

Reorganization of deep ocean circulation accompanying a Late Cretaceous extinction event

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DEEP ocean circulation may be a significant factor in determining global climate¹⁻⁵. Increases in the flux of warm, saline waters from low to high latitudes would enhance the poleward transport of heat and, thus, help maintain the warm conditions at high latitudes typical of globally warm 'greenhouse' periods. But controversy exists^{1,2} as to whether the ocean's thermohaline circulation can transport enough heat to bring about the temperature distributions of these times, such as the mid-Cretaceous and early Eocene. Here we present stable-isotope records of ocean temperature and salinity that indicate that bottom waters in Late Cretaceous oceans of the Southern Hemisphere became cooler and less saline at the same time (about 70 Myr ago) as widespread biotic changes^{3,6-8}. These findings support the idea that changes in deep ocean circulation can act as a climate switch.

The Maastrichtian Age (the last 6 Myr of the Cretaceous Period) was a time of global environmental and biotic change. Isotopic data⁹⁻¹¹, microfossil palaeobiogeography¹², and palaeobotanical evidence^{13,14} all indicate cooling at this time. Several studies have shown warming during the last ~0.5 Myr of the Maastrichtian^{12,15,16} but superimposed on, not contradicting, evidence for long-term cooling. The disappearance of epicontinental seas as well as sequence stratigraphy¹⁷ document a large-scale, mid-Maastrichtian regression. Mid-Maastrichtian bioevents include floral shifts in the American west⁶, changes among benthic foraminiferal assemblages³, and extinction among both tropical, reef-forming rudistid⁷ and widely distributed, deep-sea inoceramid⁸ bivalves.

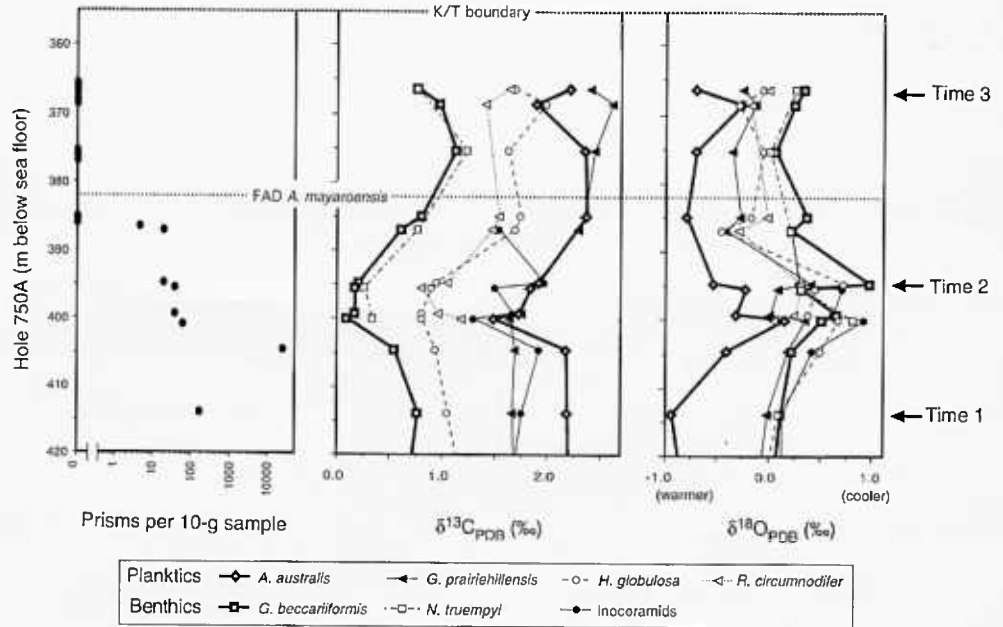
Oceanic processes may provide a causal mechanism uniting these events⁸. Global cooling and draining of tropical and subtropical epicontinental seas are expected to have reduced the production of hypersaline waters. Extinction among tropical, reef-

forming rudists has been attributed to the disappearance of hypersaline surface waters⁷; habitat restriction related to decreased influence of saline bottom waters has been invoked to explain extinction among deep-sea inoceramids⁸. Because it would have decreased oceanic heat transport, declining importance of warm, saline water masses could have affected Maastrichtian climates and indirectly led to observed changes among terrestrial floras. But direct data supporting the hypothesized palaeoceanographic changes are lacking.

To constrain better the physical properties of Maastrichtian oceans, we measured $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of size-sorted, single-taxon separates of planktic foraminifers, benthic foraminifers, and inoceramid remains from an austral (Hole 750A, ~60°S palaeolatitude) and a subtropical (Hole 761, ~35°S palaeolatitude) Ocean Drilling Program site. Preservation is uniformly good to excellent throughout the intervals studied, and local evidence of biological change (the disappearance of inoceramids) is well documented at both localities⁸. Because the inoceramid extinction is recognized globally⁸, isotopic shifts consistently associated with this event can be justifiably proposed as global phenomena. Similar-sized foraminifers were selected for each taxon analysed minimizing the possibility of size-related artefacts compromising the results¹⁸. Measured isotopic values are not significantly correlated with small variations in size that existed among samples of any taxon.

At both sites there is a ~0.7‰ negative $\delta^{13}\text{C}$ excursion in benthic foraminifers associated with the extinction of inoceramids (Figs 1 and 2). The abundance of inoceramid shell fragments (prisms) begins to decline at the initiation of this excursion, and inoceramid disappearance coincides with the point where $\delta^{13}\text{C}$ values return to original levels. Over the same stratigraphic interval there is a 0.5–1.0‰ positive shift in $\delta^{18}\text{O}$ values. Change at the subtropical site is more or less unidirectional; at the austral site, values first become heavier and then largely recover. Results from inoceramid prisms parallel the trends shown by benthic foraminifers, but exhibit consistent offsets perhaps related to chemosymbiosis in inoceramids or vital effects in benthic foraminifers¹⁹⁻²¹. Incorporating planktic data in the analysis shows that correlative changes in surface waters occurred only at the austral site. Isotopic trends in austral planktic and benthic foraminifers are parallel (Fig. 1), whereas there is no obvious correlation between planktic and benthic results in either $\delta^{13}\text{C}$ or $\delta^{18}\text{O}$ in the subtropical samples (Fig. 2).

FIG. 1 Inoceramid abundance and stable-isotope analyses $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ through the Maastrichtian in ODP Hole 750A ($\sim 60^\circ\text{S}$ palaeolatitude). Data for a benthic taxon and a shallow planktic taxon (based on average oxygen values) are highlighted. All taxa exhibit similar trends with a negative $\delta^{13}\text{C}$ and a positive $\delta^{18}\text{O}$ excursion associated with the decline of inoceramids. The seemingly anomalous inoceramid point on the $\delta^{13}\text{C}$ plot (the highest inoceramid point) could not be re-analysed because all available prisms were used in the first analysis and the next-higher sample did not contain a sufficient mass of prisms for an analysis. The dotted line labelled "FAD *A. mayaroensis*" is the first appearance datum of *Abathomphalus mayaroensis*, a planktic foraminifer commonly used as a zonal indicator for the upper Maastrichtian. Time slices 1, 2, and 3 are modelled in Fig. 3. Complementary data from this hole are presented in ref. 22. ($\delta^{13}\text{C} = [(^{13}\text{C}/^{12}\text{C})_{\text{sample}} / (^{13}\text{C}/^{12}\text{C})_{\text{standard}}] - 1$ and $\delta^{18}\text{O} = [(^{18}\text{O}/^{16}\text{O})_{\text{sample}} / (^{18}\text{O}/^{16}\text{O})_{\text{standard}}] - 1$; the PDB standard is used in both cases).



Carbon isotope data demonstrate mid-Maastrichtian palaeoceanographic change associated with inoceramid extinction but fail to yield a unique interpretation regarding the nature of that change. Further, competing hypotheses seem to make restrictive predictions. An explanation invoking a single cause for observed trends in $\delta^{13}\text{C}$ suggests a simple physical oceanographic relationship between the two sites. On the other hand, invoking separate benthic and planktic processes to explain the results requires that independent causes yielded $\delta^{13}\text{C}$ excursions of similar magnitude and direction. The lack of planktic $\delta^{13}\text{C}$ trends in subtropical samples (Fig. 2) argues against Maastrichtian redistribution of carbon among global reservoirs²²; therefore, planktic shifts at the austral site (Fig. 1) probably represent local changes in produc-

tivity. Sinking of surface-water carbon could explain parallel trends in co-occurring benthic taxa at this site. The $\delta^{13}\text{C}$ record of subtropical benthics, though, is clearly not tied to local surface values and requires either horizontal import of isotopic changes or temporal variation in benthic cycling of carbon. If the subtropical benthic $\delta^{13}\text{C}$ excursion reflects characteristics inherited from the source region(s) for bottom water, similarities between the two sites imply that they shared a single bottom-water mass (at any given time) with isotopic continuity over significant distances—differences in benthic $\delta^{13}\text{C}$ values are too small to infer transport direction^{23,24}. Alternatively, the subtropical (and austral) benthic record might indicate that changes in bottom waters (for example, temporal increase in the concentration of dissolved oxygen) led to

FIG. 2 ODP Hole 761 ($\sim 35^\circ\text{S}$ palaeolatitude). Benthic taxa exhibit the same isotope patterns associated with the inoceramid extinction observed in Hole 750A (Fig. 1), but the planktic taxa show little change across the interval studied. See text for further discussion. Asterisk indicates data from ref. 22.

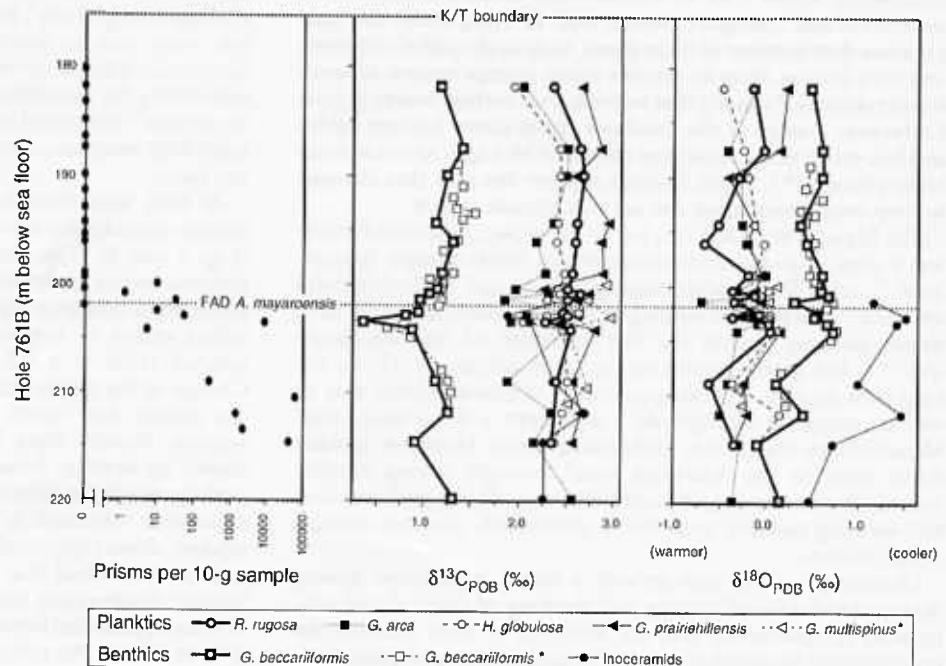
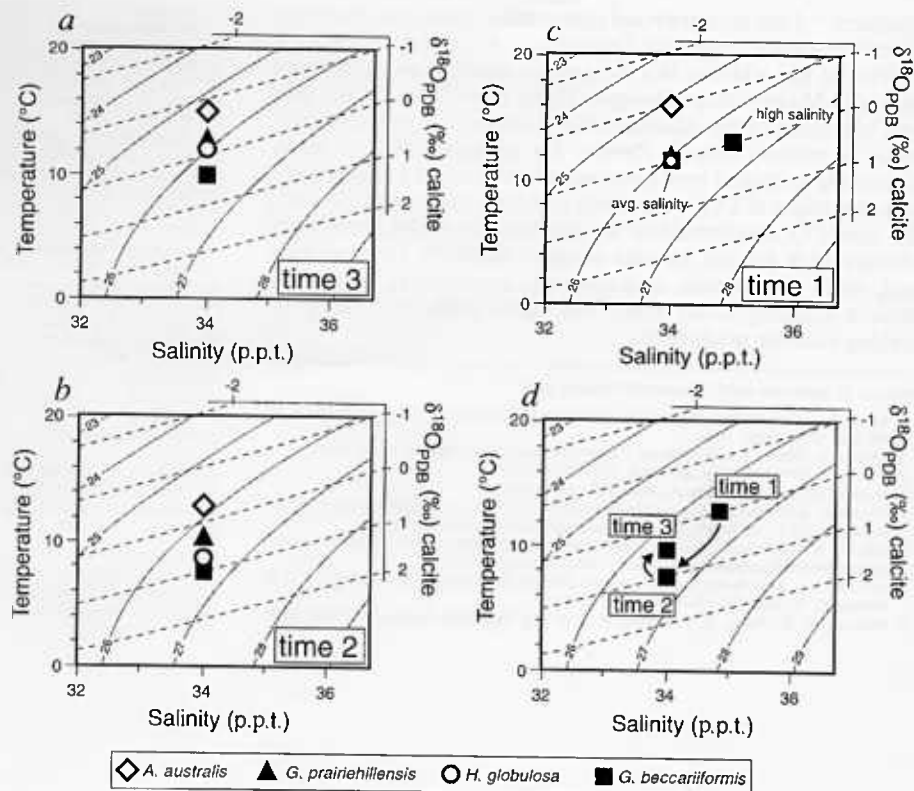


FIG. 3 $\delta^{18}\text{O}$ values of selected taxa from time slices 1, 2, and 3 (as indicated on Fig. 1) plotted in the temperature–salinity space appropriate for the Cretaceous period (after ref. 26). Sloping dashed lines represent $\delta^{18}\text{O}$ values of calcite in equilibrium with sea water of a given temperature and salinity. Curved solid lines are isopycnals expressed as σ_t (that is, $1,000(\rho_{\text{seawater}} - 1)$, where ρ is density in g cm^{-3}). a, Plot of time slice 3 (post-extinction) from Fig. 1, assuming an isohaline water column (34 p.p.t. is used as an appropriate Cretaceous value after ref. 26) showing apparent decreasing temperature and increasing density with depth. b, Plot of time slice 2 (inoceramids declining) suggesting cooler temperatures than time slice 3 but a stable, isohaline water column. c, Plot of time slice 1 (pre-extinction) suggesting a metastable water column if the water column were isohaline (*Gavellinella beccariiiformis* symbol labelled “avg. salinity”) and a stable water column if time slice 1 bottom water had elevated salinity and temperatures (*G. beccariiiformis* symbol labelled “high salinity”). d, Proposed changes in the relative temperature and salinity of bottom waters based on the analyses of *G. beccariiiformis* from time slices 1, 2 and 3. See text for further discussion.



reminalization of previously deposited, isotopically light organic carbon perturbing the vertical gradient in $\delta^{13}\text{C}$. This interpretation is less restrictive in terms of physical oceanography, but it invokes two causes for one pattern. The similarities in shape, but not timing, between the $\delta^{13}\text{C}$ curve for austral plankties and one or both of the benthic records would be coincidental.

Oxygen isotope results provide better constraints on the nature of oceanographic change. Although typically used to estimate palaeotemperature, $\delta^{18}\text{O}$ analyses can also indicate changes in salinity. Evaporation preferentially removes water containing ^{16}O , so saline waters generated by evaporative concentration are enriched in ^{18}O . Taxa living in high salinity, ^{18}O -enriched waters secrete an isotopically heavier test than those living in waters of lower salinity but the same temperature. If both the isotopic composition of the oceans and the isotopic enrichment related to evaporation and freshwater input are known, the $\delta^{18}\text{O}$ values of calcite in equilibrium with these waters can be plotted in temperature/salinity space^{25,26}. Further, because water density varies as a function of temperature and salinity, the plots can be contoured for density (isopycnals). By estimating the unknown variables, Woo *et al.* (ref. 26) constructed such a plot appropriate for Late Cretaceous oceans. Errors in estimated parameters would shift the position of the equilibrium lines, but would not greatly affect inferred differences in temperature/salinity among samples.

The $\delta^{18}\text{O}$ analyses suggest that bottom-water source regions switched from low to high latitudes coincident with the inoceramid extinction at both sites. At the austral site, if the water column was isohaline, $\delta^{18}\text{O}$ results for time 2 and time 3 (see Fig. 1) indicate a stable water column in which temperature decreases and density increases with depth (Fig. 3). However, at time 1, if the water column was isohaline, temperature would not change below the surface layer and the water column would be only metastable; further, some analyses between times 1 and 2 indicate density inversion. This paradox disappears if the pre-extinction bottom waters had higher salinity and temperature. Assuming bottom-water density at time 3 was equal to bottom-water density at time 1 (that is, cool, post-extinction bottom waters were as dense as the warm, salty bottom waters they are proposed to have replaced),

the $\delta^{18}\text{O}$ value of *Gavellinella beccariiiformis* at time 1 suggests a stable water column in which pre-extinction bottom waters were ~ 1 part per thousand (p.p.t.) saltier and $\sim 1.5^\circ\text{C}$ warmer than deep surface waters. The subtropical site gives similar results, but lacks the evidence for a peak in cooling analogous to time 2 at the austral site. Thus, initiation or intensification of intermediate and/or deep-water formation in the austral region may coincide with the cooling at time 2. These new, cool, dense waters then progressively displaced warm, saline waters at depth without demonstrably affecting mid-latitude surface water.

The salinity changes proposed above are conservative. If the density of bottom water at time 1 is estimated from results for time 2 rather than time 3, modelled pre-extinction bottom water is an additional ~ 0.5 p.p.t. saltier and $\sim 1^\circ\text{C}$ warmer. The assumption of density balance across time, though, is not required. For example, if the production of hypersaline waters declined or ceased in low-latitude seas, the extent of deep-sea regions bathed in warm, saline waters would have contracted towards the Equator without high-latitude forcing (implicitly assumed above). Thus, the upper limit of the salinity/temperature of pre-extinction bottom waters is bounded only by reasonable estimates of the ecological tolerances of contemporary benthic organisms. In addition, the possibility of a contribution from low-latitude forcing in reorganization of deep-ocean circulation means observed austral temperature changes may be more an effect than a cause of Maastrichtian change.

Uncertainty regarding cause and effect notwithstanding, we think a change in the location of the dominant source regions of bottom waters could have provided a causal mechanism that temporally focused Maastrichtian biotic changes. Whereas Late Cretaceous cooling occurred over an extended interval^{9,11,27}, palaeontological and geochemical shifts are concentrated in the mid-Maastrichtian^{7,8,22,28}. If the palaeoceanographic model presented above is accurate (that is, if a reversal in deep ocean circulation is coincident with global palaeoecological change) change in deep ocean circulation may have acted as a climate switch. A shift from low-latitude to high-latitude production of bottom waters is expected to have caused a decline in oceanic heat transport and lead to an increase in latitudinal temperature

gradients^{1,2}. Late Maastrichtian temperature gradients were steep relative to earlier in the Late Cretaceous¹, but additional data are needed to test whether this increase in gradient was correlated with mid-Maastrichtian changes. These constraints on the role and behaviour of the Maastrichtian deep ocean may have relevance to modern climate change. The reorganization of ocean circulation proposed here occurred when estimated atmospheric concentrations of CO₂ were about twice the present-day levels²⁹; the same CO₂ concentrations are employed in model simulations attempting to predict the next century's climate³⁰. Documenting and, more importantly, understanding controls on changes in ancient climates under these conditions could be crucial to making accurate predictions. □

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