Abrupt deep-sea warming at the end of the Cretaceous

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ABSTRACT

Climatic and oceanographic variations during the last 2 m.y. of the Maastrichtian inferred from high-resolution (10 k.y.) stable isotope analysis of the mid-latitude South Atlantic Deep Sea Drilling Project Site 525 reveal a major warm pulse followed by rapid cooling prior to the Cretaceous-Tertiary boundary. Between 66.85 and 65.52 Ma, cool but fluctuating temperatures average 9.9 and 15.4 °C in intermediate and surface waters, respectively. This interval is followed by an abrupt short-term warming between 65.45 and 65.11 Ma, which increased temperatures by 2–3 °C in intermediate waters, and decreased the vertical thermal gradient to an average of 2.7 °C. This warm pulse may be linked to increased atmospheric pCO_2, increased poleward heat transport, and the switch of an intermediate water source from high to low-middle latitudes. During the last 100 k.y. of the Maastrichtian, intermediate and surface temperatures decreased by an average of 2.1 and 1.4 °C, respectively, compared to the maximum temperature between 65.32 and 65.24 Ma.

INTRODUCTION

The well-documented long-term cooling trend that characterized the Late Cretaceous was terminated by a short-term warming followed by cooling near the end of the Maastrichtian (Stott and Kennett, 1990). Stable isotope studies of deep-sea sites in the Pacific and South Atlantic reveal this terminal Cretaceous warm event as a distinct 2–3 °C global warming in the deep sea and a 2–3 °C warming in surface waters in middle and high southern latitudes. Based on published data, the onset of this warming occurred near the base of Chron 29R, about 400–500 k.y. before the Cretaceous-Tertiary (K-T) boundary; the warming was terminated by an abrupt cooling 100–200 k.y. before the K-T boundary event (Stott and Kennett, 1990; Barrera, 1994; Li and Keller, 1998). However, different sample resolutions, condensed sedimentation, and short hiatuses at or near the K-T boundary make it difficult to determine the precise timing or nature of these terminal Cretaceous warming and cooling events.

The intriguing presence of a short-term warming at the end of the Maastrichtian, and subsequent cooling preceding the major mass extinction at the K-T boundary, has prompted us to take a closer look at the climatic and oceanographic changes during the last 2 m.y. of the Cretaceous. For this study, we chose DSDP (Deep Sea Drilling Project) Site 525 (Hole A), which has an excellent late Maastrichtian paleomagnetic record and apparently continuous sedimentation up to the K-T boundary clay layer. However, above the clay layer there is a hiatus, as indicated by the presence of a well-developed Zone P1c planktic foraminiferal fauna, which suggests that sediment of the first ~500 k.y. of the early Danian is missing. During the late Maastrichtian, the site was located on Walvis Ridge in the mid-latitude South Atlantic at 36°S paleolatitude, and at a paleodepth of about 1300 m (Chave, 1984; Moore et al., 1984).

MATERIAL AND METHODS

Our preliminary study of Site 525 at 1–1.5 m sample intervals revealed a continuous sediment record with well-preserved, little recrystallized foraminiferal tests, and little evidence of dissolution in all but one sample (Li and Keller, 1998). However, the similarity of the oxygen isotope records from DSDP Sites 525 and 463, and ODP (Ocean Drilling Program) Sites 690 and 689 (Stott and Kennett, 1990; Barrera, 1994; Barrera et al., 1997) indicates that original climatic signals are preserved.

We resampled the last 2 m.y. of the Cretaceous at Site 525 (Cores 40–41) every 10 cm (~10 k.y.), except for section 6 of Core 40, which was not recovered. Age estimates are based on the time scale of Cande and Kent (1995). Biostratigraphic zonation is based on the zonal scheme of Li and Keller (1998), with Zones CF1-2 and CF3 forming the upper part of the *Abathomphalus mayaroensis* Zone. Stable isotopes were measured on about 20–30 specimens of the surface-dwelling planktic foraminifer *Rugoglobigerina rugosa*, and the benthic foraminifer *Anomalinoides acuta* from the 150–250 µm size fraction and fine fraction (<38 µm). A total of 160 samples were analyzed for stable isotopes at Princeton University with a VG Optima gas-source mass spectrometer equipped with a common acid bath. To test analytical precision, standard calcite samples were spaced throughout the run and these yielded an average precision of 0.04‰ for $\delta^{18}O$ and 0.02‰ for $\delta^{13}C$. All isotopic results were calibrated to the PDB (Peedee belemnite) scale. Paleotemperature estimates are based on the equation by Erez and Luz (1983), with a value of –1.2‰ for seawater $\delta^{18}O$ for a largely ice-free world (Shackleton and Kennett, 1975).

RESULTS

Oxygen Isotopes

High-resolution stable isotope records reveal relatively stable $\delta^{18}O$ signatures between 66.8 and 65.6 Ma (Fig. 1); benthic and planktic $\delta^{18}O$ values vary mostly between 0.25‰–0.5‰ and –0.5‰ to –1.25‰, respectively. This relatively stable period was followed by larger $\delta^{18}O$ changes during the last 500 k.y. of the Maastrichtian. Beginning at 65.55 Ma, benthic $\delta^{18}O$ values gradually decreased, then abruptly dropped 0.5‰ at 65.45 Ma, and reached minimum values of –0.37‰ by 65.30 Ma. Thereafter, benthic values gradually increased about 0.6‰ toward the K-T boundary. Planktic $\delta^{18}O$ values do not mirror the pronounced benthic changes, but show fluctuations in the range of 0.2‰–0.4‰. However, the benthic $\delta^{18}O$ increase during the last 300 k.y. of the Maastrichtian was also paralleled by generally increasing planktic values (Fig. 1). Maximum surface $\delta^{18}O$ values for the 2 m.y. interval were reached about 40 k.y. before the K-T boundary (Fig. 1).

Paleotemperatures

$\delta^{18}O$ data from the benthic foraminifer *Anomalinoides acuta* at Site 525 at 1300 m paleodepth measure changes in intermediate-water temperatures (IWTs), whereas $\delta^{18}O$ data from the planktic foraminifer *Rugoglobigerina rugosa* measure sea-surface temperatures (SSTs, Fig. 2). Between 66.8 and 65.6 Ma, the IWT was relatively constant at about 10 °C, whereas the SST was more variable, fluctuating between 14 and 16 °C; thus, the surface-to-deep gradient (Δt) varied between 5 and 7 °C, based on a five-point running average (Figs. 1, 2). Between 65.55 and 65.47 Ma, intermediate waters gradually warmed by 1 °C, and

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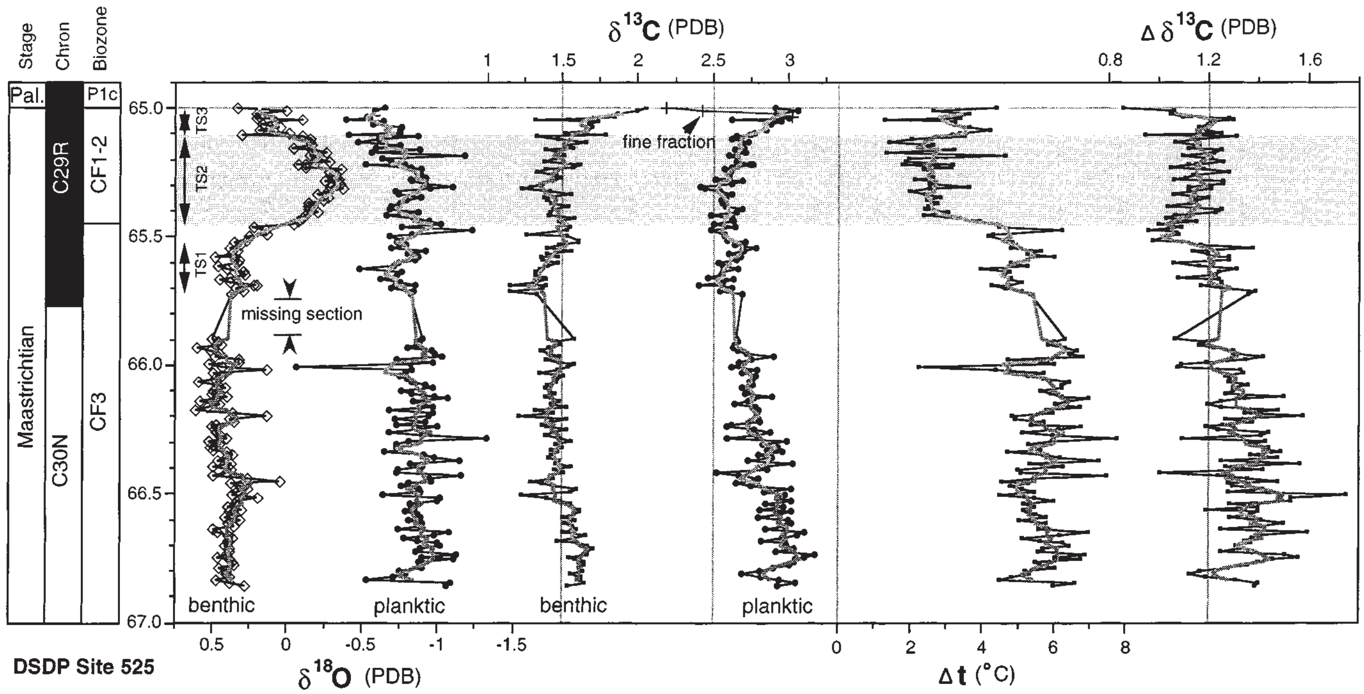

Figure 1. High resolution (10 k.y.) $\delta^{18}O$ and $\delta^{13}C$ records, surface-to-deep thermal (Δt), and carbon isotope ($\Delta\delta^{13}C$) gradients at DSDP Site 525. Magnetostratigraphy is from Chave (1984), biozonation is from Li and Keller (1998), and time scale is from Cande and Kent (1995). Light gray lines mark five-point running averages, and light gray band (TS2) marks period of climatic warming and very low vertical thermal gradients. Pal.—Paleocene, PDB—Peedee belemnite standard. Benthic isotopic values are from *Anomalinoides acuta* and planktic values from *Rugoglobigerina rugosa*.

SSTs varied between 14.5 and 15.5 °C. (A one-point maximum of 17.5 °C was omitted from the calculation of average SST values). During this time, the Δt decreased from 6 to 4.2 °C, except for a brief surface maximum (Fig. 2).

The major intermediate-water warming was marked by three steps (TS2a–TS2c, Fig. 2): Between 65.45 and 65.32 Ma (TS2a, n = 13), there was an abrupt 2.2 °C rise to a plateau averaging 12.3 °C; between 65.32 and 65.24 Ma (TS2b, n = 9), IWT reached maximum temperatures averaging 13.1 °C; and between 65.24 and 65.11 Ma (TS2c, n = 12), the IWT decreased to an average of 12.3 °C. Similarly pronounced changes are not observed in the SSTs, which generally show higher-frequency variations, particularly during the transition phases (TS2a and TS2c). Nevertheless, averaged SSTs indicate a slight warming between TS1 and TS2a from 15.0 to 15.2 °C, followed by another 0.5 °C warming in TS2b, and subsequent cooling of 0.9 °C in TS2c (Table 1, Fig. 2). During the last 100 k.y. of the Maastrichtian, the SST and IWT decreased to averages of 14.3 and 11 °C, respectively. During the warm pulse (65.45 and 65.11 Ma), the Δt averaged 2.7 °C as compared to 4.9 °C in TS1 and 3.3 °C in TS3 (Fig. 2).

Carbon Isotopes and Surface-to-Deep $\delta^{13}C$ Gradient

At Site 525, benthic and planktic $\delta^{13}C$ values show similar trends and generally covary, though with slightly greater fluctuations recorded in planktic values (Fig. 1). Between 66.8 and 65.6 Ma, benthic and planktic $\delta^{13}C$ values decreased gradually by 0.25‰ and 0.5‰, respectively, and the $\Delta\delta^{13}C$ values gradually decreased with an abrupt decrease of 0.25‰ at 65.55 Ma (Fig. 1). During the last 300 k.y. of the Maastrichtian, benthic and planktic values increased by 0.7‰ and 0.5‰, respectively; most of this increase was during the last 100 k.y. accompanied by a 0.5‰ decrease during the last 40 k.y. of the Maastrichtian (Figs. 1, 2). These data suggest only a minor decrease in surface productivity at 65.5 Ma, and a more significant decrease during the last 40 k.y. of the Maastrichtian.

However, the opposite signal is found in the measured fine-fraction $\delta^{13}C$ values for the last 40 k.y. of the Maastrichtian, which show a decrease of 1‰ and hence indicate more strongly reduced surface productivity (Fig. 2). Because the carbonate fine fraction (<38 μm) consists primarily of calcareous nannoplankton, it is possible that primary productivity was significantly reduced in this group, but not in the surface dwelling planktic foraminifers prior to the K-T boundary. Contradicting this interpretation is the absence of increased dissolution and constant high $CaCO_3$ values that suggest that there was no major overall change in the rate of $CaCO_3$ production in surface waters prior to the K-T boundary.

DISCUSSION

Paleoclimate

Relatively cool though fluctuating late Maastrichtian (66.85 and 65.52 Ma) temperatures averaging 9.9 °C in IWT and 15.4 °C in SST are indicated at the middle latitude South Atlantic Site 525 (Figs. 1, 2). This relatively cool period was interrupted by a short-term warming that increased IWTs from 10.1 °C to a maximum of 13.3 °C between 65.55 and 65.30 Ma. This maximum warm pulse was followed by a 2–3 °C cooling during the last 100–200 k.y. of the Maastrichtian. In contrast to the pronounced IWT warming, SSTs are more variable and reflect high climatic variability during the transition phases, though the average SST increased only by 0.5 °C during the maximum IWT warming (TS2b, Fig. 2).

However, surface temperatures only increased significantly in the southern and northern high latitudes, where they were accompanied by the incursion of subtropical planktic foraminifera (Huber and Watkins, 1992; Schmitz et al., 1992; Keller, 1993; Keller et al., 1993). The significantly increased SST previously observed in the $\delta^{18}O$ record at Site 525 based on low sampling resolution (1–1.5 m; Li and Keller, 1998) was not confirmed at the 10 cm sample spacing of this study and appears to have been a result of sample aliasing.

What are likely processes that could have triggered the short-term warming in the deep sea at high latitudes? Stott and Kennett (1990) suggested that the abrupt high-latitude warming near the end of the Maastrichtian may have resulted from changes in the carbon cycle. The abrupt decrease in surface productivity at 65.55 Ma at

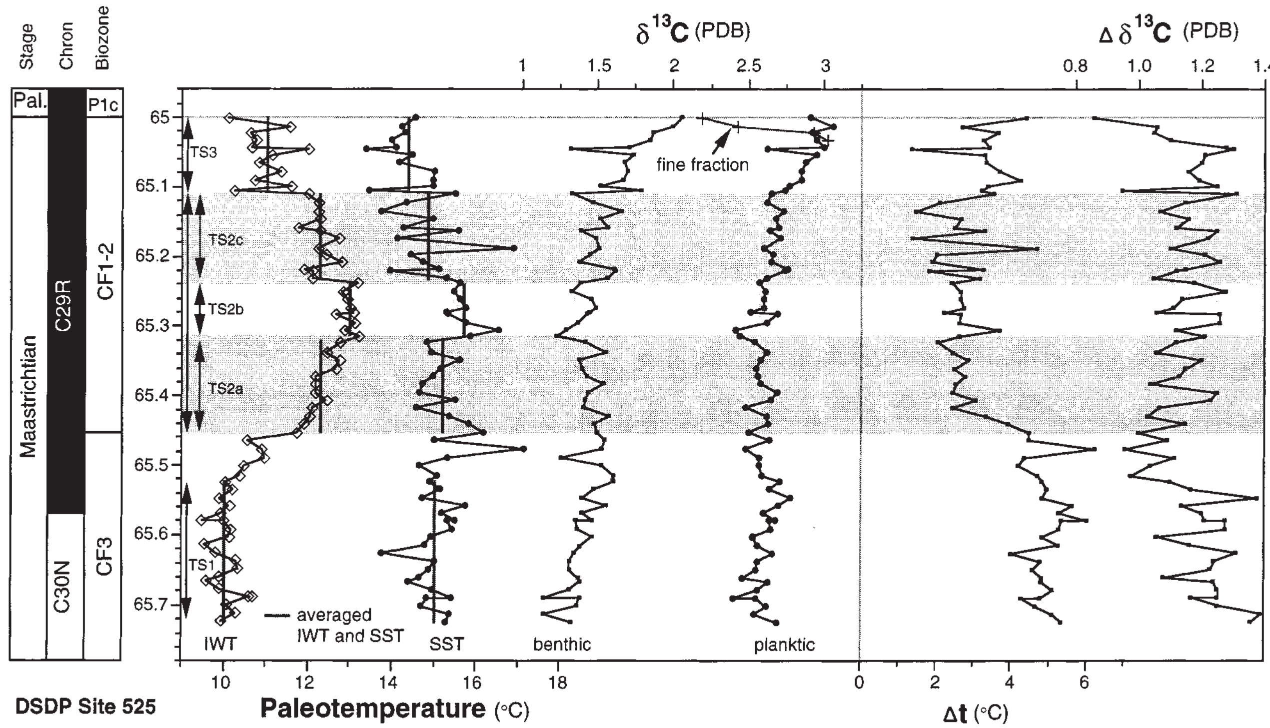

Figure 2. Expanded paleotemperature and δ^{13}C records of last 720 k.y. of Maastrichtian from Site 525. The short-term warming is marked by three stepped changes in IWT (gray band labeled TS2a–TS2c): (1) an abrupt 2.2 °C rise in TS2a, (2) maximum warming that reached temperatures of 13.1 °C in TS2b, and (3) cooling to 12.3 °C in TS2c. During this time interval, SST increased only slightly in TS2b. During last 100 k.y. of Maastrichtian, intermediate and surface waters cooled an average of 2.1 and 1.4 °C, respectively, compared to maximum warming in TS2b.

Site 525, as suggested by decreased Δδ^{13}C, implies that increased atmospheric pCO_2 is the most likely cause for the short-term warming. Deccan Trap volcanic degassing may have contributed significantly to the increased pCO_2 at this time. The major Maastrichtian Deccan Trap volcanic eruptions occurred in the lower part of Chron 29R, apparently coincident with the short-term warming at Site 525 (Courtillot et al., 1988, 1996; Jaeger et al., 1989; Baksi et al., 1994; Bhandari et al., 1995). However, climatic effects observed at Site 525 with major warming in IWTs and only minor warming in SSTs, though accompanied by high climatic variability, do not suggest the expected response of a greenhouse effect—namely the global warming of both surface and deep waters.

The observed SSTs at Site 525 and published data from low latitudes (Zachos et al., 1989) seem to support the cool tropics in the latest Cretaceous proposed by D'Hondt and Arthur (1996). This scenario is also suggested by the latitudinal SST gradient between Site 525 (36°S) and Site 690 (~65°S), which suggests a 33% decrease (from 0.18 to 0.12 °C per 1° of latitude, Table 1) between the late Maastrichtian cooling (TS1) and the subsequent warming (TS2), followed by a 17% increase to 0.14 °C per 1° of latitude during the end-Maastrichtian cooling (TS3). These data suggest significantly enhanced poleward heat transport during the short-term warming in TS2.

Paleoceanographic Implications

To evaluate changes in intermediate-water circulation, we estimated the paleodensity based on the model of Railsback et al. (1989), assuming a mean salinity of 34 parts per thousand (ppt) and the same salinity throughout the water column for the interval of TS3. Mean δ^{18}O values are used in this study (e.g., for specific intervals labeled TS1–TS3, Fig. 3) to avoid measured margin errors for single samples (TS1: $n = 21$; TS2: $n = 34$; TS3: $n = 12$).

Results show that intermediate-water signals in TS2 are significantly different from those in TS1 and TS3 (Fig. 3). The δ^{18}O of *A. acuta* indicates that the cool intermediate waters in TS1 were replaced by a 2.4 °C warmer and 1.6 ppt saltier and denser water mass during the short-term warming in TS2. During the cooling at the end of the Maastrichtian (TS3), this warmer, saltier, and denser intermediate water mass was replaced by a cooler, lower-salinity, and lower-density water mass with values close to TS1 (Fig. 3). These data suggest a different intermediate-water source during the short-term warming in TS2, which may have originated from shallow platform regions in low to middle latitudes. The switch from a high to a low-middle latitude source for the intermediate water mass during the climatic warming may be related to a eustatic transgression that resulted in increased shallow marginal-sea areas in low and middle latitudes. In contrast, values for surface paleotemperatures, salinity, and density inferred from *R. rugosa* show no major changes

TABLE 1. AVERAGED SST, IWT, AND LATITUDINAL GRADIENT AT THREE TIME SLICES AT THE END OF THE CRETACEOUS

Time slice (TS) (age in Ma)	Averaged SST (°C)	Averaged IWT (°C)	SST gradient at middle-high latitudes (°C per 1° latitude)	Site
TS3 (65.10–65.00)	14.3	11.0	0.14	DSDP 525
	10.3	9.0		ODP 690
TS2 (65.45–65.11)	15.2	12.5	0.12	DSDP 525
	11.7	9.3		ODP 690
TS2c (65.24–65.11)	14.8	12.3		DSDP 525
TS2b (65.32–65.24)	15.7	13.1		DSDP 525
TS2a (65.45–65.32)	15.2	12.3		DSDP 525
TS1 (65.72–65.52)	15.0	10.1	0.18	DSDP 525
	9.8	6.7		ODP 690

Note: ODP Site 690 data from Stott and Kennett (1990).

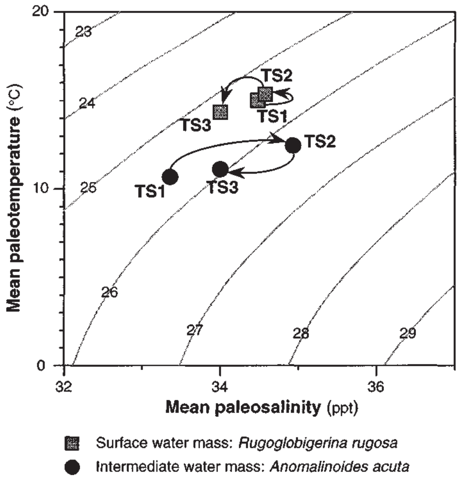

Figure 3. Mean paleotemperature-salinity profile with contours of density (curved gray lines) for the three time slices (TS) at the end of the Cretaceous at Site 525A after Railsback et al. (1989), density is expressed as σ_t = ($\rho_{seawater}$ − 1) × 1000, assuming an isohaline water column in TS3 (34 ppt for mean paleosalinity). Note that cool, low-salinity, and low-density intermediate-waters in TS1 were replaced by warmer, saltier, and denser waters during short-term warming in TS2; in TS3, the TS2 water mass was switched to a water mass with a temperature-salinity profile more similar to that of TS1. In contrast, surface waters remained relatively unchanged during these three time slices.

between TS1 and TS2, though they decreased slightly in TS3 (Fig. 3).

Water-mass stratification generally reflects the stability of the water column in the ocean. A rapid density increase with depth results in a more stable and well-stratified water column as compared with an inverse density profile. The water-mass structure during the warm event in TS2 is therefore likely to have been more stable as compared to the cool events in TS1 and TS3. This interpretation is also suggested by the low averaged vertical thermal gradient of 2.7 °C in TS2 as compared to 4.9 °C in TS1 and 3.3 °C in TS3. The higher intermediate-water density, warmer temperature, and lower latitudinal SST gradient may have resulted in a sluggish surface circulation during the short-term warming at the end of Cretaceous.

CONCLUSIONS

Climate changes during the last 2 m.y. of the Maastrichtian indicate a relatively cool mid-latitude South Atlantic interrupted by a short-term warming of 2–3 °C in IWTs at the base of Chron 29R, though surface temperatures increased only slightly. This short-term warming may have been related to increased atmospheric pCO_2, increased poleward heat transport, and a switch in the intermediate-water source from high to lower latitudes. During the last 100 k.y. of the Maastrichtian, intermediate and surface waters cooled by an average of 2.1 and 1.4 °C, respectively, compared to the maximum warming between 65.32 and 65.24 Ma.

Oxygen isotope records suggest that intermediate waters originated from a lower-latitude warm, saline, dense water source during the short-term warming between 65.45 and 65.11 Ma, and from a higher-latitude cold water source during the cool intervals between 65.72 and 65.50 Ma and 65.10 and 65.00 Ma.

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REFERENCES CITED

Barrera, E., 1994, Global environmental changes preceding the Cretaceous-Tertiary boundary: Early-late Maastrichtian transition: Geology, v. 22, p. 877–880.

Barrera, E., Savin, S. M., Thomas, E., and Jones, C. E., 1997, Evidence for thermohaline-circulation reversals controlled by sea-level change in the latest Cretaceous: Geology, v. 25, p. 715–718.

Baksi, A. K., 1994, Geochronological studies on whole-rock basalts, Deccan Traps, India: Evaluation of the timing of volcanism relative to the K-T boundary: Earth and Planetary Science Letters, v. 121, p. 43–56.

Bhandari, N., Shukla, P. N., Ghevariya, Z. G., and Sundaram, S. M., 1995, Impact did not trigger Deccan volcanism: Evidence from Anjar K/T boundary intertrappean sediments: Geophysical Research Letters, v. 22, p. 433–436.

Cande, S. C., and Kent, D. V., 1995, Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic: Journal of Geophysical Research, v. 100, p. 6093–6095.

Chave, A. D., 1984, Lower Paleocene-Upper Cretaceous magnetostratigraphy, Sites 525, 527, 528, and 529, Deep Sea Drilling Project Leg 74, *in* Moore, T. C., Jr., Rabinowitz, P. D., et al., eds., Initial reports of the Deep Sea Drilling Project, Volume 74: Washington, U.S. Government Printing Office, p. 525–531.

Courtillot, V., Feraud, G., Maluski, H., Vandamme, D., Moreau, M. G., and Besse, J., 1988, Deccan flood basalts and the Cretaceous/Tertiary boundary: Nature, v. 333, p. 843–846.

Courtillot, V., Jaeger, J. J., Yang, Z., Feraud, G., and Hofmann, C., 1996, The influence of continental flood basalts on mass extinctions; where do we stand? *in* Ryder, G., Fastovosky, D., and Gartner, S., eds., The Cretaceous-Tertiary event and other catastrophes in Earth history: Geological Society of America Special Paper 307, p. 513–525.

D'Hondt, S., and Arthur, M. A., 1996, Late Cretaceous oceans and the cool tropic paradox: Science, v. 271, p. 1838–1841.

Erez, J., and Luz, B., 1983, Experimental paleotemperature equation for planktonic foraminifera: Geochimica et Cosmochimica Acta, v. 47, p. 1025–1031.

Huber, B., and Watkins, D. K., 1992, Biogeography of Campanian-Maastrichtian calcareous plankton in the region of the southern ocean: Paleogeographic and paleoclimatic implications, *in* Kennett, J. P., and Warnke, D. A., eds., The Antarctic paleoenvironment: A perspective on global change: Washington, D.C., American Geophysical Union, Antarctic Research Series, v. 56, p. 31–60.

Jaeger, J. J., Courtillot, V., and Tapponier, P., 1989, Paleontological view of the ages of the Deccan Traps, the Cretaceous/Tertiary boundary, and the India-Asia collision: Geology, v. 17, p. 316–319.

Keller, G., 1993, The Cretaceous/Tertiary boundary transition in the Antarctic Ocean and its global implications: Marine Micropaleontology, v. 21, p. 1–45.

Keller, G., Barrera, E., Schmitz, B., and Mattson, E., 1993, Gradual mass extinction, species survivorship, and long-term environmental changes across the Cretaceous/Tertiary boundary in high latitudes: Geological Society of America Bulletin, v. 105, p. 979–997.

Li, L., and G., Keller, 1998, Maastrichtian climate, productivity and faunal turnovers in planktic foraminifera in South Atlantic DSDP Sites 525 and 21: Marine Micropaleontology, v. 33, p. 55–86.

Moore, T. C., Jr., Rabinowitz, P. D., Borella, P. E., Shackleton, N. J., and Boersma, A., 1984, History of the Walvis Ridge, *in* Moore, T. C., Jr., Rabinowitz, P. D., et al., eds., Initial reports of the Deep Sea Drilling Project, Volume 74: Washington, D.C., U.S. Government Printing Office, p. 873–894.

Railsback, L. B., Anderson T. F., Ackerly, S. C., and Cisne, J. L., 1989, Paleoceanographic modeling of temperature-salinity profiles from stable isotopic data: Paleoceanography, v. 4, p. 585–591.

Schmitz, B., Keller, G., and Stenvall, O., 1992, Stable isotope and foraminiferal changes across the Cretaceous/Tertiary boundary at Stevns Klint, Denmark: Arguments for long-term oceanic instability before and after bolide impact: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 96, p. 233–260.

Shackleton, N. J., and Kennett, J. P., 1975, Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analyses in DSDP Sites 277, 279, and 281, *in* Kennett, J. P., Houtz, R. E., et al., eds., Initial reports of the Deep Sea Drilling Project, Volume 29: Washington, D.C., U.S. Government Printing Office, p. 743–755.

Stott, L. D., and Kennett, J. P., 1990, The paleoceanographic and paleoclimatic signature of the Cretaceous/Paleogene boundary in the Antarctic: Stable isotopic results from ODP Leg 113, *in* Barker, P. F., Kennett, J. P., et al., eds., Proceedings of the Ocean Drilling Program, Scientific Results, Volume 113: College Station, Texas, Ocean Drilling Program, p. 829–848.

Zachos, J. C., Arthur, M. A., and Dean, W. E., 1989, Geochemical evidence for suppression of pelagic marine productivity at the Cretaceous/Tertiary boundary: Nature, v. 337, p. 61–64.

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