Tectonic forcings of Maastrichtian ocean-climate evolution

Tracy D. Frank
University of Nebraska-Lincoln, tfrank2@unl.edu

Michael A. Arthur
Pennsylvania State University

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Tracy D. Frank and Michael A. Arthur

Department of Geosciences, Pennsylvania State University, University Park, PA

Abstract

A global compilation of deep-sea isotopic records suggests that Maastrichtian ocean-climate evolution was tectonically driven. During the early Maastrichtian the Atlantic intermediate-deep ocean was isolated from the Pacific, Indian, and Southern Oceans; deep water formed in the high-latitude North Atlantic and North Pacific. At the early/late Maastrichtian boundary a major reorganization of oceanic circulation patterns occurred, resulting in the development of a thermohaline circulation system similar to that of the modem oceans. A combination of isotopic and plate kinematic data suggests that this event was triggered by the final breaching of tectonic sills in the South Atlantic and the initiation of north-south flow of intermediate and deep water in the Atlantic. The onset of Laramide tectonism during the mid Maastrichtian led to the concurrent draining of major epicontinental seaways. Together, these events caused cooling, increased latitudinal temperature gradients, increased ventilation of the deep ocean, and affected a range of marine biota.

1. Introduction

Although the K/T boundary interval has long been the focus of most studies of the Maastrichtian, many workers have begun to investigate the paleoceanographic and climatic conditions during the several million years preceding the boundary event. In this regard the recent recognition of roughly coeval changes in biotic, oceanographic, and climatic indicators has served to focus attention on the mid Maastrichtian. Notable mid Maastrichtian biologic events include the extinction of inoceramids [e.g., Dhondt, 1983, 1992; Kauffman, 1988; MacLeod, 1994], the loss of Caribbean-Tethyan rudist reef ecosystems [Johnson and Kauffman, 1990, 1996; Swinburne, 1990], and shifts in the latitudinal distributions of calcareous nanoplankton and planktonic foraminifera [Huber, 1992; Huber and Watkins, 1992]. Fluctuations in $\delta^{18}O$ and $\delta^{13}C$ values of planktonic and benthic foraminifera from sites in the tropical Pacific and high southern latitudes have led workers to link mid Maastrichtian extinctions to a reversal in thermohaline circulation patterns, although the driving mechanism, mode, and direction of this reversal are debated [e.g., MacLeod and Huber, 1996; Barrera et al., 1997].

A poor understanding of the causal mechanisms behind mid Maastrichtian oceanographic and biologic events may, in part, reflect the propensity of many workers to (1) base their interpretations upon records derived from a single global reservoir, geographic realm, and/or single indicator or (2) focus their efforts upon a narrow time slice with little consideration of changes that led up to or resulted from mid Maastrichtian events. Few workers have attempted to merge the available isotopic, lithologic, and paleontologic data sets to evaluate possible interactions among the various global reservoirs. In this regard we compiled and correlated previously published Maastrichtian stable isotope records from cores recovered from a global array of Deep Sea Drilling Project (DSDP) and the Ocean Drilling Program (ODP) sites (Figures 1 and 2). Because whole-rock and multitaxon isotopic records can be affected by variable vital effects and local partitioning of marine isotopic reservoirs, we relied on records derived from analyses of monospecific planktonic and monogeneric, epifaunal benthic foraminifera. A lack of published stable isotope data from Maastrichtian sediments of the Atlantic led us to generate additional records from DSDP site 357, located on the northern side of the Rio Grande Rise in the South Atlantic, DSDP site 390, situated in the North Atlantic on the Blake Nose, and DSDP site 548, located in the North Atlantic on the Goban Spur. Fluctuations chronicled in stable isotope data were interpreted in a context of changes recognized in records derived from a range of independent paleoceanographic and climatic indicators. This approach allowed the relationships between events and resulting feedbacks within and among various global reservoirs to be evaluated and, ultimately, synthesized into a more complete picture of the oceanic and climatic conditions that characterized the Maastrichtian.

2. Methods

2.1. Correlation and Chronostratigraphy

Direct comparison of published data sets has long been hampered by the use of different subdivisions and timescales that have been proposed for the Maastrichtian and the Cretaceous in general [e.g., Gradstein et al., 1994, Figure 1]. Such differences are derived, in part, from the diverse methods used in developing the various timescales. The application of Maastrichtian biostratigraphies is particularly complex because of poorly established correlations between high- and low-latitude zonation schemes [e.g., Huber, 1992] and the diachronous natures of some first- and last-occurrence events. To avoid these problems, we have developed an integrated chronostratigraphic framework
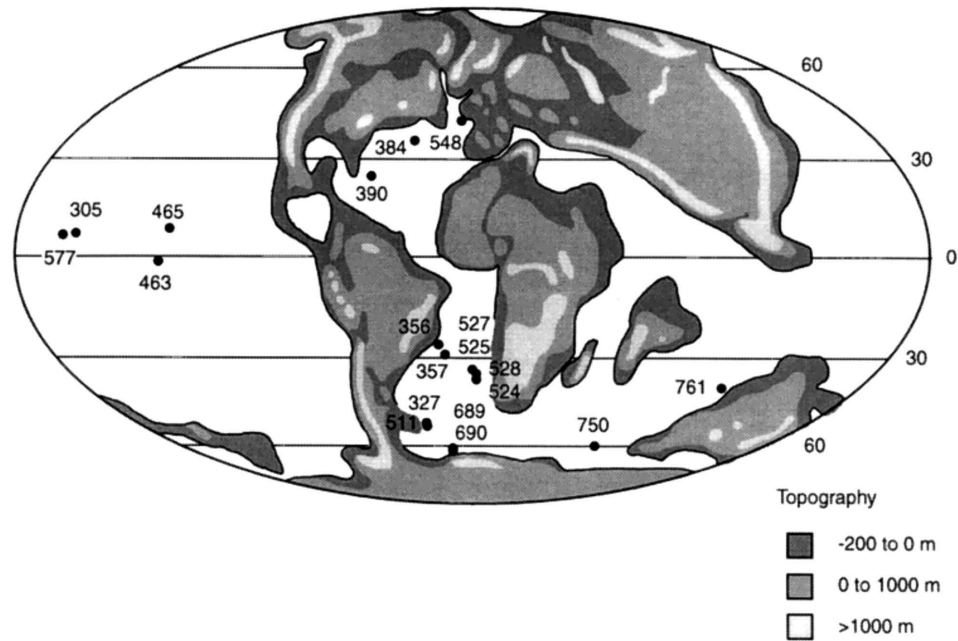
Figure 1. Paleogeographic reconstruction for the mid Maastrichtian showing continental topography and locations of Deep Sea Drilling Project (DSDP) and Ocean Drilling Program (ODP) sites. Modified from Patzkowsky et al. [1991].

(Figure 3) that is largely independent of biostratigraphy and based on the magnetostratigraphic timescale of Gradstein et al. [1994]. Isotopic data from sites for which magnetostratigraphic data were available (Table 1) were tied directly into the Gradstein et al. [1994] timescale by assuming constant rates of sedimentation between magnetic polarity zone boundaries. The strontium isotope curve, tied into the chronostratigraphic framework using Sr isotope and magnetostratigraphic data from DSDP and ODP sites 690, 689, 525, and 528 (Figure 3), provided a second means of cor-

relation. In cases where neither magnetostratigraphic nor Sr isotope data were available, isotopic records were correlated using regional biostratigraphies that have been correlated to the chronostratigraphic framework using either magnetostratigraphy or Sr isotope data (Figure 3). In such cases it was assumed that biostratigraphic zonal boundaries are synchronous within a given region. Ultimately, the highest temporal resolution, ranging from <0.5 to no more than 1.5 m.y., was achieved in cases where some combination of the above methods could be used for correlation.
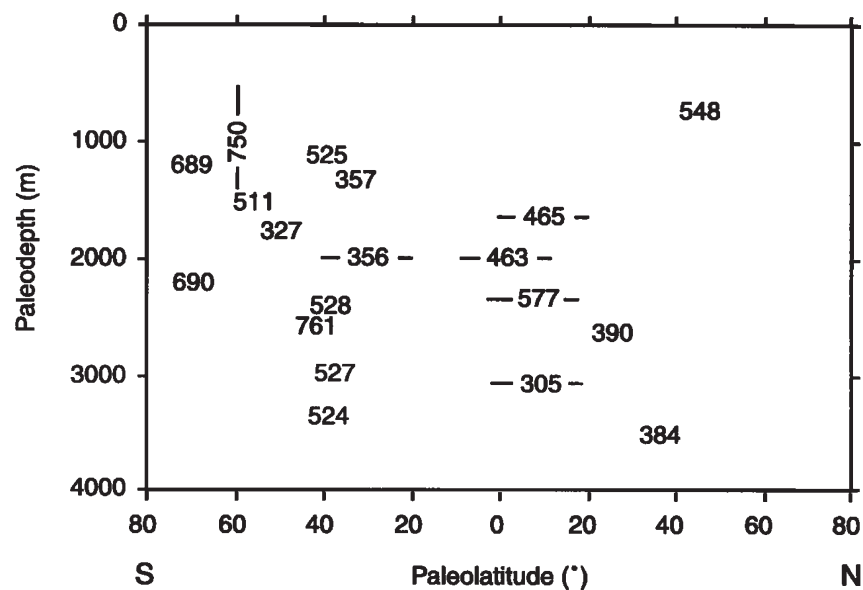

Figure 2. Estimated paleolatitude and paleodepth for DSDP and ODP sites. Maastrichtian paleodepth and paleolatitude estimates for sites 357, 384, 390, 465, 525, 527, 528, 689, and 690 are after D'Hondt and Arthur [1996]. Estimates for the remaining sites are based on sedimentary data in DSDP and ODP reports.
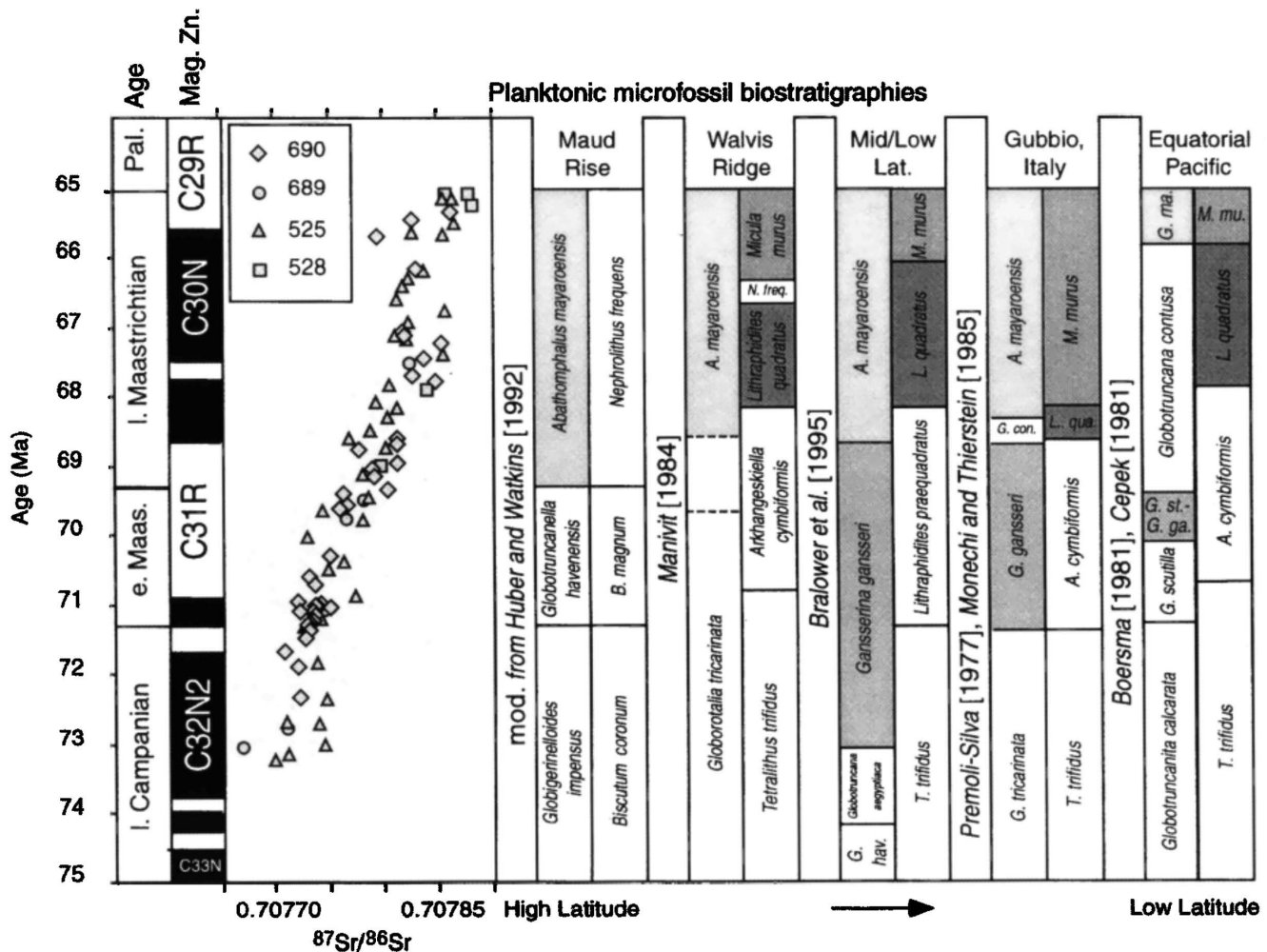
Figure 3. Chronostratigraphic framework used in this study, based on the magnetostratigraphic timescale of Gradstein et al. [1994]. Included are Sr isotope data and biostratigraphies that have been correlated using magnetopolarity data.

This resolution is significantly better than that (2-6 m.y.) provided by most planktonic foraminifer and calcareous nanofossil biostratigraphies for the Maastrichtian (Figure 3) [Bralower et al., 1995].

2.2. Sample Preparation and Analysis

Bulk sediment samples from sites 357, 390, and 548 were soaked in a 5.25% sodium hypochlorite (NaOCl) solution for 24 hours. Samples were then repeatedly rinsed in deionized water and placed in an ultrasonicator until samples were fully disaggregated and cleaned. Samples were oven-dried at <50°C. For carbon and oxygen isotope analysis, well-preserved foraminifera from the 250-354 and 354-420 μm size fractions were reacted at 90°C with anhydrous phosphoric acid in an automated carbonate device (common acid bath) coupled to a Finnigan-MAT 252 mass spectrometer. Isotopic ratios were corrected for ^{17}O contribution [Craig, 1957] and are reported in per mil (‰) relative to the Vienna Peedee belemnite (VPDB) standard; fractionation factors used to calculate the $\delta^{18}O$ and $\delta^{13}C$ values of carbonate are 0.98793 and 1.0052, respectively. Precision is better than 0.10‰ for $\delta^{13}C$ and $\delta^{18}O$ values and was monitored through multiple analyses of National Bureau of Standards (NBS) 19 and other powdered calcite standards.[1]

Identification of planktonic species is based on D'Hondt and Arthur [1995]; benthic genera were identified using Loeblich and Tappan [1988] and numerous DSDP and ODP reports. Diagenetic alteration, often subtle and not readily

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1 The new data reported in this paper are available electronically at World Data Center–A for Paleoclimatology, NOAA/NGDC, 325 Broadway, Boulder, Colorado (email paleo@mail.ngdc.noaa.gov; URL http://www.ngdc.noaa.gov/paleo/contrib-list.html).

Table 1. Site Location, Depth, Available Geochemical and Magnetostratigraphic Information, and Data Sources

| Site	Present-Day Latitude, Longitude	Present-Day Depth, m	$\delta^{18}O$, $\delta^{13}C$	$^{87}Sr/^{86}Sr$	Magneto-stratigraphy	Data Sources
689 (Maud Rise)	64° 31′ S, 3° 06′ E	2080	x	x	x	Barrera and Huber [1990] Stott and Kennett [1990]
690 (Maud Rise)	65° 10′ S, 1° 12′ E	2914				Barrera et al. [1997] Hamilton [1990]
750 (Kerguelen Plateau)	54° 36′ S, 81° 14′ E	2041	x			MacLeod and Huber [1996]
327 (Falkland Plateau)	56° 35′ S, 65° 18′ W	3812	x			Huber et al. [1995]
511 (Falkland Plateau)	51° 00′ S, 46° 58′ W	2589				
356 (São Paulo Plateau)	28° 17′ S, 41° 05′ W	3175	x	x		Boersma [1984] Martin and MacDougall [1991]
524 (Cape Basin)	29° 30′ S, 3° 31′ E	4796	x		x	He et al. [1984] Chave [1984]
525 (Walvis Ridge)	29° 04′ S, 2° 59′ E	2467	x	x	x	Sugarman et al. [1995] Chave [1984]
527 (Walvis Ridge)	28° 02′ S, 1° 46′ E	4428	x		x	Lee and Keller [1998] Shackleton et al. [1984] Chave [1984]
528 (Walvis Ridge)	28° 32′ S, 2° 19′ E	3800	x	x	x	D'Hondt and Lindinger [1994] Martin and MacDougall [1991] Chave [1984]
357 (Rio Grande Rise)	30° 25′ S, 35° 34′ W	2086	x			This study Boersma [1984]
390 (Blake Nose)	30° 09′ N, 76° 07′ W	2665	x			This study D'Hondt and Arthur [1995]
384 (J-Anomaly Ridge)	40° 22′ N, 51° 40′ W	3909	x	x		Boersma [1984] Martin and MacDougall [1991]
463 (Mid-Pacific Mts.)	21° 21′ N, 174° 40′ E	2525	x	x		Barrera et al. [1997]
465 (Hess Rise)	33° 49′ N, 178° 55′ E	2161	x			Boersma and Shackleton [1981]
482 (Shatsky Rise)	32° 25′ N, 158° 01′ E	2619	x			Boersma and Shackleton [1981]
171 (Horizon Guyot)	19° 08′ N, 169° 28′ W	2290	x			Doulas and Savin [1973, 1978]
548 (Goban Spur)	48° 55′ N, 12° 10′ W	1251	x		x	This study Townsend [1985]
577 (Shatsky Rise)	32° 27′ N, 157° 44′ E	2675	x	x		Zachos [1988] Martin and MacDougall [1991] Hess et al. [1986]
761 (Exmouth Plateau)	16° 44′ S, 115° 32′ E	2168	x	x		MacLeod and Huber [1996]
Gubbio, Italy	43° 13′ N, 12° 21′ E	n/a			x	Alvarez et al. [1977]
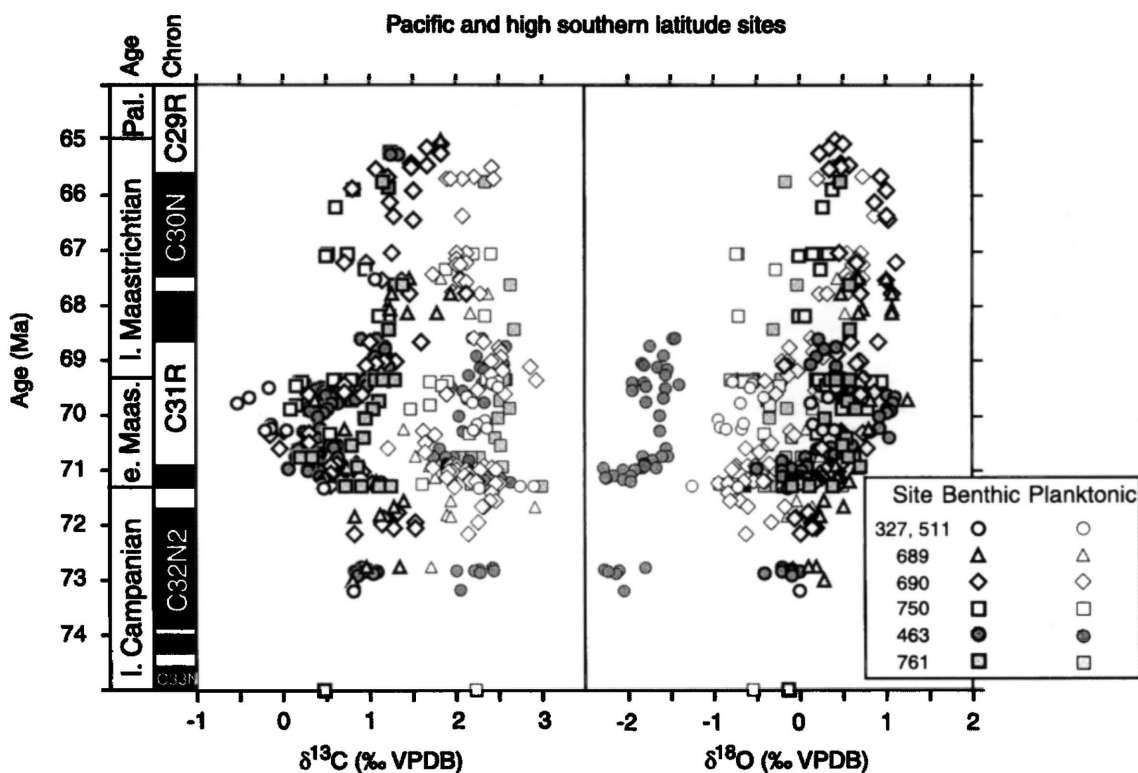
Figure 4. Correlation of benthic and corresponding planktonic (a) carbon and (b) oxygen isotope records from DSDP and ODP sites in the paleoequatorial Pacific and high southern paleolatitudes. Data are derived from benthic foraminiferal species *Gavelinella* and *Nuttalides* (all sites) and the planktonic foraminifera *Archeoglobigerina australis* (sites 689, 690, 750, 327, and 511), *Pseudoguembelina excolata* (site 463), and *Rugoglobigerina rugosa* (site 761).

visible petrographically, is a major confounding factor in interpreting oxygen isotope data from deep-sea cores [e.g., Schrag et al., 1992]. Excellent preservation of benthic and planktonic foraminifera from site 390 is suggested by a lack of calcitic overgrowth and the retention of fine-scale skeletal features on outer and inner shell surfaces when viewed in the scanning electron microscope (SEM). Evidence of minor recrystallization was observed in some specimens from sites 357 and 548. However, retained compositional differences between benthic and planktonic taxa and the consistency of isotopic trends among sites suggest that diagenetic modification of original oxygen isotope compositions was minor.

3. Stable Isotope Records

3.1. Tropical Pacific and High Southern Latitudes

Previous paleoceanographic models for the Maastrichtian are largely based on isotopic records from sites in the tropical Pacific, Southern Ocean, and high latitude South Atlantic (Figure 4) [Barrera et al., 1987, 1997; Barrera and Huber, 1990; Barrera, 1994; D'Hondt and Lindinger, 1994; Huber et al., 1995; MacLeod and Huber, 1996]. A general trend toward more positive $\delta^{18}O$ values with decreasing age in these planktonic and benthic foraminiferal records has been interpreted to reflect overall cooling of surface and deep waters through the Maastrichtian (Figure 4) [Barrera and Huber, 1990; Barrera, 1994]. Decreases in $\delta^{18}O$ and $\delta^{13}C$ surface

to deep water gradients throughout the late Maastrichtian, most notably at high southern latitude sites, have been inferred on the basis of corresponding planktonic and benthic foraminiferal records from individual sites (Figure 4). Superimposed on these general trends are prominent, parallel excursions in the carbon and oxygen isotope compositions of benthic and, at some sites, planktonic taxa [Barrera, 1994; Barrera et al., 1997], including 1-2‰ negative excursions in $\delta^{13}C$ and up to 1‰ positive excursions in $\delta^{18}O$ values (Figure 4). These excursions begin just above the Campanian/Maastrichtian boundary (base of magnetic polarity zone 32N1) and end near the lower/upper Maastrichtian boundary (upper part of magnetic polarity zone 31R).

3.2. Atlantic

The early Maastrichtian excursions in foraminiferal $\delta^{18}O$ and $\delta^{13}C$ values prominent at tropical Pacific and high southern latitude records are not evident in data from Atlantic sites (Figure 5). Rather, foraminiferal $\delta^{18}O$ and $\delta^{13}C$ values from Atlantic sites exhibit slight increases in lower Maastrichtian sections and are relatively constant in the upper Maastrichtian. One aspect that records from Atlantic and other sites have in common, however, is a gradual decrease in $\delta^{18}O$ and $\delta^{13}C$ surface to deep water gradients through the late Maastrichtian (Figures 4 and 5).

MacLeod et al. [1998] documented an early Maastrichtian negative excursion in benthic foraminiferal $\delta^{13}C$ values at site 390 [MacLeod et al., 1998, Figures 5 and 8]. This ex-
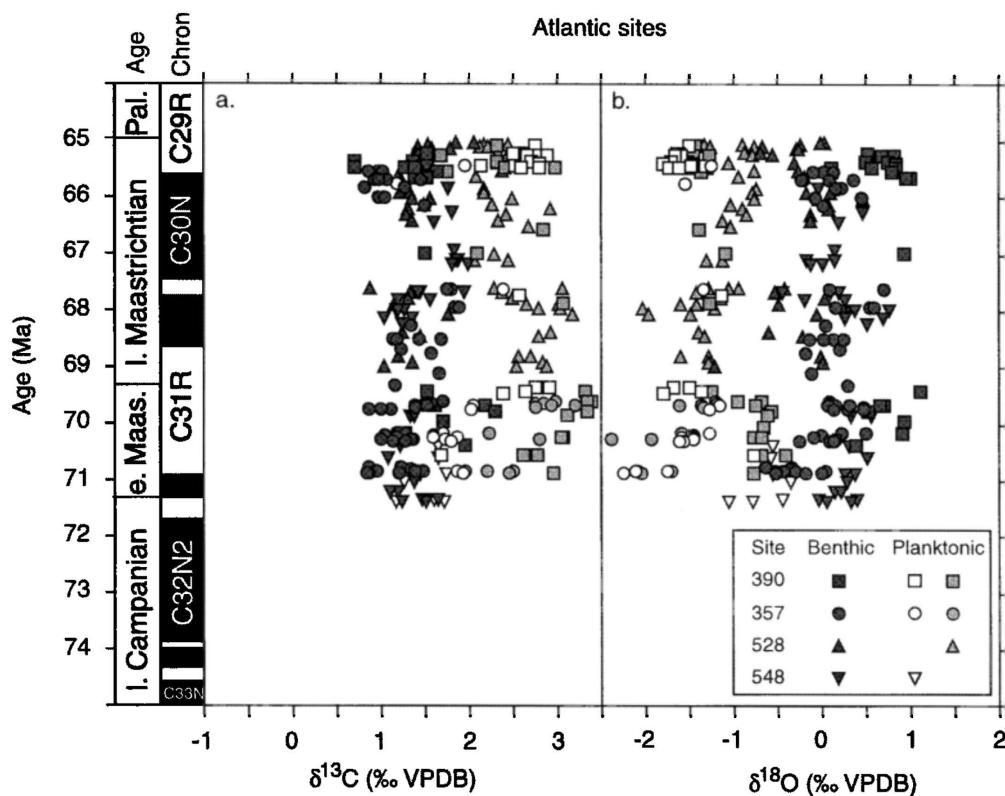
Figure 5. Correlation of benthic and corresponding planktonic (a) carbon and (b) oxygen isotope records from DSDP sites in the Atlantic. Data are derived from benthic foraminifera species *Gavelinella* and *Nuttalides* (all sites; dark shaded symbols) and the planktonic foraminifera *Rugoglobigerina rotundata* at sites 357, 390, and 548 (open symbols) and *Pseudotextularia elegans* and *Pseudotextularia ultimatudima* at sites 357, 390, and 528 (light shaded symbols).

cursion is defined by a single data point generated by the analysis of a sample composed of mixed benthic taxa. We, in contrast, focused our analyses on the benthic foraminifera *Gavelinella* and *Nuttalides*, which apparently precipitated their shells in carbon isotopic equilibrium with dissolved inorganic carbon (DIC) in the intermediate or bottom waters in which they formed [Shackleton et al., 1984]. Moreover, we sampled at a higher resolution than did MacLeod et al. [1998]. On this basis we dismiss the negative $\delta^{13}C$ excursion documented previously at site 390 as a sampling artifact.

4. Discussion

In the modem oceans a decrease in the $\delta^{13}C$ value of DIC in bottom water as it flows from the Atlantic to the Indian Ocean and North Pacific reflects the progressive addition of isotopically light CO_2 derived from the oxidation of organic matter [e.g., Kroopnick, 1985]. As such, the $\delta^{13}C$ value of bottom water DIC serves as a rough indicator of oceanic circulation patterns. Assuming the carbon isotope compositions of the benthic foraminifera *Nuttalides* and *Gavelinella* accurately reflect the composition of DIC in the waters in which they grew [Shackleton et al., 1984], temporal change in benthic foraminiferal $\delta^{13}C$ values at any particular site is interpreted primarily to signal either a modification in the distribution of carbon among global

reservoirs or a change in the character or source of the water mass bathing that site. Given the depths (<2500 m) of most sites discussed in this study (Figure 2), an important consideration is that carbon isotope data from benthic foraminifera are primarily recording water mass aging effects in intermediate water masses. Few reliable stable isotope records from paleodepths >2500 m are available, primarily because of a relatively shallow carbonate compensation depth (CCD) during much of the Late Cretaceous [Thierstein, 1979; Arthur et al., 1985] and subsequent poor preservation of foraminiferal calcite. Additional influences that could have had a minor effect on benthic foraminiferal $\delta^{13}C$ records (e.g., pCO_2 and local differences in paleoproductivity [Mackensen et al., 1994]) could not be constrained because of the sparse nature of the Maastrichtian record.

Given these caveats, one of the most prominent features of the global Maastrichtian carbon isotope record is temporal change in the degree of variation in benthic foraminiferal $\delta^{13}C$ values among sites (Figure 6). To facilitate discussion, we have divided the Maastrichtian into three time slices on the basis of distinct changes in the intersite benthic foraminiferal $\delta^{13}C$ gradient (time 1, time 2, and time 3; Figure 6). Time 1 encompasses the early Maastrichtian and is characterized by the largest $\delta^{13}C$ gradient (up to 3‰). Time 2 extends from the early/late Maastrichtian boundary to the middle of Magnetochron 30R. The base of time 2 marks an abrupt decrease in the benthic foraminiferal $\delta^{13}C$ gradient,
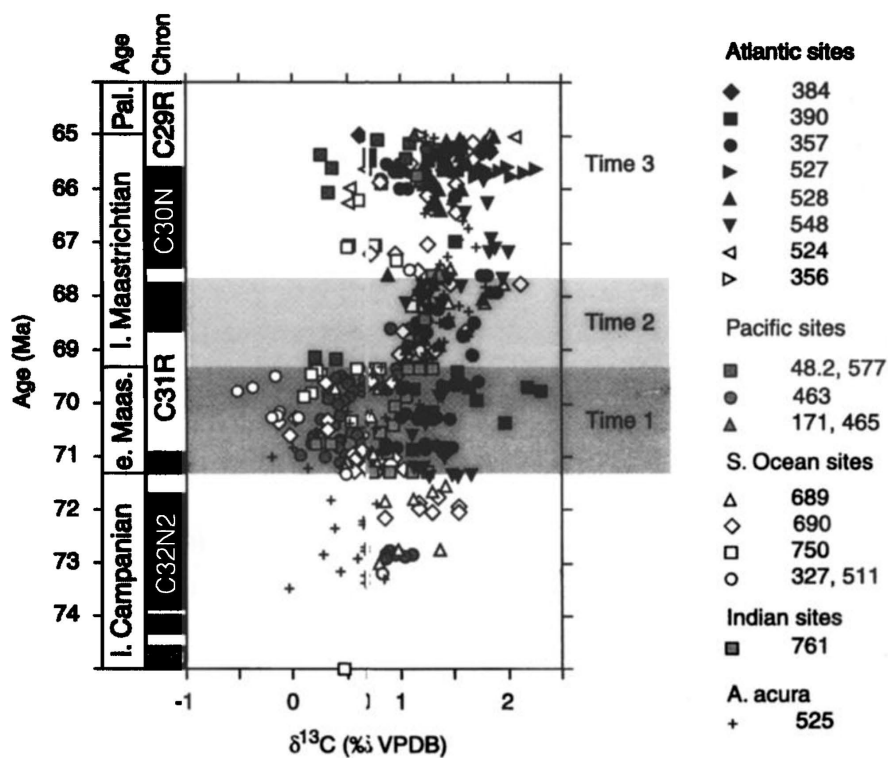
Figure 6. Compilation of carbon isotope data from the benthic foraminifera *Gavelinella* and *Nuttalides*. Time slices 1, 2, and 3 are discussed in the text.

which remains at <1‰ throughout time 2 (Figure 6). Time 3 spans the remainder of the Maastrichtian and is characterized by a gradual reestablishment of an interocean benthic foraminiferal $\delta^{13}C$ gradient, which increases to nearly 2.5‰ in the upper part of Magnetochron 30N (Figure 6). Unfortunately, there are insufficient data from the Campanian to constrain pre-early Maastrichtian conditions.

The large interocean $\delta^{13}C$ gradient during time I arises from differences in benthic foraminiferal carbon isotope records from sites in the tropical Pacific, southern Indian Ocean, and the Southern Ocean as opposed to those from Atlantic sites 357, 390, and 548. Whereas a prominent negative excursion in benthic foraminiferal $\delta^{13}C$ values characterizes lower Maastrichtian records from the former (Figure 4), corresponding $\delta^{13}C$ values from sites in the Atlantic (sites 357, 390, and 548) are generally more positive and exhibit little variation throughout the Maastrichtian (Figure 5). These differences suggest that during the early Maastrichtian, sites in the Pacific, Indian Ocean, and Southern Ocean shared a water mass that was separate from that which bathed sites 357, 390, and 548. During this time the largest differences among early Maastrichtian benthic foraminiferal $\delta^{13}C$ records occur among sites that were separated by large east-west trending tectonic features in the South Atlantic (Figure 6). The most negative benthic foraminiferal $\delta^{13}C$ values (derived from *Gavelinella* and *Nuttalides*) are found at relatively shallow sites 327 and 511, located on the southern side of the Falkland Plateau, whereas the most positive values are found at Atlantic sites to the north, including site 357, located on the northern side of the Rio Grande Rise (Figure 7). Although not directly compara-

ble because of possible vital effects, a $\delta^{13}C$ record derived from the benthic foraminifer *Anomalinoides acuta* at site 525 (Figure 6), located near the top of a south facing region of the Walvis Ridge (Figure 7), also exhibits a negative excursion of nearly 1‰ during time 1 [Li and Keller, 1998]. These differences suggest that during time 1 the Falkland Plateau in the Southern Ocean and the Walvis Ridge-Rio Grande Rise system in the southern South Atlantic may have effectively blocked the northward flow of deep and intermediate waters from the southern high latitudes and that lower $\delta^{13}C$ values at relatively shallow sites on the southern sides of such features (e.g., sites 327, 511, and 525) reflect the upwelling of deep waters as they met these barriers. This interpretation is consistent with Late Cretaceous tectonic reconstructions (Figure 7) [Sclater et al., 1977; Thiede, 1977].

At the early/late Maastrichtian boundary (base of time 2), benthic foraminiferal $\delta^{13}C$ values from tropical Pacific and high southern latitude sites shift by as much as 1.5‰ to approach the higher $\delta^{13}C$ values that characterize sites 357, 390, and 548 (Figure 6). The $\delta^{13}C$ values from *A. acuta* at site 525 exhibit a similar increase, although slightly earlier (Figure 6). As a result, differences in benthic $\delta^{13}C$ values between sites on the northern flanks of east-west ridges in the Atlantic (e.g., site 357) and their southern counterparts (sites 327 and 511) disappear. This could reflect either (1) the development of multiple point sources of bottom water or (2) an increase in the rate of overturn and/or an overall increase in dissolved oxygen levels. As such, the disappearance of the intersite benthic foraminiferal $\delta^{13}C$ gradient at the base of time 2 (Figure 6) suggests a reorganization of oceanic circulation patterns that resulted in a
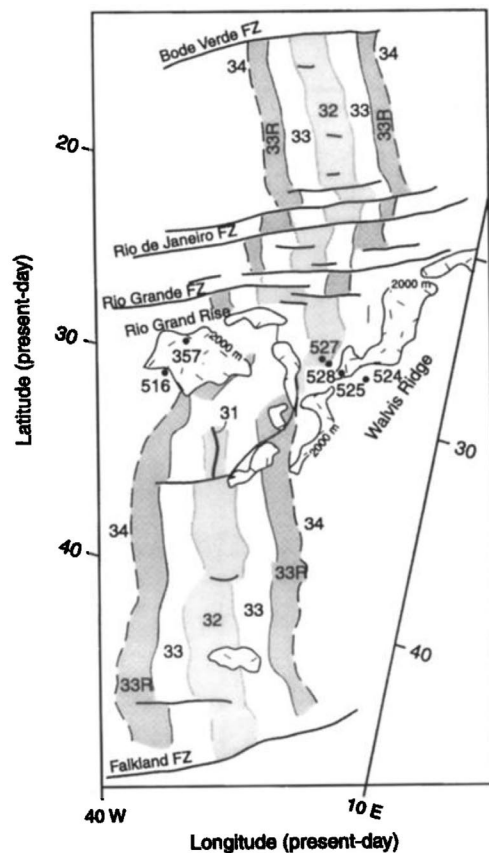
Figure 7. Reconstruction of the South Atlantic Basin at the beginning of Magnetochron 31R (early Maastrichtian) based upon the tectonic map of Cande et al. [1988]. The 2000 m contour delineates major topographic features (shaded areas), substantial portions of which were within a few hundred meters of sea level.

global homogenization of the character and carbon isotope composition of intermediate and deep waters.

Plate kinematic data from the South Atlantic suggest that the mid Maastrichtian reorganization of oceanic circulation patterns was triggered by the final breaching of major, east-west trending tectonic sills (Walvis Ridge-Rio Grande Rise complex) in the South Atlantic Basin (Figure 7). By late Maastrichtian time (Magnetochron 30), global seafloor spreading rates decreased from a high of ~75 mm y^{-1} during Magnetochron 34 (Aptian-Santonian) to 30 mm y^{-1} [Cande et al., 1988]. This decrease resulted in a global decrease in the rate of oceanic crust production (Figure 8) [Larson, 1991]. Variations in the amount of volcanic material in Campanian and Maastrichtian sediments at DSDP sites on the Walvis Ridge suggest that volcanic activity had greatly decreased in the region by late Maastrichtian time [Simon and Schmincke, 1984]. As rates of seafloor spreading and volcanic activity decreased in the South Atlantic Basin, the rate of thermal subsidence should have gradually increased along the flanks of the Mid-Atlantic Ridge. A tectonic reconstruction of South Atlantic bathymetry reveals that by the mid Maastrichtian, passages that could accommodate north-south exchange of intermediate to deep waters had developed adjacent to the east and west flanks of the Rio Grande Rise (Figure 7)

[Cande et al., 1988]. Moreover, this reconstruction (Figure 7) indicates that such passages would have been extremely narrow at the base of Magnetochron 32 (late late Campanian) and virtually nonexistent during anomaly 33N (early late Campanian). The low temporal resolution of reversals in magnetopolarity makes it difficult to precisely pinpoint the time at which subsidence and seafloor spreading in the South Atlantic first allowed intermediate and deep water exchange across the barrier imposed by the Walvis Ridge and Rio Grande Rise. The timing implied by the reconstruction in Figure 7, however, is consistent with the mid Maastrichtian timing suggested by benthic foraminiferal δ^{13}C records as well as a sharp decrease in the Ce anomaly within lower Maastrichtian (69-68 Ma) pelagic carbonate sediments at site 516 in the South Atlantic, which has been interpreted to reflect increased dissolved oxygen levels due to the onset of deep water circulation over the Rio Grande Rise [Hu et al., 1988]. The removal of barriers to north-south flow in the Atlantic during the mid Maastrichtian had the potential to initiate a complex series of changes and corresponding feedbacks that affected thermohaline circulation, climate, and marine biota (Figure 9).

4.1. Thermohaline Circulation

On the basis of water mass aging gradients inferred using benthic foraminiferal δ^{13}C values [e.g., Kroopnick, 1985] and relatively positive benthic foraminiferal δ^{18}O values at tropical Pacific sites, Barrera et al. [1997] suggested that during the early Maastrichtian (time 1), bottom waters formed in the northern Pacific and reached sites in the tropical Pacific and Southern Ocean. Relatively low benthic foraminiferal values at sites 327 and 511 on the Falkland Plateau suggest that these waters reached as far as the Atlantic sector of the Southern Ocean (Figure 4); a negative excursion in the early Maastrichtian δ^{13}C record from *A. acuta* at site 525 [Li and Keller, 1998] suggests that these. waters may have also reached the southern flanks of the Walvis Ridge (Figure 6). Among sites that exhibit the negative excursion in benthic foraminiferal δ^{13}C values during time 1, relatively high δ^{13}C values at site 761 (Figure 4) have been interpreted to reflect the input and mixing of relatively warm and somewhat more saline bottom waters forming at low latitudes in the Indian Ocean [MacLeod and Huber, 1996]. Alternatively, benthic foraminiferal δ^{13}C values at site 761 may record mixing of Pacific deep waters with an Atlantic source through the Tethys [e.g., Barron and Peterson, 1989; Schmidt and Mysak, 1996].

The similarity among early Maastrichtian benthic foraminiferal δ^{13}C records from sites 357, 390, and 548 makes it difficult to discern the sources of intermediate and deep water for sites in the Atlantic north of the Walvis Ridge and Rio Grande Rise during time 1. Because benthic foraminifera at these sites have considerably higher δ^{18}O values (by ~2‰) relative to planktonic counterparts, Atlantic intermediate and deep water likely had a cool source. Possible shallow connections to the Arctic through the Labrador Sea and Baffin Bay could have provided deep waters during the Campanian and Maastrichtian [Umpleby, 1979; Grant, 1980; Gradstein and Srivastava, 1980], although re-
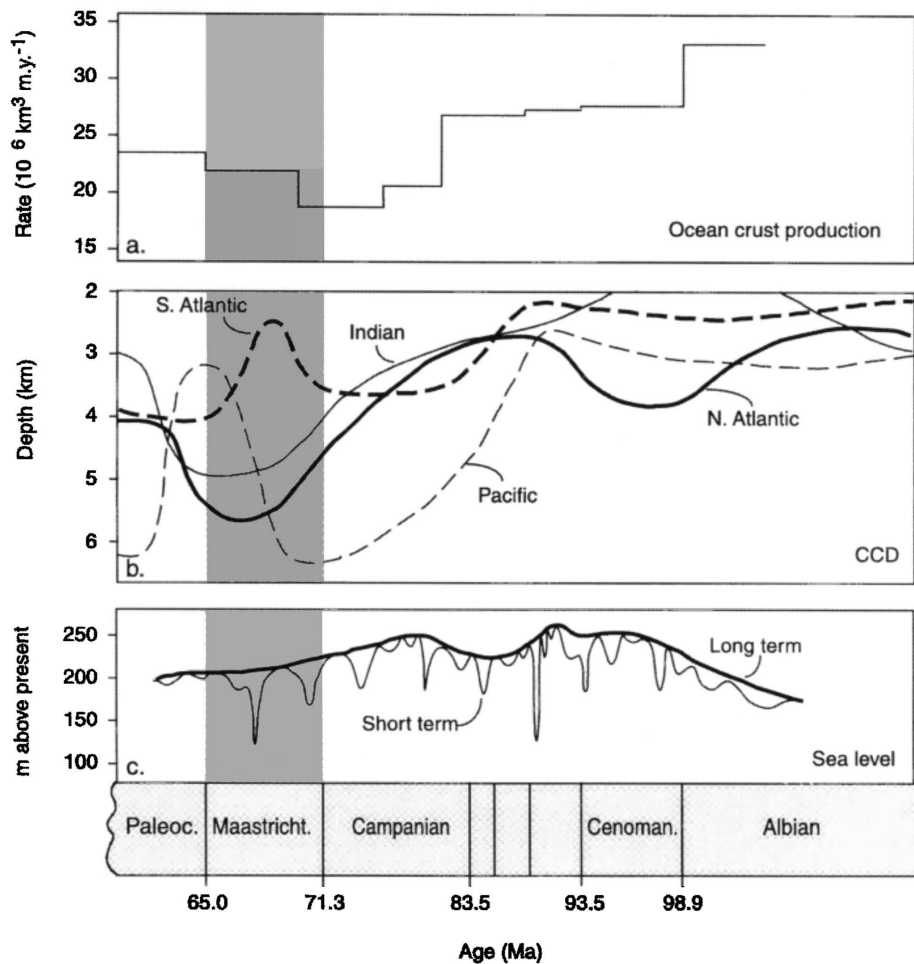
Figure 8. Upper Cretaceous variations in (a) ocean crust production rate [modified from Larson, 1991], (b) carbonate compensation depth (CCD) [modified from Arthur et al.. 1985], and (c) sea level [modified from Haq et al., 1987]. CCD estimates of Arthur et al. [1985] were determined using average core $CaCO_3$ concentrations, paleodepths calculated from thermal subsidence data, and biostratigraphic and magnetostratigraphic data.

gions of seafloor spreading in the North Atlantic were narrow and did not extend north of ~60°N at that time [Sclater et al., 1977; Srivastava and Tapscott, 1986]. Alternatively, bottom water production could have been a winter phenomenon at lower latitudes, driven by seasonal cooling and/or differences in evaporation-precipitation fluxes [e.g., Robinson et al., 1992].

Although the disappearance of the interocean benthic foraminiferal $\delta^{13}C$ gradient hinders any isotopically based inferences regarding the exact nature of the change in oceanic circulation patterns at the early/late Maastrichtian boundary (base of time 2), the resulting effects can be discerned in the gradual reestablishment of an interbasinal $\delta^{13}C$ gradient during time 3. During that time the most positive benthic foraminiferal $\delta^{13}C$ values are found at Atlantic sites 384, 390, and 548, suggesting that these sites were relatively proximal to areas of deep water production. A decrease in benthic foraminiferal $\delta^{13}C$ values to the south at sites on the Rio Grande Rise (site 357) and Walvis Ridge (sites 527 and 528) could reflect the aging of deep waters as they traveled from the North Atlantic. Relatively low benthic values at site 524 may reflect the isolation of the Cape Basin from the influence of deep water formed in the North Atlantic by the

Walvis Ridge, which appears to have separated the eastern South Atlantic into two distinct basins during the late Maastrichtian (Figure 7) [Sclater et al., 1977]. Alternatively, given the 3500 m paleodepth for site 524 (Figure 2), low benthic foraminiferal $\delta^{13}C$ values may be monitoring an older, deeper water mass relative to other sites. If so, these relationships suggest that the mid Maastrichtian circulation changes proposed here may have been restricted to intermediate-deep water depths (e.g., 1000–2500 m) and had little influence on bottom water circulation patterns. Stable isotope data suggest that intermediate and/or deep waters may also have been forming at high latitudes in the Southern Ocean during time 3 [e.g., Barrera, 1994]. Benthic foraminiferal $\delta^{13}C$ values at sites 689 and 690 are as high as those at sites 384, 390, and 548 in the North Atlantic (Figure 6). Moreover, planktonic and benthic foraminiferal oxygen isotope records at site 690 merge during time 3, indicating the absence of a surface to deep water $\delta^{18}O$ gradient (Figure 4) [Stott and Kennett, 1990]. Although records are less complete, the lowest benthic foraminiferal $\delta^{13}C$ values within time 3 tend to be from sites in the tropical Pacific, suggesting that as in the modern oceans, these areas were located furthest from sites of intermediate and bottom water production.
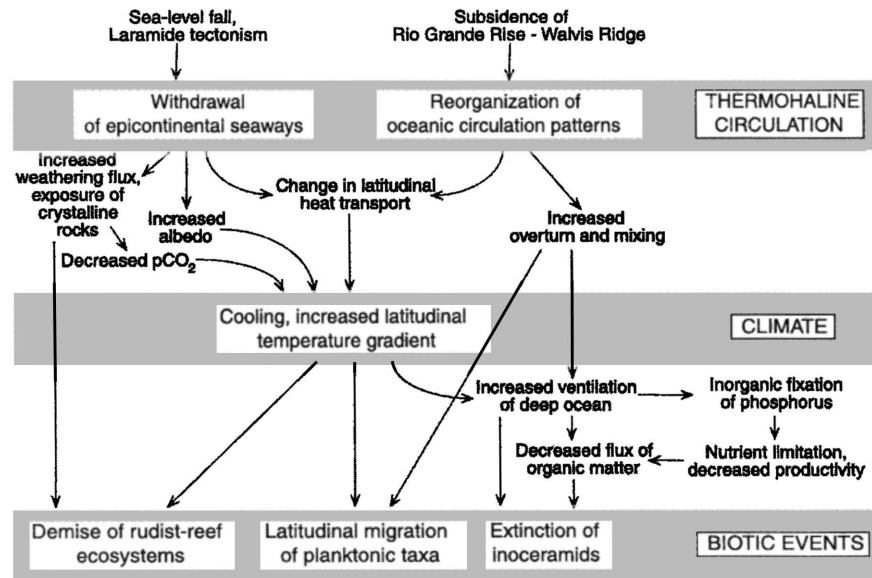
Figure 9. Flowchart illustrating the influence of Maastrichtian tectonic events on thermohaline circulation, climate, and biota.

Isotopic data suggest that the initiation of north-south flow of intermediate and deep water in the Atlantic during the mid Maastrichtian led to rapid shifts in the sites of deep water production and the direction of thermohaline circulation. Basin-specific, mid Maastrichtian fluctuations in the CCD [Thierstein, 1979; Arthur et al., 1985; Arthur and Dean, 1986], are consistent with this interpretation (Figure 8). These reconstructions define the CCD as the paleodepth at which pelagic sediments at DSDP sites contain <20% CaCO$_3$. Following a progressive deepening that had begun in the Coniacian, the CCD in the North Atlantic and Indian Oceans leveled out during the mid Maastrichtian and remained constant through the K/T boundary [Arthur and Dean, 1986]. The CCD in the South Atlantic Ocean initially rose sharply to ~2.5 km during the mid Maastrichtian and then deepened to ~4 km by the end of the Cretaceous [Arthur et al., 1985]. During this same interval the CCD in the Pacific shallowed by nearly 3 km from an initial depth of ~6 km [Thierstein, 1979]. The deepening of the CCD in the North Atlantic, the rise and fall of the CCD in the South Atlantic, and the abrupt shallowing of the CCD in the Pacific during the mid Maastrichtian could, in part, reflect a cessation of circulation driven by the flow of waters that formed in the northern Pacific (time 1) and the initiation of inter-ocean basin flow of intermediate and deep waters generated in the North Atlantic (time 2).

4.2. Climate

With the exception of a short-lived warming event during the last ~0.5 m.y. of the Maastrichtian, a range of oxygen isotope [e.g., Boersma and Shackleton, 1981; Boersma, 1984; Barrera et al., 1987; Barrera and Huber, 1990; Frakes et al., 1994], paleobiogeographic [Huber, 1992; Huber and Watkins, 1992], and paleobotanical indicators [Frederiksen, 1989; Spicer and Parrish, 1990] suggest that the Maastrichtian was characterized by overall cooling. Oxygen isotope records derived from near-surface dwelling planktonic for-aminifera from a latitudinal distribution of sites exhibit an increase in the rate of change toward higher δ^{18}O values beginning at the early/late Maastrichtian boundary (Figure 10; base of time 2). Records from sites 689, 690, and 750 in the Southern Ocean indicate greater cooling than do records from mid- to low-latitude sites (Figure 10), suggesting an increase in the latitudinal temperature gradient. Increased high-latitude cooling during the late Maastrichtian is also supported by oxygen isotope data from molluscan macrofossils from James Ross Island, which suggest that subpolar conditions were established during the late Maastrichtian [Ditchfield et al., 1994].

One exception to the pattern described above is the planktonic foraminiferal record from site 390, in which δ^{18}O values decrease abruptly from ~ –0.5 to –1.5‰ at the early/late Maastrichtian boundary (Figure 10). Planktonic δ^{18}O values below the boundary are similar to those from southern high latitude sites, whereas δ^{18}O values above the boundary match records derived from extra-Atlantic sites at similar paleolatitudes. Because this shift coincides with a mid Maastrichtian unconformity at site 390 [Benson et al., 1978], relatively positive δ^{18}O values from early Maastrichtian foraminifera may reflect diagenetic alteration in cool waters at the seafloor during the hiatus. Alternatively, the negative shift in planktonic δ^{18}O values at site 390 may reflect changes in surface circulation patterns or evaporation-precipitation balances associated the initiation of north-south deep water flow in the Atlantic and increased rates of deep water export. Increased warming of surface waters along the eastern margin of North America could reflect the development of a proto-Gulf Stream Current. This interpretation is supported by the retention of compositional differences between early Maastrichtian planktonic and benthic taxa at site 390 (Figure 5) and an inferred mid Maastrichtian warming in eastern North America and the southern western interior [Wolfe and Upchurch, 1987]. In addition, late Maastrichtian benthic foraminiferal faunas at abyssal site 384 in the North Atlantic suggest high nutrient
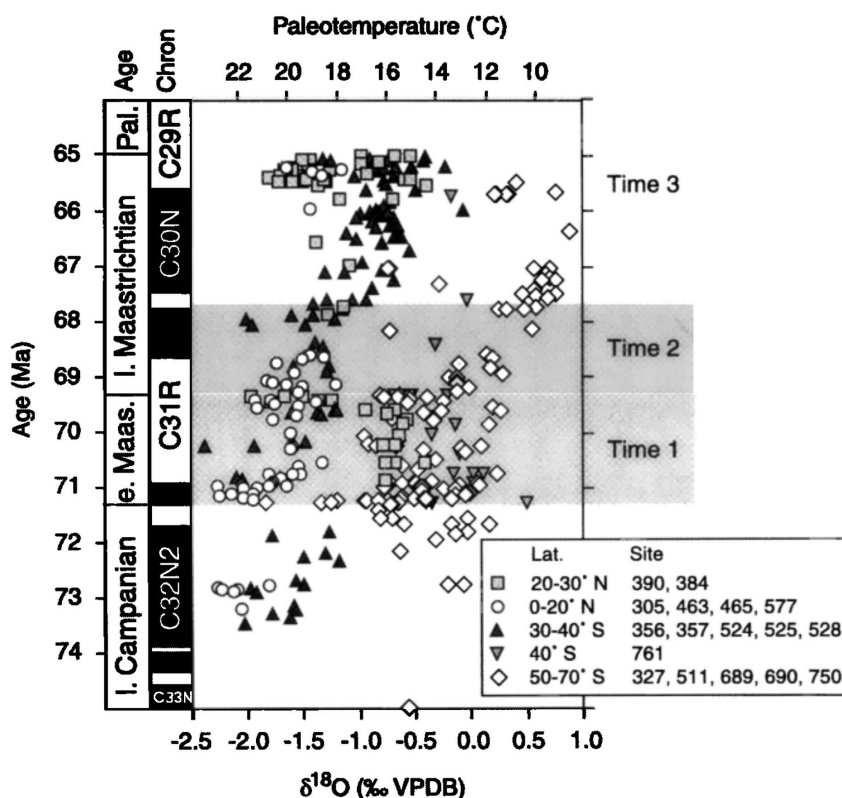
Figure 10. Planktonic foraminiferal oxygen isotope records generated in this study and compiled from literature sources (Table 1). Also shown are Maastrichtian sea surface temperature estimates calculated using the paleotemperature equation of Erez and Luz [1983]. Symbols reflect paleolatitude. Planktonic foraminifera are *Pseudoguembelina* (0–20° N), *Rugoglobigerina* (20–30° N, 40° S), *Pseudotextularia* and (20-30° N, 30–40° S), and *A. australis* (50–70° S). With the possible exception of *Pseudoguembelina*, foraminifera are inferred to be warm season, near-surface dwelling planktonic foraminifera [D'Hondt and Arthur, 1995].

availability and are consistent with increased productivity, possibly driven by the mixing of a warm surface current from the western North Atlantic with a cold surface current from the north [Widmark and Speijer, 1997].

Regardless of how data from site 390 are interpreted, oxygen isotope records derived from near-surface dwelling planktonic foraminifera from a range of sites suggest that the rate of global cooling accelerated at the early/late Maastrichtian boundary. Assuming a mean oceanic $\delta^{18}O$ value of –1‰ SMOW [Shackleton and Kennett, 1975] and ignoring any differential salinity effects between water masses [Railsback et al., 1989], estimates of sea surface temperatures (SSTs) using the oxygen isotope paleotemperature equation of Erez and Luz [1983] indicate that the latitudinal temperature gradient in the Southern Hemisphere between 50°–70° S and 0°–20° N increased from ~6–8°C during the early Maastrichtian (time 1) to as much as 10°–12°C during the late Maastrichtian (time 3). However, the late Maastrichtian latitudinal SST gradient still was substantially lower than that characteristic of the modern oceans. It has been suggested that low latitudinal temperature gradients were maintained throughout the Late Cretaceous by either elevated levels of meridional heat transport or less extreme latitudinal differences in the balance of radiation to and from the sea surface [D'Hondt and Arthur, 1996].

The onset of accelerated cooling at the early/late Maastrichtian boundary (base of time 2) suggests that the mid

Maastrichtian change in circulation patterns may have affected global climate by changing the efficiency of latitudinal heat transport (Figure 9). Southern Hemisphere cooling during the late Maastrichtian has been previously attributed to a change from the production of warm, saline bottom water in the tropics to deep water formation at high latitudes driven by a change in sea level [Barrera, 1994; MacLeod and Huber, 1996]. Foraminiferal carbon and oxygen isotope data (Figures 4 and 5), however, suggest that deep water was produced at relatively high latitudes throughout the Maastrichtian and that inferred circulation changes served to relocate the major sites of deep water formation from the northern Pacific to the high-latitude North Atlantic and Southern Ocean. As such, it is unlikely that the mid Maastrichtian circulation change was the sole contributor to late Maastrichtian cooling.

The mid Maastrichtian also saw the retreat of the major Cretaceous continental interior seaways in North America, Europe, Asia, South America, and Africa [Reyment and Dingle, 1987; Vakhrameev, 1987; Guillande et al., 1990; Elder and Kirkland, 1994; Roberts and Kirschbaum, 1995]. Their disappearance likely reflects falling eustatic sea level (Figure 8) [Haq et al., 1987] and, at least in the western interior foreland basin, a decrease in the rate of subsidence following the shift from Sevier-style to Laramide-style orogeny [Cross, 1986; Stott, 1993; Elder and Kirkland, 1994]. Circulation in north-south trending continental interior seaways may have

played a significant role in the maintenance of low latitudinal temperature gradients during much of the Cretaceous (Figure 9). Estuarine circulation in the Western Interior Seaway of North America, for example, could have enhanced heat transport to and from the high latitudes by drawing in warm Tethyan and cool Boreal waters while simultaneously exporting bottom waters along the limbs of a strong, counterclockwise gyre system that occupied the entire north-south extent of the seaway [Slingerland et al., 1996]. The effect of the dissipation of north-south seaways on the distribution of Earth surface temperatures was likely compounded by a concomitant increase in surface albedo (Figure 9), especially in the Northern Hemisphere, and, although difficult to quantify, changes in the balance between net precipitation and evaporation as the seaways withdrew. In addition, an increase in the rate of change toward higher $^{87}Sr/^{86}Sr$ ratios beginning at ~71 Ma (Figure 3) [Nelson et al., 1991; Barrera et al., 1997] suggests that the weathering flux changed as the area of exposed land mass increased and crystalline basement rocks were newly exhumed by block uplift. Weathering reactions, which consume CO_2 over the long term [Holland, 1984], may have also contributed to cooling by effecting a drawdown of atmospheric pCO_2 (Figure 9).

4.3. Biotic Events

4.3.1. Distribution of planktonic taxa. Changes in the diversity and latitudinal distribution of planktonic foraminifera and calcareous nanoplankton [Douglas, 1969; Sliter, 1972; Huber, 1992; Huber and Watkins, 1992] are consistent with the inferred mid Maastrichtian reorganization of oceanic circulation patterns. Total and keeled planktonic foraminiferal species diversity at high latitudes was low during the late Campanian and early Maastrichtian, leading to a high gradient in latitudinal diversity during this time [Douglas, 1969, Sliter, 1972; Huber, 1992]. During the early late Maastrichtian, several low-latitude keeled and nonkeeled planktonic foraminiferal taxa (e.g., *Globigerinelloides subcarinatus, Globotruncanella citae, Globotruncana bulloides,* and *Globotruncana subcircumnodifer*) began to migrate toward the high southern latitudes [Huber, 1990, 1991, 1992; Huber and Watkins, 1992]. These migrations resulted in a poleward shift of biogeographic realm boundaries and an increase in total- and keeled-species diversity at high southern latitude sites during the late Maastrichtian [Huber, 1992]. Beginning at the early/late Maastrichtian boundary, the planktonic foraminifer *Abathomphalus mayaroensis* and the calcareous nanoplankton *Nephrolithus frequens*, both upper Maastrichtian marker species that originated in southern high latitudes [Worsley, 1974; Wise, 1988], began a migration to low latitudes in the Atlantic [Pospichal and Wise, 1990; Huber and Watkins, 1992].

While temperature influences the distribution of Recent planktonic foraminifera [e.g., Bé, 1977], many other factors play a role, including the vertical structure of the water column, which can affect the depths of the chlorophyll maximum zone and the pycnocline [Fairbanks and Wiebe, 1980]. Oxygen isotope rankings of Maastrichtian planktonic foraminifera suggest that keeled globotruncanids, including those taxa that migrated poleward throughout the mid Maastrichtian, did not consistently occupy the deepest or coldest paleohabitats but, as a whole, inhabited niches spanning a wide thermal range [D'Hondt and Arthur, 1995]. As such, the concomitant equatorward migration of high-latitude taxa, the poleward migration of keeled planktonic foraminifera, and the expansion of biogeographic realm boundaries toward high southern latitudes during an interval marked by decreasing sea-surface temperatures [Douglas and Savin, 1975; Boersma and Shackleton, 1981; Barrera and Huber, 1990] suggest that the density structure and stability of surface waters may have played a larger role in the distribution of Maastrichtian planktonic taxa [Huber, 1992]. Possibly, as barriers to north-south flow in the Atlantic dissipated, the development of a global deep water circulation system resulted in increased surface-water communication, in effect leading to a global homogenization of surface water properties during the late Maastrichtian (Figure 9).

4.3.2. Extinction events. The mid Maastrichtian pulse of extinction, which occurred over a ~0.5 m.y. interval that spans the early/late Maastrichtian boundary [Kauffman and Hart, 1995], included the major lineages of Inoceramidae [Dhondt, 1983, 1992; MacLeod, 1994; MacLeod et al., 1996], the collapse of rudistid bivalve-dominated reef ecosystems including associated mollusk and coral taxa [Johnson and Kauffman, 1990, 1996; Swinburne, 1990], and a drastic decrease in the diversity of warm water benthic biofacies containing numerous bivalves (e.g., *Exogyra*), gastropods (e.g., *Neriniidae* and *Actaeonellidae*) and echinoids [Kauffman, 1988]. Biologic diversity remained low throughout the remainder of the Cretaceous [Barnes et al., 1995]. Although the scale of the mid Maastrichtian event, one that simultaneously influenced taxa across a wide range of water depths and latitudes, suggests that the driving mechanism must have been extraordinary, an often overlooked feature of the event is its relatively selective nature. With minor exceptions, the biota most affected were benthic detritus or filter feeders. Shallow water, nonbenthic biota appear to have been little influenced. Planktonic foraminifera and calcareous nanoplankton exhibit shifts in their latitudinal distributions but do not decrease in overall diversity or abundance [Huber, 1992; Huber and Watkins, 1992]. Likewise, with the exception of some specialized lineages [Barnes et al., 1995], ammonites show little change in abundance or diversity during the mid Maastrichtian [Wiedmann, 1988; Ward and Kennedy, 1993].

4.3.3. Inoceratnids. Despite a wide geographic and bathymetric range, Maastrichtian inoceramid bivalves are purported to have been adapted to a relatively narrow ecological niche associated with low concentrations of dissolved oxygen in bottom waters [MacLeod et al., 1996]. On the basis of rapid increases in the population of burrowing organisms and the abundance and diversity of benthic foraminifera across the mid Maastrichtian extinction interval, increased ventilation of the deep sea has been proposed to account for the extinction of deep-sea inoceramids [MacLeod, 1994]. Although previously linked to a shift from low- to high-latitude deep water formation at the early/late Maastrichtian boundary [MacLeod and Huber, 1996], accelerated high-latitude cooling, and increased rates of overturn and mixing associated with the mid Maastrichtian circulation changes proposed here would have also served to increase dissolved O_2 levels in the deep sea (Figure 9).

An independent line of evidence supporting higher dissolved O_2 levels during the late Maastrichtian lies in the record of marine phosphate deposition. The distribution of economic phosphate deposits indicates that the Campanian through early Eocene oceans were marked by widespread phosphorite accumulation, with peaks near the Santonian/Campanian and Paleocene/Eocene boundaries, respectively [Cook and McElhinny, 1979; Arthur and Jenkyns, 1981]. Between these maxima, phosphorite accumulation declined by an order of magnitude from the late early Maastrichtian to the late early Paleocene [Arthur and Jenkyns, 1981]. Because a primary source of phosphorus to the deep marine realm is organic matter, the amount available in the sediment column for phosphatization depends largely upon the amount of organic matter that survives the transit from surface waters to the seafloor. Progressive ventilation of the deep sea during the late Maastrichtian would have led to a decrease in this source of phosphate. Given that falling organic particulates are a primary nutrient source for deep-sea filter feeders, a decrease in the flux of organic matter may have also contributed to the demise of inoceramids (Figure 9). Overall, the main source of phosphorus to the oceans is riverine input, which may have increased during the late Maastrichtian as weathering rates increased. If ocean deep water conditions are oxidizing, however, large amounts of phosphorus may be effectively removed from the system by inorganic fixation as carbonate fluorapatite or by adsorption on Fe oxyhydroxides (Figure 9) [Berner, 1973; Arthur and Jenkyns, 1981]. During the mid Maastrichtian, Fe oxyhydroxides may have become more widespread on ridge crests as a result of increased deep water ventilation. Because phosphorus is a limiting nutrient, a decrease in surface water productivity would be expected as the amount of phosphorus available for the biomass decreased (Figure 9). In turn, decreased surface water productivity (or decreases in mean oceanic $[PO_4^{3-}]$) should result in a decreased surface to deep water $\delta^{13}C$ gradient [e.g., Bender, 1984]. A progressive decrease in surface to deep water $\delta^{13}C$ gradients beginning at the early/late Maastrichtian boundary is evident at all sites examined in the present study (Figures 4 and 5). Despite increased weathering rates during the late Maastrichtian that might have delivered somewhat more phosphorus to the oceans, foraminiferal carbon isotope records are consistent with an overall increase in the removal rate of phosphorus through inorganic fixation and a rapid increase in the concentration of dissolved O_2 in intermediate and deep waters during the late Maastrichtian.

4.3.4. Rudistid-reef ecosystems. Johnson et al. [1996] linked middle and Late Cretaceous collapses of rudist-dominated reefs to the enhanced export of "warm, salty" deep water from low-latitude shelves occupied by carbonate platforms [Johnson et al., 1996, Figure 1]. The high rates of heat export were suggested to have caused cooling and habitat elimination. The evidence presented here, however, indicates that high-latitude deep water formation prevailed throughout the Maastrichtian. Although our hypothesis is consistent with tropical cooling as a possible reason for platform demise, we do not believe enhanced export of heat from the tropics by warm saline bottom water production played a

role. We attribute the extinction of rudist-dominated ecosystems primarily to the rapid deterioration of warm tropical habitats and decreased calcification rates associated with the onset of accelerated global cooling at the early/late Maastrichtian boundary.

5. Conclusions

A synthesis of stable isotope data from deep-sea cores indicates that Maastrichtian oceanic evolution was characterized by three distinct phases. During the early Maastrichtian (time 1), stable isotope data indicate that characteristics of water masses differed on either side of the east-west trending Walvis Ridge-Rio Grande Rise complex in the South Atlantic. Deep water appears to have formed at high latitudes in both the North Atlantic and North Pacific. The disappearance of interbasin differences among water masses at the early/late Maastrichtian boundary suggests a reorganization of oceanic circulation patterns that resulted in a global homogenization in the character of intermediate and, possibly, deep waters. Plate kinematic data suggest that this event was triggered by the final breaching of the Walvis Ridge and Rio Grande Rise. The reorganization of circulation patterns caused increased oceanic overturn and ventilation of the deep ocean and may have led to decreases in the inventory of dissolved phosphate and the flux of particulate organic matter to the deep sea. The extinction of benthic inoceramid bivalves may have been one result of these oceanographic changes. Increased land area (albedo) and weathering rates associated with mid Maastrichtian tectonism and the withdrawal of epicontinental seaways contributed to global cooling and increased latitudinal thermal gradients during the late Maastrichtian (times 2 and 3). The subsequent deterioration of warm tropical habitats and a decrease in calcification rates as surface temperatures cooled probably contributed to the final disappearance of rudist-dominated ecosystems. Falling sea surface temperatures and increased weathering ultimately led to the transfer of alkalinity to the oceans, an overall deepening of the CCD, and enhanced preservation of pelagic carbonates during the late Maastrichtian.

References

Alvarez, W., M. A. Arthur, A. G. Fischer, W. Lowrie, G. Napoleone, I. Premoli-Silva, and W. M. Roggenthen, Upper Cretaceous-Paleocene magnetic stratigraphy at Gubbio, Italy, V. Type section for the Late Cretaceous-Paleocene geomagnetic reversal timescale, *Geol. Soc. Am. Bull.,* 88, 383-389, 1977.

Arthur, M. A., and W. E. Dean, Cretaceous paleoceanography of the western North Atlantic Ocean, in *The Geology of North America, Vol. M, The Western North Atlantic Region,* edited by P. R. Vogt, pp. 617-630, Geol.

Soc. Am., Boulder, CO, 1986

Arthur, M. A., and H. C. Jenkyns, Phosphorites and paleoceanography, *Ocean. Acta,* 4, 83-96, 1981.

Arthur, M. A., W. E. Dean, and S. O. Schlanger, Variations in the global carbon cycle during the Cretaceous related to climate, volcanism and changes in atmospheric CO$_2$, in *The Carbon Cycle and Atmospheric CO$_2$: Natural Variations Archean to Present,* Geophys. Monogr. Ser., vol. 32, edited by E. T. Sundquist and W. S Broecker, pp. 504-529, AGU, Washington, DC, 1985.

Barnes, C., A. Hallam, D. Kaljo, E. G. Kauffman, and O. H. Walliser, Global event stratigraphy, in *Global Events and Event Stratigraphy in the Phanerozoic,* edited by O. H. Walliser, pp. 319-333, Springer-Verlag, New York, 1995.

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

Barrera, E., and B. T. Huber, Evolution of Antarctic waters during the Maestrichtian: Foraminifer oxygen and carbon isotope ratios, ODP Leg 113, *Proc. Ocean Drill. Program, Sci. Results,* 113, 813-827, 1990

Barrera, E., B. Huber, S. Savin, and P. Webb, Antarctic marine temperatures: Late Campanian through early Paleocene, *Paleoceanography,* 2, 21-47, 1987.

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

Barron, E. J., and W. H. Peterson, Model simulation of the Cretaceous ocean circulation, *Science,* 244, 684-686, 1989.

Bé, A. W. H., An ecological, zoogeographic and taxonomic review of recent planktonic foraminifera, in *Oceanic Micropaleontology,* edited by A. T. S. Ramsay, pp. 1-100, Academic, San Diego, Calif., 1977.

Bender, M. L., On the relationship between ocean chemistry and atmospheric *p*CO$_2$ during the Cenozoic, in Climate Processes and Climate Sensitivity, Geophys. Monogr. Ser. vol. 29, edited by J. E. Hansen and T. Takahashi, pp. 352-359, AGU, Washington, D C., 1984.

Benson, W. E., et al, Sites 389 and 390 North rim of Blake Nose, *Initial Rep. Deep Sea Drill. Proj.,* 44, 69-151, 1978.

Berner, R A, Phosphate removal from sea water by adsorption on volcanogenic ferric oxides, *Earth Planet. Sci. Lett.,* 18, 77-86, 1973.

Boersma, A., Cretaceous and early Tertiary foraminifera from Deep Sea Drilling Project Leg 62 sites in the central Pacific, in *Initial Rep. Deep Sea Drill. Proj.,* 62, 377-396, 1981.

Boersma, A., Campanian through Paleocene paleotemperatures and carbon isotope sequence and the Cretaceous-Tertiary boundary in the Atlantic Ocean, in *Catastrophes and Earth History: The New Uniformitarianism,* edited by W. A. Berggren and J. A. Van Couvering, pp. 247-277, Princeton Univ. Press, Princeton, NJ, 1984.

Boersma, A., and N. J. Shackleton, Oxygen and carbon isotope variations and planktonic foraminifer depth habitats, late Cretaceous to Paleocene, Central Pacific, Deep Sea Drilling Sites 463 and 465, *Initial Rep. Deep Sea Drill. Proj.,* 62, 513-526, 1981.

Bralower, T. J., R. M. Leckie, W. V. Sliter, and H. R. Thierstein, An integrated Cretaceous microfossil biostratigraphy, in *Geochronology, timescales and global stratigraphic correlation,* SEPM Spec. Publ., 54, edited by W. A. Berggren et al., pp. 65-79, 1995.

Cande, S. C., J. L. LaBrecque, and W. F. Haxby, Plate kinematics of the South Atlantic: Chron C34 to present, *J. Geophys. Res.,* 93, 13,479-13,492, 1988.

Cepek, P., Mesozoic calcareous-nannoplankton stratigraphy of the central North Pacific (Mid-Pacific Mountains and Hess Rise), Deep Sea Drilling Project Leg 62, *Initial Rep. Deep Sea Drill. Proj.,* 62, 397-418, 1981.

Chave, A.D., Lower Paleocene-Upper Cretaceous magnetostratigraphy, Sites 525, 527, 528 and 529, Deep Sea Drilling Project Leg 74, *Initial Rep. Deep Sea Drill. Proj.,* 74, 525-532, 1984.

Cook, P. J., and M. W. McElhinny, A re-evaluation of the spatial and temporal distribution of sedimentary phosphate deposits in the light of plate tectonics, *Econ. Geol.,* 74, 315-330, 1979.

Craig, H., Isotopic standards for carbon and oxygen and correction factors for mass spectrometric analysis of carbon dioxide, *Geochim. Cosmochim. Acta,* 12, 133-149, 1957.

Cross, T. A., Tectonic controls of foreland basin subsidence and Laramide style deformation, western United States, in *Foreland basins,* IAS Spec. Publ., vol. 8, edited by P.A. Allen and P. Homewood, pp. 15-39, Blackwell Sci., Cambridge, MA, 1986.

Dhondt, A. V., Campanian and Maastrichtian Inoceramids A review, *Zitteliana,* 10, 689-701, 1983.

Dhondt, A. V., Cretaceous inoceramid biogeography A review, *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 92, 217-232, 1992.

D'Hondt, S., and M. A. Arthur, Interspecies variation in stable isotopic signals of Maastrichtian planktonic foraminifera, *Paleoceanography,* 10, 123-125, 1995.

D'Hondt, S., and M. A. Arthur, Late Cretaceous oceans and the cool tropic

paradox, *Science,* 271, 1838-1841, 1996.

D'Hondt, S., and M. Lindinger, A stable isotopic record of the Maastrichtian ocean-climate system: South Atlantic DSDP Site 528, *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 112, 363-378, 1994

Ditchfield, P., J. D. Marshall, and D. Pirrie, High latitude paleotemperature variation: New data from James Ross Island, Antarctica. *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 107, 79-101, 1994.

Douglas, R. G., Upper Cretaceous planktonic foraminifera in northern California, *Micropaleontology,* 15, 151-209, 1969.

Douglas, R. G., and S. M. Savin, Oxygen and carbon isotope analysis of Cretaceous and Tertiary foraminifera from the central North Pacific, *Initial Rep. Deep Sea Drill. Proj.,* 17, 591-605, 1973.

Douglas, R. G., and S. M. Savin, Oxygen and carbon isotope analyses of Tertiary and Cretaceous microfossils from the Shatsky Rise and other sites in the North Pacific Ocean, *Initial Rep. Deep Sea Drill. Proj.,* 31, 509-520, 1975.

Douglas, R. G., and S. M. Savin, Oxygen isotopic evidence for the depth stratification of Tertiary and Cretaceous foraminifera, *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 3, 175-196, 1978.

Elder, W. P., and J. I. Kirkland, Cretaceous paleogeography of the southern western interior region, in *Mesozoic Systems of the Rocky Mountain Region,* edited by M. V. Caputo, J. A. Peterson, and K. J. Franczyk, 415-440, SEPM Spec Publ., 1994.

Erez, J., and B. Luz, Experimental paleotemperature equation for planktonic foraminifera, *Geochim. Cosmochim. Acta,* 47, 1025-1031, 1983.

Fairbanks, R. G., and P. H. Wiebe, Foraminifera and chlorophyll maximum: Vertical distribution, seasonal succession and paleoceanographic significance, *Science,* 209, 1524-1525, 1980.

Frakes, L. A., J.-L. Probst, and W. Ludwig, Latitudinal distribution of paleotemperature on land and sea from early Cretaceous to middle Miocene, *C. R. Acad. Sci. Ser. 2,* 318, 1209-1218, 1994

Frederiksen, N. O., Changes in floral diversities, floral turnover rates, and climates in Campanian and Maastrichtian time, North Slope of Alaska, *Cretaceous Res.,* 10, 249-266, 1989.

Gradstein, F. M., and S. P. Srivastava, Aspects of Cenozoic stratigraphy and paleoceanography of the Labrador Sea and Baffin Bay, *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 30, 261-295, 1980.

Gradstein, F. M., F. P. Agterberg, J. G. Ogg, J. Hardenbol, P. van Been, J. Thierry, and Z. Huange, A Mesozoic timescale, *J. Geophys. Res.,* 99, 24,501-24,074, 1994.

Grant, A. C., Problems with plate tectonics; the Labrador Sea, *Bull. Can. Pet. Geol.,* 28, 252-278, 1980.

Guillande, R., D. Pons, and J. Bourgois, Decouverte d'une microflore continentale du Maestrichtien inferieur dans la Cordillere orientale de Colombie, consequences geodynamiques, *C. R. Acad. Sci., Ser. 2,* 310, 947-954, 1990.

Hamilton, N., Mesozoic magnetostratigraphy of Maud Rise, Antarctica, *Proc. Ocean Drill. Program Sci. Results,* 113, 255-260, 1990.

Haq, B. U., J. Hardenbol, and P. R. Vail, Chronology of fluctuating sea levels since the Triassic, *Science,* 235, 1156-1167, 1987.

He, Q., J. A. McKenzie, H. Oberhaensli, Stable-isotope and percentage-of-carbonate data for Upper Cretaceous/lower Tertiary sediments from Deep Sea Drilling Project Site 524, Cape Basin, South Atlantic, *Initial Rep. Deep Sea Drill. Proj.,* 73, 749-754, 1984.

Hess, J., M. L. Bender, and J. -G. Schilling, Evolution of the ratio of strontium-87 to strontium-86 in seawater from Cretaceous to Present, *Science,* 231, 979-984, 1986.

Holland, H. D., *The Chemical Evolution of the Atmosphere and Oceans,* 582 pp., Princeton Univ. Press, Princeton, NJ, 1984.

Hu, X., Y. L. Wang, and R. A. Schmitt, Geochemistry of sediments on the Rio Grande Rise and the redox evolution of the South Atlantic Ocean, *Geochim. Cosmochim. Acta,* 52, 201-207, 1988.

Huber, B. T., Maestrichtian planktonic foraminifer biostratigraphy of the Maud Rise (Weddell Sea, Antarctica): ODP Leg 113 Holes 689B and 690C, *Proc. Ocean Drill. Program Sci. Results,* 113, 489-514, 1990.

Huber, B. T., Planktonic foraminifer biostratigraphy of Campanian-Maestrichtian sediments from Sites 698 and 700, southern South Atlantic, *Proc. Ocean Drill. Program Sci. Results,* 114, 281-297, 1991.

Huber, B. T., Paleobiogeography of Campanian-Maastrichtian foraminifers in the southern high latitudes, *Palaeogeogr. Palaeoclimatol. Palaeoecol.,* 92, 325-360, 1992.

Huber, B. T., and D. K. Watkins, Biogeography of Campanian-Maastrichtian calcareous plankton in the region of the Southern Ocean: Paleogeographic and paleoclimatic implications, in *The Antarctic Paleoenvironment: A Perspective on Global Change,* Antarct. Res. Ser., vol. 56, edited by J. P. Kennett and D. A. Warnke, pp. 31-60, AGU, Washington, DC, 1992.

Huber, B. T., D. A. Hodell, and C. P. Hamilton, Middle-Late Cretaceous climate of the southern high latitudes, Stable isotopic evidence for minimal equator-to-pole thermal gradients, *Geol. Soc. Am. Bull.,* 107, 1164-1191, 1995.

Johnson, C. C., and E. G. Kauffman, Originations, radiations and extinctions of Cretaceous rudistid bivalve species in the Caribbean Province, in *Extinction Events in Earth History*, edited by E. G. Kauffman and O. H. Walliser, pp. 305-324, Springer-Verlag, New York, 1990.

Johnson, C. C., and E. G. Kauffman, Maastrichtian extinction patterns of Caribbean Province rudistids, in *Cretaceous-Tertiary Mass Extinctions: Biotic and Environmental Changes*, edited by N. MacLeod and G. Keller, pp. 231-274, W. W. Norton, New York, NY, 1996.

Johnson, C. C., E. J. Barron, E. G. Kauffman, M. A. Arthur, P. J. Fawcett, and M. K. Yasuda, Middle Cretaceous reef collapse linked to ocean heat transport, *Geology*, 24, 376-380, 1996.

Kauffman, E. G., The dynamics of marine stepwise mass extinction, in *Paleontology and Evolution: Extinction Events*, edited by M. A. Lamolda, E. G. Kauffman, and O. H. Walliser, *Rev. Esp. Paleontol.*, vol. Extraordinario. 57-71, 1988.

Kauffman, E. G, and M. B. Hart, Cretaceous bioevents, in *Global Events and Event Stratigraphy in the Phanerozoic*, edited by O. H. Walliser, pp 285-312, Springer-Verlag, New York, 1995

Kroopnick, P.M., The distribution of ^{13}C and ΣCO_2 in the world oceans, *Deep Sea Research, Part A*, 32, 57-84, 1985.

Larson, R. L., Latest pulse of Earth: Evidence for a mid-Cretaceous super-plume, *Geology*, 19, 547-550, 1991.

Li, L., and G. Keller, Maastrichtian climate, productivity and faunal turnovers in planktic foraminifera in South Atlantic DSDP sites 525A and 21, *Mar. Micropaleontol.*, 33, 55-86, 1998.

Loeblich, A. R., Jr., and H. Tappan, *Foraminiferal genera and their classification*, pp. 970 and 847 pl., Van Nostrand Reinhold, New York, NY, 1988.

Mackensen, A., H. Grobe, H.-W. Hubberten, and H. Kuhn, Benthic foraminiferal assemblages and the δ^{13}C-signal in the Atlantic sector of the Southern Ocean: Glacial to interglacial contrasts, in *Carbon Cycling in the Glacial Ocean: Constraints on the Ocean's Role in Global Change*, NATO ASI Ser., vol. 117, edited by R. Zahn et al., pp. 105-144, Springer-Verlag, New York, 1994.

MacLeod, K. G., Bioturbation, inoceramid extinction and mid Maastrichtian ecological change, *Geology*. 22, !139-142, 1994.

MacLeod, K. G., and B. T. Huber, Reorganization of deep ocean circulation accompanying a Late Cretaceous extinction event, *Nature*, 380, 422-425, 1996.

MacLeod, K. G., B. T. Huber, and P. D. Ward, The biostratigraphy and paleobiogeography of Maastrichtian inoceramids, in *The Cretaceous-Tertiary Event and Other Catastrophes in Earth History*, edited by G. Ryder, D. Fastovsky, and S. Gartner, Geol. Soc. Am. Spec. Pap., 307, 361-373, 1996.

MacLeod, K. G., B. T. Huber, and M. L. Ducharme, Paleontological and geochemical constraints on changes in the deep ocean during the Cretaceous greenhouse interval, in *Warm Climates in Earth History*, edited by B. T. Huber, K. G. MacLeod, and S. Wing, Cambridge Univ. Press, in press, 1998.

Manivit, H., Paleogene and Upper Cretaceous calcareous nannofossils from Deep Sea Drilling Project Leg 74, *Initial Rep. Deep Sea Drill. Proj.*, 74, 475-499, 1984.

Martin, E. E., and J. D. MacDougall, Seawater Sr isotopes at the Cretaceous/Tertiary boundary, *Earth Planet. Sci. Lett.*, 104, 106-181, 1991.

Monechi, S., and H. R. Thierstein, Late Cretaceous-Eocene nannofossil and magnetostratigraphic correlations near Gubbio, Italy, *Mar. Micropaleontol.*, 9, 419-440, 1985.

Nelson, B. K., G. K. MacLeod, and P. D. Ward, Rapid change in strontium isotopic composition of sea water before the Cretaceous/Tertiary boundary, *Nature*, 351, 644-647, 1991.

Patzkowsky, M. E., L. H. Smith, P. J. Markwick, C. J. Engberts, and E. D. Gyllenhaal, Application of the Fujita-Ziegler paleoclimate model: Early Permian and Late Cretaceous examples, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 86, 67-85, 1991.

Pospichal, J. J., and S. W. Wise, Jr., Maestrichtian calcareous nannofossil biostratigraphy of Maud Rise, ODP Leg 113 Sites 689 and 690, Weddell Sea, *Proc. Ocean Drill Program Sci. Results*, 113, 465-488, 1990.

Premoli-Silva, I., Upper Cretaceous-Paleocene magnetic stratigraphy at Gubbio, Italy; II, Biostratigraphy, *Geol. Soc. Am. Bull.*, 88, 371-374, 1977.

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

Reyment, R. A., and R. V. Dingle, Palaeogeography of Africa during the Cretaceous Period, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 59, 96-116, 1987.

Roberts, L. N. R., and M. A. Kirschbaum, Paleogeography of the Late Cretaceous of the Western Interior of Middle North America—Coal distribution and sediment accumulation, *U.S. Geol. Surv. Prof Pap.*, 1.561, 115, 1995.

Robinson, A. R., et al., General circulation of the eastern Mediterranean, *Earth Sci. Rev.*, 32, 285-309, 1992.

Schmidt, G. A., and L. A. Mysak, Can increased poleward oceanic heat flux explain the warm Cretaceous climate?, *Paleoceanography*, 11, 579-593, 1996

Schrag, D. P., D. J. DePaolo, and F. M. Richter, Oxygen isotope exchange in a two-layer model of oceanic crust, *Earth Planet. Sci. Lett.*, 111, 305-317, 1992.

Sclater, J. G., S. Hellinger, and C. Tapscott, The paleobathymetry of the Atlantic Ocean from the Jurassic to the present, *J. Geol.*, 85, 509-552, 1977.

Shackleton, N., and J. P. Kennett, Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon analyses in DSDP sites 277, 279 and 281, *Initial Rep. Deep Sea Drill. Proj.*, 29, 743-755, 1975.

Shackleton, N. J., M. A. Hall, and A. Boersma, Oxygen and carbon isotope data from Leg 74 foraminifers, *Initial Rep. Deep Sea Drill. Proj.*, 74, 599-612, 1984.

Simon, M., and H.-U. Schmincke, Late Cretaceous volcaniclastic rocks from the Walvis Ridge, southeast Atlantic, Leg 74, *Initial Rep. Deep Sea Drill. Proj.*, 74, 765-791, 1984.

Slingerland, R., L. R. Kump, M. A. Arthur, P. J. Fawcett, B. B. Sageman, and E. J. Barron, Estuarine circulation in the Turonian western interior seaway of North America, *Geol. Soc. Am. Bull.*, 108, 941-952, 1996,.

Sliter, W. V., Upper Cretaceous foraminiferal zoogeography and ecology—Eastern Pacific margin, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 12, 15-31, 1972.

Spicer, R. A., and J. T. Parrish, Late Cretaceous-early Tertiary palaeoclimates of northern high latitudes: A quantitative view, *J. Geol. Soc. London*, 147, 329-341, 1990.

Srivastava, S. P. and C. R. Tapscott, Plate kinematics of the North Atlantic, in *The Geology of North America, vol. M., The Western North Atlantic Region*, edited by P. R. Vogt and B. E. Tucholke, pp. 379-404, Geol. Soc. Am., Boulder, Colo., 1986.

Stott, D. F., Evolution of Cretaceous foredeeps: A comparative analysis along the length of the Canadian Rocky Mountains, in *Evolution of the Western Interior Basin*, edited by W. G. E. Caldwell and E. G. Kauffman, Geol. Assoc. Can. Spec. Pap., 39, 131-150, 1993.

Stott, L. D., and J. P. Kennett, The paleoceanographic and paleoclimatic signature of the Cretaceous/Paleogene boundary in the Antarctic: Stable isotopic results from ODP Leg 113, *Proc. Ocean Drill. Program Sci. Results*, 113, 839-848, 1990.

Sugarman, P. J., K. G. Miller, D. Bukry, and M. D. Feigenson, Uppermost Campanian-Maestrichtian strontium isotopic, biostratigraphic and sequence stratigraphic framework of the New Jersey Coastal Plain, *Geol. Soc. Am. Bull.*, 107, 19-37, 1995

Swinburne, N. H. M., The extinction of the rudist bivalves, Ph.D. thesis, 175 pp., Open University, Milton Keynes, Eng., UK, 1990.

Thiede, J., Subsidence of aseismic ridges; evidence from sediments on Rio Grande Rise (southwest Atlantic Ocean), *AAPG Bull.*, 61, 929-940, 1977

Thierstein, H R, Paleoceanographic implications of organic carbon and carbonate distribution in Mesozoic deepsea sediments, in *Deep Drilling Results in the Atlantic Ocean Continental Margins and Paleoenvironment*, Maurice Ewing Ser, v 3, edited by M. Talwani, W. Hay, and W. B. F. Ryan, pp. 249-274, AGU, Washington, D.C., 1979.

Townsend, H. A., The paleomagnetism of sediments acquired from the Goban Spur on Deep Sea Drilling Project Leg 80, *Initial Rep. Deep Sea Drill. Proj.*, 80, 389-414, 1985.

Umpleby, D. C., Geology of the Labrador shelf, *Pap. Geol. Surv. Can.*, 79-13, 34, 1979.

Vakhrameev, V. A., Cretaceous paleogeography of the USSR, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 59, 57-68, 1987.

Ward, P. D., and W. J. Kennedy, Maastrichtian Ammonites from the Biscay region (France, Spain), *Paleo. Soc. Mem.*, 34, 58, 1993.

Widmark, J. G. V., and R. P. Speijer, Benthic foraminiferal faunas and trophic regimes at the terminal Cretaceous Tethyan seafloor, *Palaios*, 12, 354-371, 1997.

Wiedmann, J., The Basque coastal sections of the K/T boundary: A key to understanding "mass extinction" in the fossil record, *Rev. Esp. Paleo.*, vol. Extraordinario, 127-140, 1988.

Wise, S. W., Jr., Mesozoic-Cenozoic history of calcareous nannofossils in the region of the Southern Ocean, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 67, 157-179, 1988.

Wolfe, J. A., and G. R. Upchurch Jr., North American nonmarine climates and vegetation during the Late Cretaceous, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 61, 33-77, 1987.

Worsley, T. R., The Cretaceous/Tertiary boundary event in the ocean, in *Studies in Paleoceanography*, edited by W. W. Hay, SEPM Spec. Publ., 20, 94-125, 1974.

Zachos, J., Aspects of Late Cretaceous and Paleogene oceanic climate and productivity, Ph.D., thesis, 565 pp., Grad. School of Oceanogr., University of Rhode Island, Kingston, RI, 1988