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Notes

Evolution of middle to Late Cretaceous oceans—A 55 m.y. record of Earth's temperature and carbon cycle

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ABSTRACT

A new 55 m.y. global compilation of benthic foraminifera $\delta^{13}C$ and $\delta^{18}O$ for the middle to Late Cretaceous shows that there was widespread formation of bottom waters with temperatures >20 °C during the Cretaceous greenhouse world. These bottom waters filled the silled North Atlantic and probably originated as thermocline or intermediate waters in the tropical oceans. Carbon burial during the Cretaceous oceanic anoxic events produced a positive $\delta^{13}C$ shift in global carbon reservoirs, but this is not particularly large, especially by comparison with the remarkable Late Paleocene carbon maximum. The interbasin $\delta^{13}C$ gradient was unusually large during the Cretaceous hot greenhouse, probably because the North Atlantic sills prevented the free exchange of waters in the deep basin. The hot greenhouse ended when the Equatorial Atlantic Gateway opened sufficiently to flood the deep North Atlantic with relatively cool polar waters formed in the Southern Ocean.

INTRODUCTION

Recent studies have shown that Earth's climate and oceans underwent significant changes during the Cretaceous Period. These changes include the increase of global temperatures, peaking in the warmest ever reported temperatures during the early Late Cretaceous hot greenhouse (surface waters >35 °C and deep-ocean water >20 °C; Huber et al., 2002; Norris et al., 2002; Forster et al., 2007; Friedrich et al., 2008). There is also evidence for changing ocean circulation related to gateway opening (Jones et al., 1995; Frank and Arthur, 1999), high pCO_2 (Bice and Norris, 2002; Bice et al., 2006), repeated episodes of burial of organic matter during oceanic anoxic events (OAEs; Arthur et al., 1988; Schlanger and Jenkyns, 1976; Wilson and Norris, 2001), and probably short-lived glaciation events (Miller et al., 1999; Bornemann et al., 2008). While the dynamics of short-term events like OAEs are well studied, the broad evolution of Cretaceous climate and oceans is poorly resolved due to low-resolution sampling and large time gaps (Huber et al., 2002; Cramer et al., 2009) or diagenetically altered samples (Clarke and Jenkyns, 1999). We have produced new stable isotope data sets from benthic foraminifera and compiled them with literature data (Fig. 1) to construct a global deep-sea isotope record (see the GSA Data Repository[1]).

METHODS

Our new stable isotope data were generated from an ~100-m-thick, C_{org}-rich, laminated marlstone sequence of Cenomanian to Santonian age containing exceptionally well preserved (glassy) foraminiferal tests (e.g., Norris et al., 2002; Bornemann et al., 2008; Friedrich et al., 2008) from the western equatorial Atlantic at Demerara Rise (Ocean Drilling Program Leg 207; Fig. DR1 and Table DR1 in the Data Repository). We also analyzed samples from Turonian to Campanian sediments from the tropical Pacific Ocean (Deep Sea Drilling Project Sites 305 and 463; Fig. DR1, Tables DR2 and DR3, respectively). Samples were picked for monospecific samples of benthic foraminifera (10–100 individuals) and run on a Fisons Prism III isotope ratio mass spectrometer at the University of California–Santa Cruz (Sites 305, 463, 1259) and on a Finnigan MAT 251 mass spectrometer at the Leibniz-Labor (Kiel, Germany; all other sites). The analytical precision was <0.05‰ and <0.07‰ for $\delta^{18}O$ and $\delta^{13}C$, respectively. Paleotemperatures are estimated by applying the equation of Bemis et al. (1998), assuming a $\delta^{18}O_{seawater}$ of −1.0 ‰ (Shackleton and Kennett, 1975) and isotopic equilibrium of foraminifera with seawater.

RESULTS

The new Cretaceous (115–65 Ma; Fig. 2) benthic oxygen isotope ($\delta^{18}O_b$) compilation is characterized by a total $\delta^{18}O_b$ range of ~5.5‰. Before 97 Ma, the data show a pronounced trend that culminates in the lowest $\delta^{18}O_b$ values between 97 and 90 Ma. A >2‰ increase in $\delta^{18}O_b$ occurred during the next 12 m.y., followed by relatively stable $\delta^{18}O_b$ values up to 71 Ma. The last 6 m.y. of the Cretaceous are characterized by short-term oscillations in $\delta^{18}O_b$ superimposed by a >1‰ positive $\delta^{18}O_b$ excursion around the Campanian-Maastrichtian boundary (Fig. 2; Fig. DR2).

The benthic carbon isotope ($\delta^{13}C_b$) compilation shows a total range of ~6‰ (Fig. 2; Fig. DR2). Before 97 Ma, $\delta^{13}C_b$ values range between 0‰ and 2‰ and include substantial interocean gradients. Between 97 and 90 Ma, $\delta^{13}C_b$ values increase by ~1‰, followed by a long-lasting decrease of >2‰ to 70 Ma. Tropical Atlantic data (Ocean Drilling Program Leg 207) are 2‰–3‰ lower than contemporaneous Pacific Ocean and Southern Ocean data. After 70 Ma, $\delta^{13}C_b$ values increase globally by >1‰ (Fig. 2; Fig. DR2).

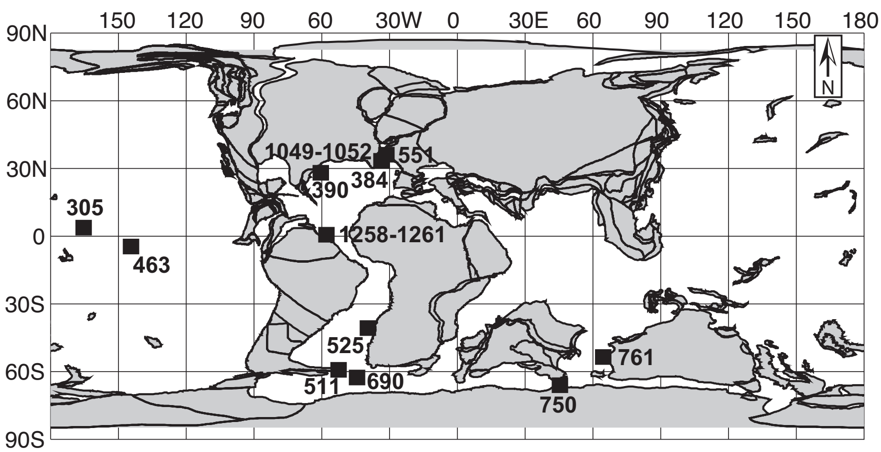

Figure 1. Paleogeographic reconstruction for 95 Ma (after Hay et al., 1999) showing sites used in this study.

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[1]GSA Data Repository item 2012043, materials, data source and age models, references, and Table DR1 (location of sites), Tables DR2–DR4 (new isotope data), and Figures DR1–DR3 (new data, isotope trends, and compilation with Zachos et al., 2008), is available online at www.geosociety.org/pubs/ft2012.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

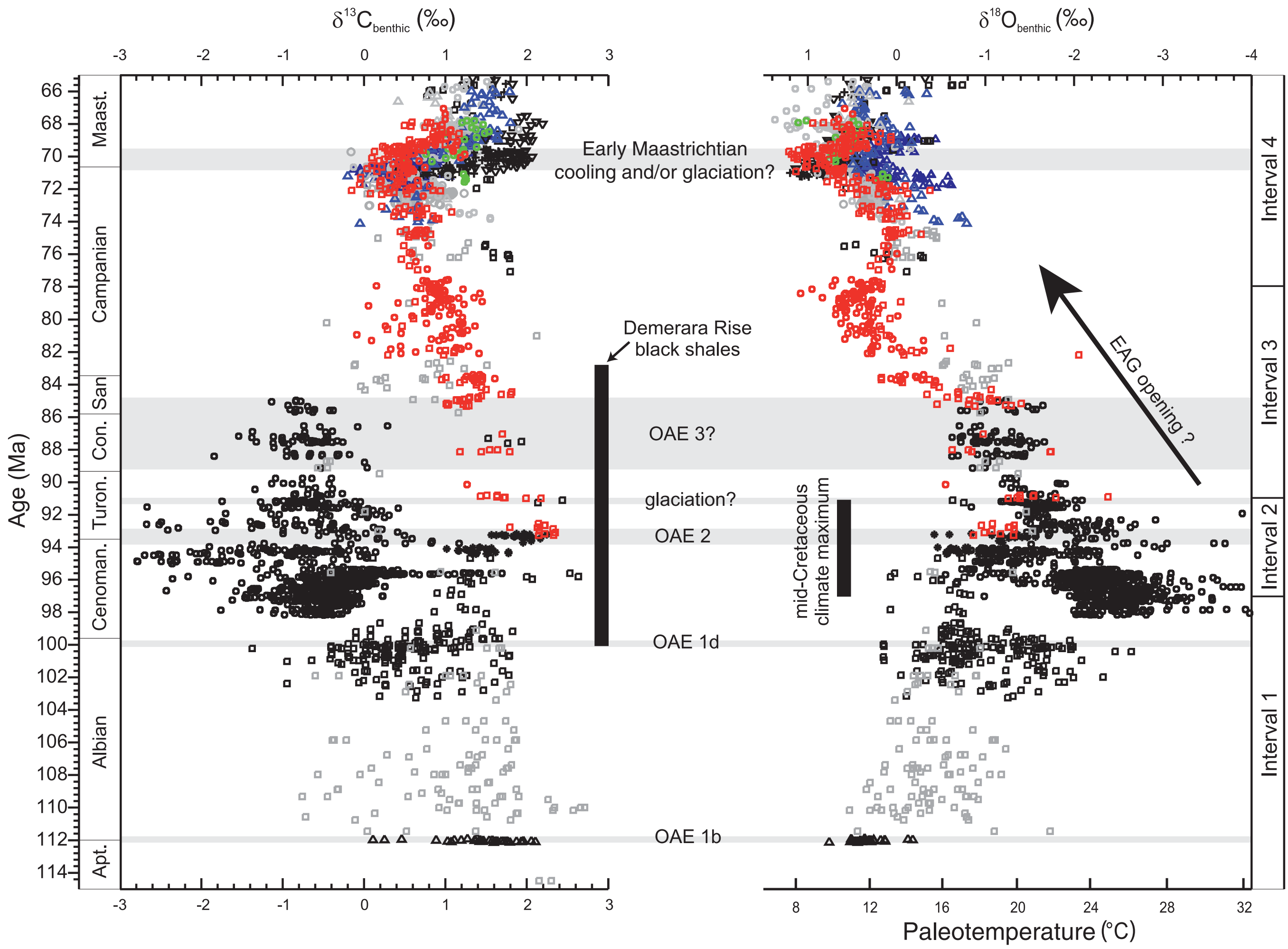

Figure 2. Stable oxygen and carbon isotope compilation of benthic foraminifera for 115–65 Ma. Black—North Atlantic Ocean, gray—southern high latitudes, red—Pacific Ocean, blue—subtropical South Atlantic Ocean, green—Indian Ocean. OAE—oceanic anoxic event; EAG—Equatorial Atlantic Gateway. Data sources are in the GSA Data Repository (see footnote 1).

DISCUSSION

New Benthic Isotope Compilation for the Cretaceous

We interpret $\delta^{18}O_b$ to be largely a function of variations in temperatures on the assumption that, with possibly a few exceptions (e.g., Bornemann et al., 2008; Miller et al., 1999), the middle to Late Cretaceous was ice free. There has been a long-standing hypothesis that variations in ocean salinity may also have contributed significantly to the $\delta^{18}O_b$ record (e.g., Brass et al., 1982); however, we judge this to be a small effect compared to temperature variations for open-ocean settings (see following discussion).

We have separated the Cretaceous isotope compilation into four intervals based upon our interpretation of the data: (1) the increasing temperatures before 97 Ma; (2) the subsequent Cretaceous hot greenhouse interval (paralleled by increasing $\delta^{13}C_b$ values); (3) the long-lasting cooling trend and decrease in $\delta^{13}C_b$ between 91 and 78 Ma; and (4) the interval after 78 Ma with small interocean $\delta^{13}C_b$ and $\delta^{18}O_b$ values and a major increase in both $\delta^{18}O_b$ and $\delta^{13}C_b$ between 70 and 68 Ma (Fig. 2).

The increasing warming before 97 Ma (interval 1) and subsequent extremely warm period between 97 and 91 Ma (interval 2) is known from low-resolution stable isotope studies (Clarke and Jenkyns, 1999; Huber et al., 2002) and has been related to increasing amounts of greenhouse gases in the atmosphere (Bice et al., 2006). Stable isotope and Mg/Ca data of planktic foraminifera (Wilson et al., 2002; Norris et al., 2002; Bice et al., 2006; Bornemann et al., 2008) and the TEX_{86} proxy (Forster et al., 2007) suggest tropical sea-surface temperatures >35 °C for this peak interval of the Cretaceous hot greenhouse, and therefore surface ocean temperatures ~7–8 °C hotter than today. Our benthic foraminifer stable isotope data suggest that intermediate to bottom waters (IBW) in the southern high latitudes and the Pacific Ocean were as warm as 20 °C, whereas the tropical proto–North Atlantic shows even higher temperatures (Fig. 2; Friedrich et al., 2008). These high temperatures can be partly explained by polar warmth at the sites of high-latitude deep-water overturning (Huber et al., 2002). The high temperatures in the deep proto–North Atlantic likely reflect the existence of shallow sills that prevented all but warm thermocline or intermediate waters from entering the basin (Poulsen et al., 2001). Formation of warm, saline water within the surrounding epicontinental shelf seas may also have contributed to waters filling the deeper North Atlantic (Friedrich et al., 2008; Berrocoso et al., 2010), a mechanism working on a local scale rather than influencing the whole ocean. The existence of these warm, saline waters from 97 to 91 Ma may have resulted in δ_w having been more positive, meaning that our paleotemperature estimates (Fig. 2) are probably too low.

The massive storage of organic carbon in black shales led to a positive trend in $\delta^{13}C_b$ between 97 and 90 Ma. The $\delta^{13}C_b$ values of the tropical Atlantic Ocean, however, are extremely

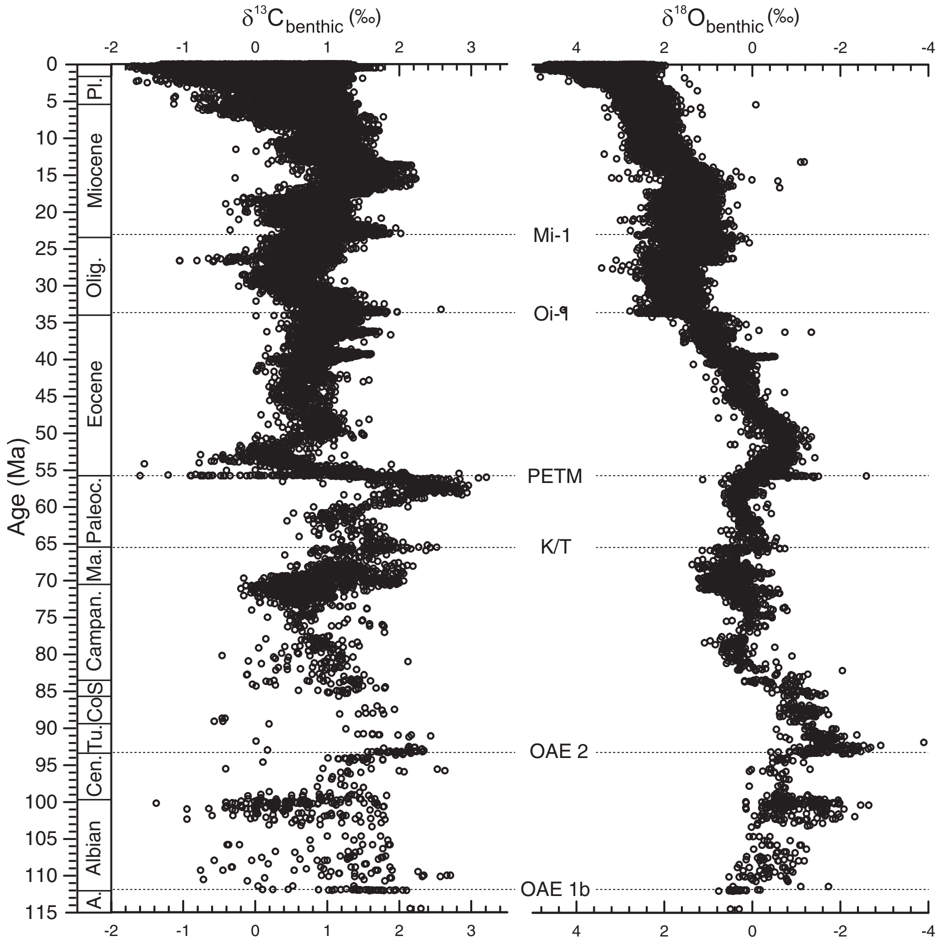

Figure 3. Stable oxygen and carbon isotope compilation of benthic foraminifera for past 115 m.y. Cenozoic data after Cramer et al. (2009). Note that Cenomanian to early Turonian benthic $\delta^{18}O$ data of Demerara Rise are not included in this compilation due to influence of warm, saline intermediate to bottom waters (IBW). The benthic $\delta^{13}C$ data from Demerara Rise are not included due to negative offset (see text). Mi-1—major Miocene glaciation event, Oi-1—major Oligocene glaciation event, PETM—Paleocene-Eocene thermal maximum; K/T—Cretaceous-Tertiary boundary, OAE—oceanic anoxic event, A.—Aptian, Cen.—Cenomanian, Tu.—Turonian, Co—Coniacian, S—Santonian, Campan.—Campanian, Ma.—Maastrichtian, Paleoc.—Paleocene, Olig.—Oligocene, Pl.—Pliocene.

negative when compared to other ocean basins (Fig. 2). This phenomenon likely reflects a long residence time of local bottom waters due to relatively shallow sills between the North Atlantic and the Pacific and South Atlantic Oceans. Nonetheless, the most significant carbon cycle perturbation is not associated with OAE formation during the middle Cretaceous, but occurred in the Middle to Late Paleocene, a time of relatively cool climate (Fig. 3). Although the Cretaceous OAEs were major periods of oil formation, they may not involve as large a global episode of carbon burial and subsequent exhumation as in the Paleocene.

Interval 3 (91–78 Ma) shows two distinctive parts. Between 91 and 84 Ma, $\delta^{18}O_b$ values of all ocean basins (where data are available) are similar. The common trend may result from the initial opening of the Equatorial Atlantic Gateway (EAG) between Africa and South America. Deepening this gateway would allow South Atlantic intermediate water to invade the proto–North Atlantic, filling the deep Atlantic with cool high-latitude waters. While several studies claim that the EAG was breached in the early Turonian (Tucholke and Vogt, 1979; Summerhayes, 1987; Pletsch et al., 2001), climate models (Handoh et al., 1999), benthic foraminifera (Friedrich and Erbacher, 2006), and sedimentological evidence (van Andel et al., 1977; Jones et al., 1995) suggest that a permanent deep-water connection was not established until the Santonian to Campanian. Our data support a substantial connection between the North and South Atlantic by the late Turonian. Between ca. 84 and 78 Ma, we see a separation of water masses between the Southern Ocean and the cooler temperatures and higher $\delta^{13}C$ values of the Pacific, possibly reflecting multiple IBW sources.

The last 13 m.y. of the Cretaceous are characterized by small interbasin gradients for both $\delta^{18}O_b$ and $\delta^{13}C_b$. This could be related to a full connection between all ocean basins (especially due to the complete opening of the EAG). Nonetheless, the North Atlantic shows more positive $\delta^{13}C_b$ values throughout the Maastrichtian than the other oceans, which may reflect local formation of deep water (Frank and Arthur, 1999; Friedrich et al., 2004). The Pacific, in contrast, exhibits the lowest $\delta^{13}C_b$ values during this time, possibly reflecting a more distant source region. A pronounced positive $\delta^{13}C_b$ excursion between 70 and 68 Ma is obvious in all data sets (Fig. 1), reflecting global carbon burial accompanied by a positive $\delta^{18}O_b$ excursion interpreted to reflect a short-term glaciation event (e.g., Barrera and Savin, 1999; Miller et al., 1999) and/or reorganization of IBW sources (e.g., Barrera et al., 1997; Friedrich et al., 2009).

115 MA COMPILATION

Comparison of our data with the Cenozoic compilation (0–65.5 Ma; Cramer et al., 2009; for comparison with the compilation of Zachos et al., 2008, see Fig. DR3) shows that the interval between 97 and 91 Ma was by far the warmest greenhouse climate during the past 115 m.y. After this interval of extreme warmth, the deep oceans started to cool, a process continuing to today. Only two greenhouse periods interrupted the cooling, i.e., the Early Eocene (47–55 Ma) and the Early to Middle Miocene (15–20 Ma; Fig. 3), both of which are comparable in their duration with the middle Cretaceous thermal maximum. Intermediate and bottom waters of the Cenozoic warm periods never reached the high temperatures of the Cretaceous hot greenhouse climate. Instead, extreme Cretaceous warmth was probably ended by the tectonic dissection of silled North Atlantic basins and the opening of oceanic connections to the poles that allowed waters cooled in the polar winter to flood the rest of the deep ocean.

The reason for the general cooling trend between 91 and 78 Ma is still unresolved. Low-resolution isotope data from the southern high latitudes and the North Atlantic show only a small decrease in temperatures, leading to the suggestion that the entire interval was part of the Cretaceous hot greenhouse (Huber et al., 2002). Based on our compilation, however, we propose that deepening of the EAG and the parallel reorganization of the oceanic circulation with longitudinal water-mass and heat exchange within the Atlantic Ocean favored cooling by the late Turonian. These tectonic effects may have been strengthened by the parallel decrease of greenhouse gases (Bice et al., 2006; Pucéat et al., 2007), decreasing ocean-crust production (Larson, 1991), and diminished formation of warm and saline intermediate water masses.

This explanation is also supported by the general decrease in $\delta^{13}C_b$ values that is obvious in all ocean basins. A better connection of the former restricted proto–North Atlantic basin, and therefore an input of oxygenated waters from the south, would allow the oxidization of the organic-rich sediments formed in this basin and mark the end of the proto–North Atlantic as a carbon sink.

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