}$ of freshwater? Using again the estimated modern Southern Ocean freshwater mixing ratio of 3%, in order to decrease local δ_w by 2‰ using only the composition of freshwater, local precipitation/runoff $\delta^{18}\text{O}$ would have had to decrease by 58‰ or more.

[31] Modern precipitation reaches a minimum $\delta^{18}\text{O}$ of -50‰ SMOW over the South Pole [Dansgaard, 1964]. Experiments with atmospheric general circulation models that include water isotopes tracers indicate that high-latitude precipitation would be more enriched in ^{18}O in a warmer climate. Using modern boundary conditions, Jouzel *et al.* [2000] show that the climatic changes accompanying an increase in CO_2 from 315 to 630 ppm CO_2 would increase high-latitude precipitation $\delta^{18}\text{O}$ by 3–4‰ in the north and by 3–7‰ in the south. Recent experiments with an isotope-capable version of GENESIS and mid-Cretaceous boundary conditions [Bice *et al.*, 2002] are in agreement with the Jouzel *et al.* model prediction of isotopically enriched high-latitude precipitation. In the GENESIS model [Mathieu *et al.*, 2002] with 4500 ppm CO_2 , the predicted $\delta^{18}\text{O}$ of precipitation over Falkland Plateau varies between -5 and -7‰ SMOW (summer-winter). If the model predictions that a warmer world has a lower gradient in precipitation $\delta^{18}\text{O}$ are correct, then we cannot invoke more depleted high-latitude freshwater to explain the Site 511 record.

[32] Where do waters 2‰ lower than the global mean occur today? Ocean water $\delta^{18}\text{O}$ -2.0‰ or lower is recorded in the North Atlantic near 65°N , where the East Greenland Current supplies ^{18}O -depleted Arctic water [Azetsu-Scott and Tan, 1997] to water depths less than 50 m. Water $\delta^{18}\text{O}$

values lower than -2‰ have also been measured in fiords, within ~ 100 km of the melting terminus of a tidewater glacier [Azetsu-Scott and Tan, 1997]. In one study near the Antarctic West Ice Shelf at $63\text{--}65^\circ\text{S}$, Frew *et al.* [1995] found no water $\delta^{18}\text{O}$ value lower than -0.5‰ , consistent with all GEOSECS [1987] Southern Ocean measurements. It therefore appears that the modern analogs to the δ_w required (if the late Turonian data are to be explained by δ_w change) are associated with Arctic ocean outflow and tidewater glacier melting. However, the appearance of local glacial runoff in the Turonian is not supported by any lithologic changes that might be expected in this scenario (e.g., increased clastic flux), and benthic foraminifer $\delta^{18}\text{O}$ do not record coincident changes in temperature or δ_w , although such changes might be expected given a glacial mechanism.

6. Concluding Remarks

[33] Our new results, combined with those of Huber *et al.* [1995], clearly show that subpolar planktonic foraminifera over Falkland Plateau became strongly ^{18}O depleted in the late Turonian, about the time of the warmest tropical temperatures thus far estimated for the open ocean [Wilson *et al.*, 2002]. We do not yet have a satisfactory explanation for the extremely depleted planktonic foraminifer $\delta^{18}\text{O}$ values above 412.18 mbsf at Site 511. Diagenesis appears unlikely as a mechanism given the good separation among planktonic $\delta^{18}\text{O}$ values, lack of evidence for ^{18}O -depleted pore fluids, low burial temperature, and absence of any textural evidence for alteration or secondary calcification.

[34] We do not rule out warming: The nannofossil assemblage changes at 412.18 mbsf have been interpreted as consistent with some warming (S. Wise, personal communication, 2002), and the upper ocean temperatures derived here are consistent with those from $\delta^{18}\text{O}$ of well-preserved late Turonian bivalves [Ditchfield *et al.*, 1994], if our assumption of -3.5‰ δ_w at James Ross Island is adequate. However, it is difficult to accept tropic-like temperatures at such high latitudes, especially when available benthic foraminifer data do not indicate coincident bottom water warming. Both the planktonic foraminifera we analyzed at Site 511 and the oysters analyzed by Ditchfield *et al.* [1994] from James Ross Island could have grown during the austral summer. Climate model experiments require very high atmospheric CO_2 concentrations to reproduce the summer temperatures estimated from the two high-latitude data sets (Figure 4). These CO_2 amounts are consistent with some estimates of mid-Cretaceous atmospheric CO_2 concentrations from fossil leaf stomatal index measurements of *Ginkgo* specimens but are significantly higher than estimates reconstructed from marine organic carbon [Freeman and Hayes, 1992; Retallack, 2001; Bice and Norris, 2002]. However, in order to explain both the planktonic and benthic $\delta^{18}\text{O}$ records from Site 511 as wholly temperature records we would have to imagine that summer upper ocean temperatures increased locally while temperatures at the sites of deep water formation during winter (the more likely time of deep water formation) did not. No suitable records exist to make this analysis.

[35] We also cannot rule out isotopically depleted upper ocean water. However, if the Site 511 record reflects an interval of decreased upper ocean water $\delta^{18}\text{O}$, we need a satisfactory mechanism(s) that would lead to depletion of local waters by $\sim 2\text{‰}$ relative to the global mean. Today, such depletion is associated with polar ocean outflow and tidewater glacier melting, but there is no evidence to support such a mechanism for the late Turonian greenhouse. Isotope-capable general circulation models support high-latitude precipitation and runoff that is more enriched in ^{18}O during a warmer climate, not more depleted. Tropical warming can be expected to lead to a stronger hydrologic cycle [Manabe, 1996] with higher precipitation rates at high latitudes and consequent increased runoff. However, whether through direct rainfall or runoff, the increase in freshwater flux to Site 511 required to effect a 2‰ decrease in local water would likely have driven a salinity decrease of 6–7 psu. In contrast, the Site 511 record and conventional interpretations of abundance and diversity changes in foraminifera and nannofossils would argue for a change to more normal marine salinities above 412.18 mbsf, not a freshening.

[36] The Cretaceous South Atlantic was a smaller, more restricted basin than the modern (Figure 1). We can speculate that the planktonic foraminifer low $\delta^{18}\text{O}$ interval might record isotopic depletion due to increased basin isolation. However, the paleogeography that included a restricted South Atlantic existed long before the late Turonian and yet no record of severe δ_w depletion has been identified in the Albian section on Falkland Plateau [Huber et al., 1995; Fassell and Bralower, 1999]. Also, if tectonic change led to severe restriction of the local basin, we might expect a coincident change in bottom water conditions, but none is indicated by the benthic foraminifer isotope records.

[37] We can speculate that perhaps Site 511 sat at the juncture of two water masses [e.g., Ciesielski et al., 1977] and that atmosphere and/or ocean circulation changes occurred to bring lower $\delta^{18}\text{O}$, normal salinity waters onto Falkland Plateau. Presumably, such changes might have been related to the tropical warmth noted in the Site 144 record [Wilson et al., 2002]. Two problems exist with a speculative high-latitude water mass mixing source, how-

ever. Such water would be expected to be cool and therefore inconsistent with the nannofossil assemblage change (section 3.1) interpreted as consistent with some warming (S. Wise, personal communication, 2002). Also, in order to explain the Site 511 shift, a lower $\delta^{18}\text{O}$ water mass from high latitudes would have to be at least as depleted as modern Arctic outflow or tidewater glacial drainage. As discussed above (section 5.3), this seems unlikely in the warm mid-Cretaceous.

[38] Alternatively, we can speculate that the record is the result of some combination of warming and increased freshwater flux. Any plausible argument must explain upper ocean change without coincident bottom water change at Site 511. We must be careful to not base too much on one data point from James Ross Island, but the most satisfactory explanation for the Site 511 record is one that would also explain extreme ^{18}O depletion (-6.09‰) [Ditchfield et al., 1994] in what has been termed pristine oyster carbonate [McArthur et al., 2000] of approximately equal age.

[39] For now, we are left with a paradoxical record in the upper Turonian on Falkland Plateau. Understanding the exact cause of the low $\delta^{18}\text{O}$ interval here will require, as a minimum, coeval upper ocean records from equivalent latitudes in either hemisphere, preferably both within and outside the South Atlantic basin. In addition, more benthic foraminifer records are needed in order to understand what global bottom water and high-latitude changes occurred approximately coeval with the extreme tropical warmth of the late Turonian.

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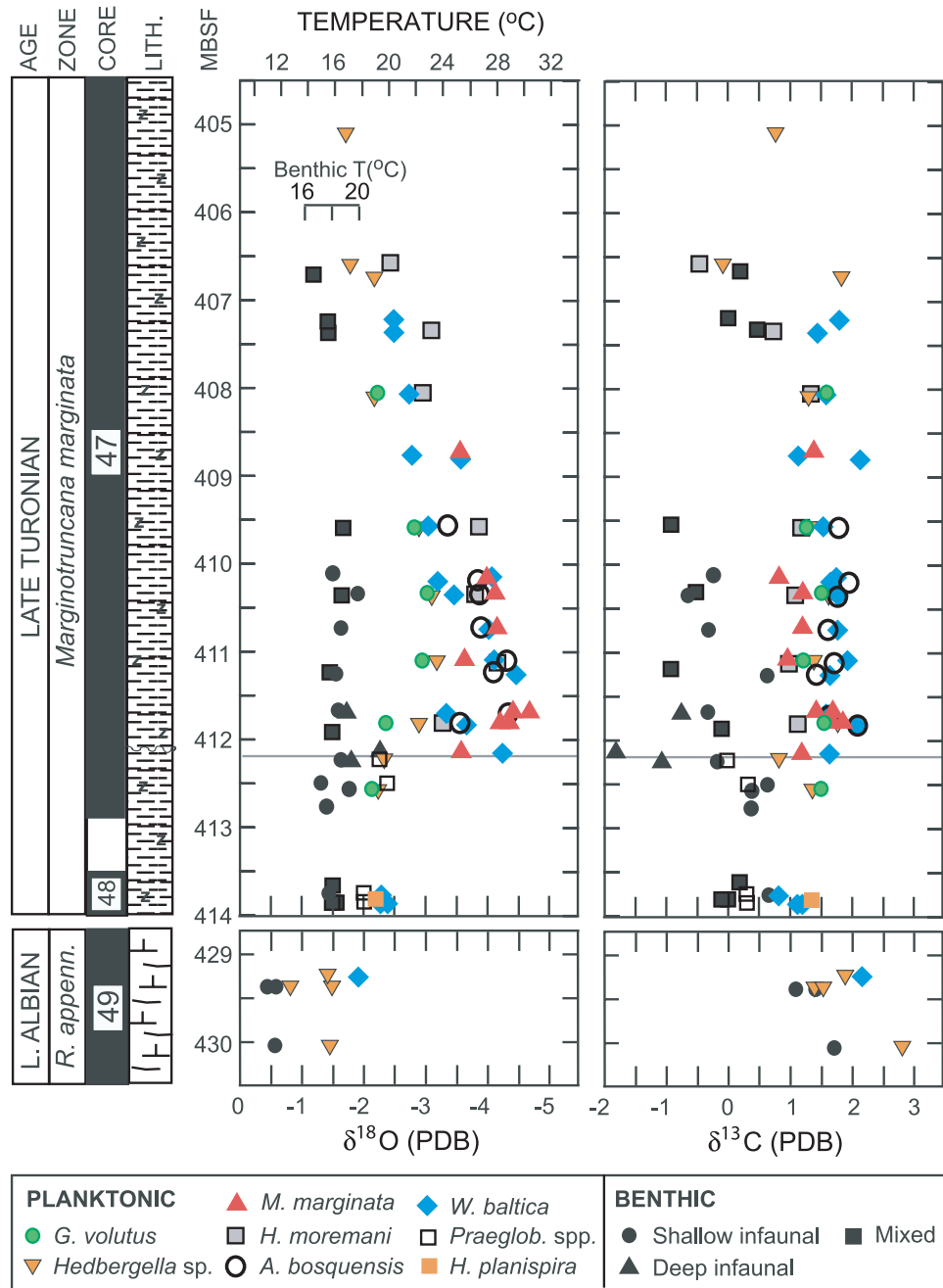


Figure 3. Oxygen and carbon isotope data for DSDP Site 511, cores 47-1 through 49-5. Planktonic foraminifer temperature calculations assume $\delta_w = -1.5\text{‰}$ SMOW. Benthic foraminifer temperature calculations assume δ_w at 1000 m water depth = -1.0‰ . The horizontal line indicates the position of a subtle lithologic change observed in the archive half of core 511-47-6 at 412.18 mbsf (see text). Other such lithologic changes occur in core 47 but their positions are not noted here. The lithologies are clay stone with rare zeolites (Turonian) and calcareous ooze (Albian). Black fill represents core recovered. The core number is shown in the core column.

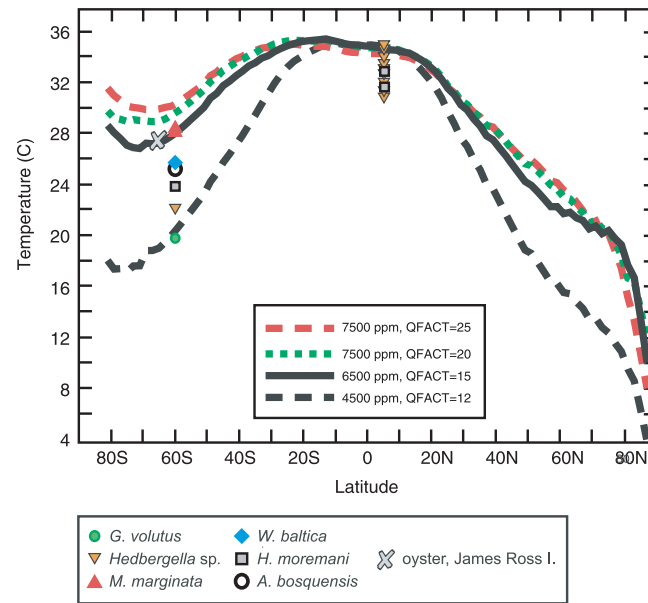


Figure 4. Curves show mean January (austral summer) sea surface temperature profiles from four GENESIS version 2.0 AGCM experiments. Symbols show estimated upper ocean temperatures from $\delta^{18}\text{O}$ of approximately coeval late Turonian samples. Isotopic temperature calculations assume the following δ_w values: Site 144 foraminifera (5°N lat.) $\delta_w = -0.25\text{‰}$ SMOW; Site 511 foraminifera (60°S lat.) $\delta_w = -1.5\text{‰}$; James Ross Island mollusc (65°S lat.) $\delta_w = -3.5\text{‰}$ (see text). Symbol shapes for foraminiferal species are the same as those used in Figure 3. All Site 144 estimates are from *H. moremani* (squares) and *Hedbergella* sp. (inverted triangles). Site 144 data are from *Wilson et al.* [2002]. The James Ross Island datum is from *Ditchfield et al.* [1994].