Greenhouse world and the Mesozoic Ocean

by

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INTRODUCTION

The Earth's climate has alternated between greenhouse (warm) and icehouse (cool) modes throughout the Phanerozoic (Fig. 1A). Earth is in the midst of an icehouse climate at present. Nevertheless, the rise of industrialization in the last two centuries has led to a dramatic increase in atmospheric CO_2 from the burning of fossil fuels, which, in turn, has led to significant global warming (e.g., Ruddiman, 2000). Global warming could profoundly impact human life because of various environmental changes, including global sea-level rise, more numerous and more powerful hurricanes and heavier rain and snow precipitation. Understanding the ocean–climate system during past greenhouse climate modes is essential for more accurate prediction of future climate and environmental changes in the warming Earth.

The Mesozoic–early Cenozoic is known as a typical greenhouse period caused largely by increased CO_2 from elevated global igneous activity (Fig. 1A–C). The mid-Cretaceous marked a major warming peak (Fig. 1D), as it is characterized by globally averaged surface temperatures that were >14°C higher than those of today (Tarduno et al., 1998), a lack of permanent ice sheets (Frakes et al. 1992) and ~100–200 m higher sea level than that of today (Haq et al., 1987; Miller et al., 2005a; Fig. 1E). Studies using DSDP (Deep Sea Drilling Program) and ODP (Ocean Drilling Program) cores have advanced our understanding of Mesozoic oceanography and climate. These studies demonstrated that Mesozoic ocean circulation and marine ecosystems differed greatly from those of today. This paper reviews significant achievements of DSDP and ODP research, and discusses future prospects of IODP (Integrated Ocean Drilling

Program) in the field of Mesozoic oceanography.

NEW INSIGHTS ON MESOZOIC OCEANOGRAPHY FROM ODP AND DSDP RESEARCH

Determination of Mesozoic Ocean Temperature History

An important achievement of DSDP and ODP was the reconstruction of the history of Mesozoic ocean temperature changes based on geochemical methods such as oxygen isotopes, TEX_{86} and alkenone analyses. Oxygen isotope data have provided the greatest source of paleotemperature reconstructions from ancient oceans. However, the increasing prevalence of diagenetic alteration in older or more deeply buried rocks limits or prevents reliable isotopic data from biogenic calcite preserved in terrestrial outcrops. Compared to many land-based sections, calcareous microfossils of Cretaceous age recovered from samples drilled at DSDP and ODP sites are often better preserved, and usually have not been as seriously affected by complex tectonic and/or weathering processes. Exquisitely preserved foraminifera from the low-latitude Demerara Rise (ODP Site 1258–1261; 4–15˚N), mid-latitude Blake Nose (ODP Sites 1049, 1050, 1052; 30˚N) and high-latitude Falkland Plateau (DSDP Site 511; 60˚S) have been especially useful for reconstructing vertical and latitudinal temperature gradients of the mid- through Late Cretaceous ocean (Figs. 2 & 3). The TEX_{86} method is especially useful for organic carbon-rich sediments, and has provided excellent paleo-temperature determinations (e.g., Schouten et al., 2003; Jenkyns et al., 2004).

According to isotopic records of surface dwelling planktic foraminifera, sea surface temperatures reached a maximum of 42˚C at Demerara Rise (Bice et al., 2006), 33˚C at Blake Nose (Huber et al., 2002), and 31˚C at Falkland Plateau (Huber et al., 2002; Bice et al., 2003) during the Turonian (Figs. 2 & 3). At comparable latitudes in the modern ocean, August surface water temperatures are 25–28˚C at 0–20˚N, 20–28˚C at 20–40˚N, and 0–5˚C at 60˚S (Thurman and Trujillo, 1999). These data consequently suggest that Cretaceous warming was most prominent at high latitudes where the difference of temperature between the mid-Cretaceous and the present oceans is nearly 30˚C (Fig. 2). Bice and Norris (2002) estimate that at least 4500 ppm CO_2 is required to match the above-mentioned maximum temperatures, which is >11 times the modern atmospheric concentration. Using a more recent climate model Bice et al. (2006) conclude that 3500 ppm or greater atmospheric CO_2 concentration is required to reproduce the estimated maximum sea surface temperatures of the Mesozoic tropical ocean.

Since the Mesozoic paleo-temperature estimates based on geochemical proxies are still insufficient in sediments older than Albian and in the areas outside of the Atlantic Ocean, further investigations are needed to reconstruct a reliable spatial and temporal temperature history during the greenhouse climate of the Mesozoic.

Oceanic Anoxic Events

Defining the concept of Oceanic Anoxic Events (OAEs) was one of the most important achievements of the early DSDP. Cretaceous marine sediments in Europe are mainly made up of white limestone and chalk; however, distinct black, laminated organic rich layers, termed "black shales", are occasionally intercalated within these sequences (Fig. 4). Because organic carbon is preferentially preserved under anoxic conditions, earlier workers suggested that these black shales had accumulated locally in a weakly ventilated, restricted basin under regional anoxic conditions. In the mid-1970s, however, the discovery of black shales at many DSDP drill sites from the Atlantic, Indian and Pacific Oceans led to recognition of widespread anoxic conditions in the global ocean spanning limited stratigraphic horizons (Fig. 5). Schlanger and Jenkyns (1976) termed these wide spread depositional black shale intervals "Oceanic Anoxic Events" (OAEs).

Burial of organic carbon, which preferentially sequesters isotopically light carbon during OAEs, resulted in a positive $\delta^{13}C$ ($^{13}C/^{12}C$) excursion of 2-3‰ in the geologic record (Fig. 3). Even if black shales are not visible in terrestrial rocks such as dark gray- to black-colored mudstones, carbon isotope excursions are a useful marker for recognizing the OAEs (Takashima et al., 2004). Recent advances in biostratigraphy and correlation using carbon isotopes have revealed that OAEs occurred at least 8 times in the Cretaceous and at 1 to 4 times in the Jurassic (Fig. 3). The Toarcian OAE, Weissert OAE, OAE 1a and OAE 2 are global scale anoxic events associated with prominent positive excursions of $\delta^{13}C$ and worldwide distribution of black shales (Fig. 3).

Two models, that of a stagnant ocean or expansion of the oxygen-minimum layer, have been proposed to explain the formation of black shales in the OAEs (e.g., Pedersen and Calvert, 1990). The stagnant ocean model (STO model) attributes OAEs to depletion of bottom water oxygen as a result of dense vertical stratification of the ocean

(Fig. 6A). A modern analogue is seen in stratified silled-basins such as the Black Sea. The expanded oxygen-minimum layer model (OMZ model) proposes that increased surface ocean productivity caused expansion of the oxygen minimum layer in the water column (Fig, 6B). Upwelling sites such as the Moroccan and Peruvian margins provide a modern analogue for this model.

These two models predict different vertical thermal gradient profiles of the water column that can be inferred from the oxygen isotope of planktic and benthic foraminifera. For example, the OAE 1b in the earliest Albian (about 112 Ma) is characterized by a sudden increase in surface water temperatures and strengthening of the vertical stratification of the water column (Erbacher et al., 2001), suggesting similarity to the STO model (Fig. 7A). On the other hand, the OAE 2 (about 94 Ma) shows sudden warming of deep-water and collapse of vertical stratification (Huber et al., 1999), which probably induced enhanced upwelling and productivity similar to the expanded OMZ model (Fig. 7B). Warming of deep-waters may have contributed to a decrease in oxygen solubility in the deep ocean, as well as triggering the disassociation of large volumes of methane hydrate buried in sediments of the continental margins. Oxidation of the released methane could have further consumed dissolved oxygen in the water column, while simultaneously releasing CO_2 to the atmosphere (Gale, 2000; Jahren, 2002). However, since there really is no modern analog for global ocean anoxia, these models suffer from the lack of an analog.

OAEs have had a significant influence on the evolution and diversity of ancient marine communities through the Phanerozoic. Numerous records demonstrate a high turnover rate of microfossils at or near OAE intervals (Jarvis et al., 1988; Erbacher et al., 1996; Premoli Silva and Sliter, 1999; Leckie et al., 2002; Erba, 2004). During the Cenomanian-Turonian (C/T) boundary OAE 2, for example, anoxic environments expanded from the photic zone (Damsté and Köster, 1998; Pancost et al., 2004) to >3500 m depth in the Atlantic (Thurow et al., 1992), resulting in about 20% extinction of marine organisms in various habitats within an interval of less than 1 million years (Fig. 1F). Black shales in the OAEs, especially OAE 1a and OAE 2, frequently yield no calcareous nannofossils, planktic foraminifera or radiolarians, suggesting that anoxic conditions had expanded to within the euphotic zone of the surface the water column (e.g. Hart and Leary, 1991; Coccioni and Luciani, 2005). Discovery of abundant cyanobacteria biomarkers (e.g., Kuypers et al., 2004), nonthermophilic archaea (e.g., Kuypers et al., 2001), and green sulfur bacteria (e.g., Damesté and Köster, 1998) within the black shales provides strong support for this hypothesis. These proxies further indicate that anoxic conditions occasionally occurred at very shallow water depths during the C/T OAE.

OAEs also served as an effective thermostat for the greenhouse Earth. Since the change in organic burial in the pelagic sections for OAEs was 2 to 3 orders of magnitude greater than the mean conditions at other time interval, burial of massive organic carbon during OAEs may have drawn down CO_2 from the ocean–atmosphere by burying organic carbon in black shales thereby punctuating long-term global warmth (e.g., Arthur et al., 1988; McElwain et al., 2005). The Late Devonian anoxic event could be an extreme example where widespread anoxia caused not only significant biotic extinction (about 40%), but also induced glaciation after deposition of black shales (Caplan and Bustin, 2001).

OAEs have benefited human life because they are a major cause of the large volumes of oil and gas that we consume today. These hydrocarbons were derived from organic-rich sediments that formed under anoxic conditions. Indeed, many petroleum source rocks were formed during in greenhouse warming peaks between the Late Jurassic and mid-Cretaceous (Fig. 1G).

Mesozoic sea level changes and existence of ice-sheet

Rising sea level attributed to global warming is one of the most serious and imminent problems for mankind because of the concentration of human populations on the coastal plains. Fluctuations in global sea level result from changes in the volume of ocean or the volume of ocean basins. The former depends mainly on the growth and decay of continental ice sheets and fluctuates on short (10^4–10^6 year) time scales. On the other hand, the latter fluctuates on longer (10^7–10^8 year) time scales resulting from tectonic effects such as variations in seafloor spreading rates, ocean ridge lengths and collision/break-up of continents (e.g., Ruddiman, 2000; Miller et al., 2005a, b). Because the Mesozoic period exhibited the break-up of Gondwana, primarily ice-free climates, high rates of seafloor spreading, as well as the emplacement of large igneous plateaus on the ocean floor, the Mesozoic ocean was characterized by much higher sea level than at present. Sea level peaked in mid- to Late Cretaceous time (~100–75 Ma), during which continents were flooded more than 40% in area of present land resulting in the expansion of continental shelf environments and intra-continental seaways (e.g., Hays and Pitmann III, 1973; Fig. 5).

The most widely cited reconstructions of past sea level changes were established by Exxon Production Research Company (EPR) (Haq et al., 1987), which have been up-dated in the past decade (e.g., Hardenbol et al., 1998). These sea level curves consist of short- (10^5–10^6 year) and long-term (10^7–10^8 year) curves that are correlated with detailed chrono-, bio- and magnetostratigraphies for last 250 million years (Fig. 3). According to the EPR curves, Late Cretaceous sea level rose as much as 260 m above the present level. The EPR curves, however, have been criticized because of the following reasons: 1) the supporting data are proprietary, 2) the sequence boundaries cannot be translated into a eustatic origin, and 3) inferred amplitudes of sea level fluctuations seem to be conjectural (e.g., Christie-Blick et al., 1995). ODP drilling on the New Jersey passive continental margin (ODP 174AX) provided on new insights into the amplitudes of, and mechanisms for, sea level changes for last 100 Ma. The area around the drilling sites is an excellent location for sea level studies because of quiescent tectonics and well-constructed biostratigraphic and Sr isotopic age control (Sugarman et al., 1995). The proposed sea level curve by the New Jersey drilling is well correlated with those of Russian platform and EPR curves, but the estimated maximum global sea level amplitude is ~100 m during Late Cretaceous (Miller et al., 2005a; Fig. 3), which is contrast to the much greater estimate by EPR.

Since the Mesozoic greenhouse period is generally assumed to have been equably ice-free interval, it has long been debated about the mechanism for the large and rapid changes observed in Cretaceous sea level (e.g., Skelton et al., 2003). Miller and his colleagues demonstrated that several rapid sea level falls recorded in New Jersey could be explained only by glacio-eustacy (Miller et al., 1999; 2005a). According to integration between occurrences of ice-rafted and/or glacial deposits around the polar regions, positive oxygen isotope values of foraminifera and intervals of rapid sea level fall, it is quite possible that the glacial events did occur during greenhouse climate. Although there still remains uncertainness in age and ice volume, several geologically short-term glacial events during Cretaceous have been proposed (e.g., middle Cenomanian [96 Ma], middle Turonian [92–93 Ma], middle Campanian, and earliest and late Maastrichtian [71 and 66.1 Ma]). These results imply that greenhouse periods had much greater short-term climatic variability instead of previously proposed long-term stable and equable climates.

Biocalcification crises during the Mesozoic ocean

The Mesozoic is marked by the poleward expansion shallow-water carbonate platforms as well as several occurrences of their global “drowning” or “collapse” events (e.g., Johnson et al., 1998; Simo et al., 1993). These drowning events were not due to sea level rise because shallow-water carbonate platforms usually grow-up much faster than sea level fluctuation. Although eutrophication of surface oceans associated OAEs were considered to be the cause of these drowning events, ODP Legs 143 and 144 revealed that some shallow-water carbonate platforms survived during OAE 1a in the central Pacific (Wilson et al., 1998). Weissert and Erba (2004) pointed out that the coincidence between drowning events of shallow-water carbonate platforms and the crisis of heavily calcified plankton groups, and termed these events “Biocalcification crises”. Although the mechanism responsible for biocalcification crises remains poorly constrained, recent hypotheses blame elevated pCO_2-induced lowered surface ocean pH, which affected carbonate-secreting organisms (e.g., Leckie et al., 2002; Weissert and Erba, 2004).

FUTURE PROSPECTS OF THE IODP FOR MESOZOIC OCEANOGRAPHY

The greenhouse climate of the mid-Cretaceous was likely related to major global volcanism and associated outgassing of CO_2. OAEs may be recognized as a negative feedback in response to sudden warming episodes, by preventing further acceleration of warming through removal of organic carbon from the ocean-atmosphere (CO_2) reservoir to sediment reservoirs. This process resulted in the emplacement of a large volume of organic matter during the mid-Cretaceous, which now serves as a major source of fossil fuels (Larson, 1991). However, present human activities are rapidly consuming these fuels, returning the carbon to the ocean-climate system. Pre-industrial CO_2 levels of about 280 ppm have increased over the past 200 years to the current levels exceeding 380 ppm, mainly as a result of human activities. Bice et al. (2006) estimated that Cretaceous atmospheric concentration ranged between 600 and 2400 ppm, 1.5 to 6 times the present concentration. If the current rate of CO_2 increase continues, Cretaceous values may be attained within 1500–6000 years, but current trends are already having clear affects on the both the ocean-climate system and the biosphere. In fact, a recent ocean-climate model predicts that rapid atmospheric release of CO_2 will produce changes in ocean chemistry that could affect marine ecosystems significantly, even under future pathways in which most of the remaining fossil fuel CO_2 is never released (Caldeira and Wickett, 2005).

Improved understanding of the Mesozoic ocean-climate system and formation of OAEs are important to better predict environmental and biotic changes in a future greenhouse world. However, Cretaceous DSDP and ODP cores with continuous recovery and abundant well-preserved fossils suitable for isotopic study are very limited. A denser global array of deep-sea cores is needed to provide more detailed reconstructions of global climate changes and oceanographic conditions in order to better understand the ocean-climate dynamics of the Mesozoic greenhouse Earth. Though far from complete, the Mesozoic record is much better studied in areas of the Atlantic Ocean and Tethys Sea than in the Indo–Pacific Oceans because most of ocean-floor formed in Mesozoic time has already subducted under continents. Therefore, much less is known about Mesozoic paleoceanographic condition in the Indo-Pacific (Bralower et al., 1993). The Mesozoic marine sequences deposited at middle–high latitudes of the Pacific, such as the continental margin of eastern Asia and the Bering Sea, are appropriate future drilling targets. Submerged continental rift sites such as the Somali Basin should also be targeted as they record a continuous paleoceanographic history from the Early Cretaceous or older. We expect that new IODP research from these Cretaceous sites could provide new insights to the process of abrupt global warming and its impact on the Earth's biosphere.

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REFERENCES

Abramovich, S., G. Keller, D. Stüben and Z. Berner. 2003. Characterization of late Campanian and Maastrichtian planktonic foraminiferal depth habitats and vital activities based on stable isotopes. *Palaeogeography, Palaeoclimatology, Palaeoecology* 202: 1-29.

Arthur, M. A., S. O.Schlanger, and H. C. Jenkyns. 1987. The Cenomanian–Turonian oceanic anoxic event, II. Palaeoceanographic controls on organic-matter production and preservation. Pp. 401-422 in *Marine Petroleum Source Rocks*. J. Brooks and A. J. Fleet, eds. Geological Society Special Publication 26, Blackwell, Oxford.

Arthur, M. A., W. E. Dean, and L. M. Pratt. 1988. Geochemical and climatic effects of increased marine organic carbon burial at the Cenomanian/Turonian boundary. *Nature* 335: 714-717.

Berner, R. A. in press. GEOCARBSULF: A combined model for Phanerozoic atmospheric O_2 and CO_2. *Geochimica et Cosmochimica Acta*.

Bice, K. L. and R. D. Norris. 2002. Possible atmospheric CO_2 extremes of the middle Cretaceous (late Albian-Turonian). *Paleoceanography* 17: 1029/2002PA000778.

Bice, K. L., B. T. Huber and R. D. Norris. 2003. Extreme polar warmth during the Cretaceous greenhouse?: Paradox of the Late Turonian d^{18}O record at DSDP Site 511, *Paleoceanography*, 18: 1029/2002PA000848.

Bice, K. L., D. Birgel, P. A. Meyers, K. A. Dahl, K-U. Hinrichs and R. D. Norris. 2006. A multiple proxy and model study of Cretaceous upper ocean temperatures and atmospheric CO_2 concentrations. *Paleoceanography* 21: 1029/2005PA001203.

Bralower, T. J., W. V. Sliter, M. A. Arthur, R. M. Leckie, D. Allard, and S. O. Schlanger. 1993. Dysoxic/anoxic episodes in the Aptian-Albian (Early Cretaceous). Pp. 5-37 in *The Mesozoic Pacific: Geology, Tectonics and Volcanism*. M. S. Pringle et al., eds. American Geophysical Union, Geophysical Monograph 77, Washington, D. C.

Caldeira, K. and M. E. Wickett. 2005. Ocean model predictions of chemistry changes from carbon dioxide emissions to the atmosphere and ocean. *Journal of Geophysical Research* 110; C09S04.

Caplan, M. L. and M. R. Bustin. 2001. Devonian–Carboniferous Hangenberg mass extinction event, widespread organic-rich mudrock and anoxia: causes and consequences. *Palaeogeography, Palaeoclimatology, Palaeoecology* 148: 187-207.

Christie-Blick, N., G. S. Mountain and K. G. Miller. 1990. Seismic stratigraphic record of sea-level change. Pp. 116-140 in *Sea-Level Change*. R. R. Revelle, eds, National Academy Press, Washington, D. C.

Coccioni, R. and V. Luciani. 2005. Planktonic foraminifers across the Bonarelli Event (OAE2, latest Cenomanian): The Italian record. *Palaeogeography,*

Palaeoclimatology, Palaeoecology 224: 167-185.
Damesté, J. S. S. and J. Köster. 1998. A euxinic southern North Atlantic Ocean during the Cenomanian/Turonian oceanic anoxic event. *Earth and Planetary Science Letters* 158: 165-173.
Dromart, G., J-P. Garcia, F. Gaumet, S. Picard, M. Rousseau, F. Atrops, C. Lecuyer and S. M. F. Sheppard. 2003. Perturbation of the carbon cycle at the Middle/Late Jurassic transition: geological and geochemical evidence. *American Journal of Science* 303: 667-707.
Erba, E. 2004. Calcareous nannofossils and Mesozoic oceanic anoxic events. *Marine Micropaleontology* 52: 85-106.
Erbacher, J., J. Thurow, R. Littke. 1996. Evolution patterns of radiolaria and organic matter variations: a new approach to identify sea-level changes in mid-Cretaceous pelagic environments. *Geology* 24: 499-502.
Erbacher, J., B. T. Huber, R. D. Norris and M. Markey. 2001. Increased thermohaline stratification as a possible cause for an oceanic anoxic event in the Cretaceous period. *Nature* 409: 325-327.
Frakes, L. A., J. E. Francis and J. I. Syktus. 1992. *Climate modes of the Phanerozoic*. Cambridge University Press, Cambridge, 274pp.
Fisher, J. K., G. D. Price, M. B. Hart and M. J. Leng. 2005. Stable isotope analysis of the Cenomanian–Turonian (Late Cretaceous) oceanic anoxic event in the Crimea. *Cretaceous Research* 26: 853-863.
Gale, A. S. 2000. The Cretaceous world, Pp. 4-19 in *Bioic response to global change: The last 145 million years*. S. J. Culver and P. F. Raqson, eds, Cambridge University press, Cambridge.
Haq, B. U., J. Hardenbol and P. R. Vail. 1987. Chronology of fluctuating sea levels since the Triassic. *Science* 235:1156-1167.
Hardenbol, J., J. Thierry, M. B. Farley, T. Jaquin, P.-C. de Graciansky and P. R. Vail. 1998. Cretaceous chronostratigraphy, Pp. 3-13 in *Mesozoic and Cenozoic sequence chronostratigraphic framework of European basins*. P.-C. Graciansky, J. Hardenbol, T. Jaquin, P. R. Vail, eds. SEPM Spec. Pub. 60.
Hart, M. B. and P. N. Leary. 1991. Stepwise mass extinctions: the case for the Late Cenomanian event. *Terra Nova* 3: 142-147.
Hays, J. D. and W. C. Pitman III. 1973. Lithospheric plate motion, sea level changes

and climatic and ecological consequences. *Nature* 246: 18-22.

Hesselbo, S. P., D. Gröcke, H. C. Jenkyns, C. J. Bjerrum, P. Farrimond, H. S. M. Bell and O. R. Green. 2000. Massive dissociation of gas hydrate during a Jurassic oceanic anoxic event. *Nature* 406: 392-395.

Huber, B. T., R. M. Leckie, R. D. Norris, T. J. Bralower and E. CoBabe. 1999. Foraminiferal assemblage and stable isotopic change across the Cenomanian–Turonian boundary in the subtropical North Atlantic. *Journal of Foraminiferal Research* 29: 392-417.

Huber, B. T., R. D. Norris and K. G. MacLeod. 2002. Deep-sea paleotemperature record of extreme warmth during the Cretaceous. *Geology* 30: 123-126.

Jahren, A. H. 2002. The biogeochemical consequences of the mid-Cretaceous superplume. *Journal of Geodynamics* 34: 177-191.

Jarvis, I., G. A. Garson, M. K. F. Cooper, M. B. Hart, P. N. Leary, B. A. Tocher, D. Horne and A. Rosenfeld. 1988. Microfossil assemblages and the Cenomanian-Turonian (Late Cretaceous) Oceanic Anoxic Event. *Cretaceous Research*, 9: 3-103.

Jarvis, I., A. Mabrouk, R. T. J. Moody and S. D. Cabrera. 2002. Late Cretaceous (Campanian) carbon isotope events, sea-level change and correlation of the Tethyan and Boreal realms. *Palaeogeography, Palaeoclimatology, Palaeoecology* 188: 215-248.

Jenkyns, H.C. 1991. Impact of Cretaceous sea level rise and anoxic events on the Mesozoic carbonate platform of Yugoslavia. *American Association of Petroleum Geologists Bulletin* 75: 1007-1017.

Jenkyns, H. C., A. S. Gale and R. M. Corfield. 1994. Carbon- and oxygen-isotope stratigraphy of the English chalk and Italian Scaglia and its paleoclimatic significance. *Geological Magazine* 131: 1-34.

Jenkyns, H. C., A. Forster, S. Schouten and J. S. Sinninghe Damsté. 2004. High temperatures in the Late Cretaceous Arctic Ocean. *Nature* 432: 888-892.

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

Jones, C. E. and H. C. Jenkyns. 2001. Seawater strontium isotopes, oceanic anoxic events, and seafloor hydrothermal activity in the Jurassic and Cretaceous. *American*

Journal of Science 301: 112-149.

Kassab, A. S. and N. A. Obaidalla. 2001. Integrated biostratigraphy and inter-regional correlation of the Cenomanian-Turonian deposits of Wadi Feiran, Sinai, Egypt. *Cretaceous Research* 22: 105-114.

Klemme, H. D. and G. F. Ulmishek. 1991. Effective petroleum source rocks of the world: stratigraphic distribution and controlling depositional factors. *American Association of Petroleum Geologists Bulletin* 75: 1809-1851.

Kuypers, M. M. M., P. Blokker, J. Erbacher, H. Kinkel, R. D. Pancost, S. Schouten and . J. P. S. Damsté. 2001. Massive expansion of marine archaea during a mid-Cretaceous oceanic anoxic event. *Science*, 293: 92-94.

Kuypers, M. M. M., S. Schouten, E. Erba, and J. P. Sinninghe Damsté. 2004. N_2-fixing cyanobacteria supplied nutrient N for Cretaceous oceanic anoxic events. *Geology*, 32: 853-856.

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

Lebedeva, N. K. and K. V. Zverev. 2003. Sedimentological and palynological analysis of the Cenomanian–Turonian event in northern Siberia. *Geologiya I Geofizika* 44: 769-780.

Leckie, R.M., T. J. Bralower and R. Cashman. 2002. Oceanic anoxic events and plankton evolution: biotic response to tectonic forcing during the mid-Cretaceous. *Paleoceanography* 17: 1-29.

McElwain, J. C., J. Wade-Murphy and S. P. Hesselbo. 2005. Changes in carbon dioxide during an oceanic anoxic event linked to intrusion into Gondwana coals. *Nature* 435: 479-482.

Miller, K. G., E. Barrera, R. K. Olsson, P. J. Sugarman and S. M. Savin. 1999. Does ice drive early Maastrichtian eustasy? *Geology* 27: 783-786.

Miller, K. G., M. A. Kominz, J. V. Browning, J. D. Wright, G. S. Mountain, M. E. Katz, P. J. Sugarman, B. S. Cramer, N. Christie-Blick and S. F. Pekar. 2005a. The Phanerozoic record of global sea-level change. *Science* 310: 1293-1298.

Miller, K. G., J. D. Wright and J. V. Browning. 2005b. Visions of ice sheets in a greenhouse world. *Marine Geology* 217: 215-231.

Morettini, E., M. Santantonio, A. Bartolini, F. Cecca, P. O. Baumgartner and J. C. Hunziker. 2002. Carbon isotope stratigraphy and carbonate production during the

Early–Middle Jurassic: examples from the Umbria–Marche–Sabina Apennines (central Italy). *Palaeogeography, Palaeoclimatology, Palaeoecology* 184: 251-273.

Pancost, R. D., N. Crawford, S. Magness, A. Turner, H. C. Jenkyns and J. R. Maxwell. 2004. Further evidence for the development of photic-zone anoxic events. *Journal of the Geological Society, London* 161: 353-364.

Pedersen, T. F. and S. E. Calvert. 1990. Anoxia vs. Productivity: What controls the formation of organic-carbon-rich sediments and sedimentary rocks? *American Association of Petroleum Geologists Bulletin* 74: 454-466.

Premoli Silva, I. and W. V. Sliter. 1999. Cretaceous paleoceanography: Evidence from planktonic foraminiferal evolution, Pp. 301-328 in *Evolution of the Creaceous Ocean–Climate System*. E. Barrera and C. C. Johnson, eds, Geological Society of America Special Paper 332, Boulder.

Raup, D. M. and J. J. Jr. Sepkoski. 1986. Periodic extinction of Families and Genera. *Science* 231: 833-836.

Ridgwell, A. 2005. A mid-Mesozoic revolution in the regulation of ocean chemistry. *Marine Geology* 217: 339-357.

Royer, D. L. in press. CO_2-forced climate thresholds during the Phanerozoic. *Geochimica et Cosmochimica Acta*.

Ruddiman, W. F. 2000. *Earth's Climate: Past and Future*. W. H. Freeman and Company. New York, 465p.

Schlanger, S. O. and H. C. Jenkyns. 1976. Cretaceous oceanic anoxic events: causes and consequences. *Geologie en Mijnbouw* 55: 179-184.

Schlanger, S. O., M. A. Arthur, H. C. Jenkyns and P. A. Scholle. 1987. The Cenomanian–Turonian oceanic anoxic event, I. Stratigraphy and distribution of organic carbon-rich beds and the marine $\delta^{13}C$ excursion, Pp. 371-399 in *Marine Petroleum Source Rocks*. J. Brooks and A. J. Fleet, eds. Geological Society Special Publication 26, Blackwell, Oxford.

Schouten, S., E. C. Hopmans, A. Forster, Y. v. Breugel, M. M. M. Kuypers and J. S. Sinninghe Damsté. 2003. Extremely high sea-surface temperatures at low latitudes during the middle Cretaceous as revealed by archaeal membrane lipids. *Geology* 31: 1069-1072.

Simo J. A., R. W. Scott and J. -P. Masse. 1993. Cretaceous carbonate platforms: An overview, Pp. 1-14 in *Cretaceous Carbonate Platforms*, J.A. Toni Simo, R. W.

Scott and J. –P. Masse, eds. American Association of Petroleum Geologists Memoir 56, Tulsa.

Skelton, P. W., Spicer, R. A., Kelley, S. P. and Gilmour, L. 2003. *The Cretaceous World*. Cambridge University Press, Cambridge, 360p.

Stanley, S. M. and L. A. Hardie. 1998. Secular oscillations in the carbonate mineralogy of reef-building and sediment-producing organisms driven by tectonically forced shifts in seawater chemistry. *Paleogeography, Paleoclimatology, Paleoecology* 144: 3-19.

Stanley, S. M. 1999. *Earth system history*. W. H. Freeman and Company, New York, 615p.

Sugarman, P. J., Miller, K. G., Bukry, D. and M. D. Feigenson. 1995. Uppermost Campanian–Maestrichtian strontium isotopic, biostratigraphic, and sequence stratigraphic framework of the New Jersey Coastal Plain. *Geological Society of America Bulletin* 107: 19-37.

Takashima, R., F. Kawabe, H. Nishi, K. Moriya, R. Wani and H. Ando. 2004. Geology and stratigraphy of forearc basin sediments in Hokkaido, Japan: Cretaceous environmental events on the Northwest Pacific margin. *Cretaceous Research* 25: 365–390.

Tarduno, J. A., D. B. Brinkman, P. R. Renne, R. D. Cottrell, H. Scher and P. Castillo. 1998. Evidence for extreme climatic warmth from the Late Cretaceous Arctic vertebrates. *Science* 282: 2241-2244.

Thurman, H. V. and A. P. Trujillo. 1999. *Essentials of Oceanography–6th edition*. Prentice-Hall, New Jersey. 257p.

Thurow, J., H. J. Brumsack, J. Rullkötter, R. Littke and P. Meyers. 1992. The Cenomanian/Turonian boundary event in the Indian Ocean –a key to understand the global picture–, Pp. 253-273 in *Synthesis of Result from Scientific Drilling in the Indian Ocean*. American Geophysical Union, Geophysical Monograph 70, Washington D. C.

Van de Schootbrugge, B., T. R. Bailey, Y. Rosenthal, M. E. Katz, J. D. Wright, K. G. Miller, S. Feist-Burkhardt and P. G. Falkowski. 2005. Early Jurassic climate change and the radiation of organic-walled phytoplankton in the Tethys Ocean. *Paleobiology* 31: 73-97.

Wang, C. S., X. M. Hu, L. Jansa, X. Q. Wan and R. Tao. 2001. The

Cenomanian–Turonian anoxic event in southern Tibet. *Cretaceous Research* 22: 481-490.

Weissert, H. and H. Mohr. 1996. Late Jurassic climate and its impact on carbon cycling. *Palaeogeography, Palaeoclimatology, Palaeoecology* 122: 27-43.

Weissert, H. and E. Erba. 2004. Volcanism, CO_2 and palaeoclimate: a Late Jurassic–Early Cretaceous carbon and oxygen isotope record. *Journal of Geological Society, London* 161: 695-702.

Wilson, P. A., H. C. Jenkyns, H. Elderfield and R. L. Larson. 1998. The paradox of drowned carbonate platforms and the origin of Cretaceous Pacific guyots. *Nature* 392: 889-984.

Yurtsever, T. S., U. K. Tekin and Í. H. Demirel. 2003. First evidence of the Cenomanian/Turonian boundary event (CTBE) in the Alakirçay Nappe of the Antalya Nappes, southwest Turkey. *Cretaceous Research* 24: 41-53.

Figure Caption

Figure 1: Compilation showing the changes in climate, geological and paleontological events through the Phanerozoic.

Figure 2: Latitudinal variations of surface ocean paleo-temperature derived from oxygen isotopes of planktonic foraminifera and TEX_{86}. Modified from Bice et al. (2003), Huber et al. (2002) and Jenkyns et al. (2004).

Figure 3: Compilation showing Jurassic–Cretaceous changes in sea level, ocean crust production, paleo-temperature, bulk carbon isotopes, carbonate platform drowning events and OAEs. Data of Large Igneous Provinces are from Jones and Jenkyns (2001). Bulk carbon isotopes are from [1] Van de Schootbrugge et al. (2005), [2] Hesselbo et al. (2000), [3] Morettini et al. (2002), [4] Dromart et al. (2003), [5] Weissert et al. (1998), [6] Erbacher et al. (1996), [7] Jenkyns et al. (1994), [8] Jarvis et al. (2002), [9] Abramovich et al. (2003). Ages of the carbonate platform drowning are from Simó et al. (1993) and Weissert and Mohr (1996).

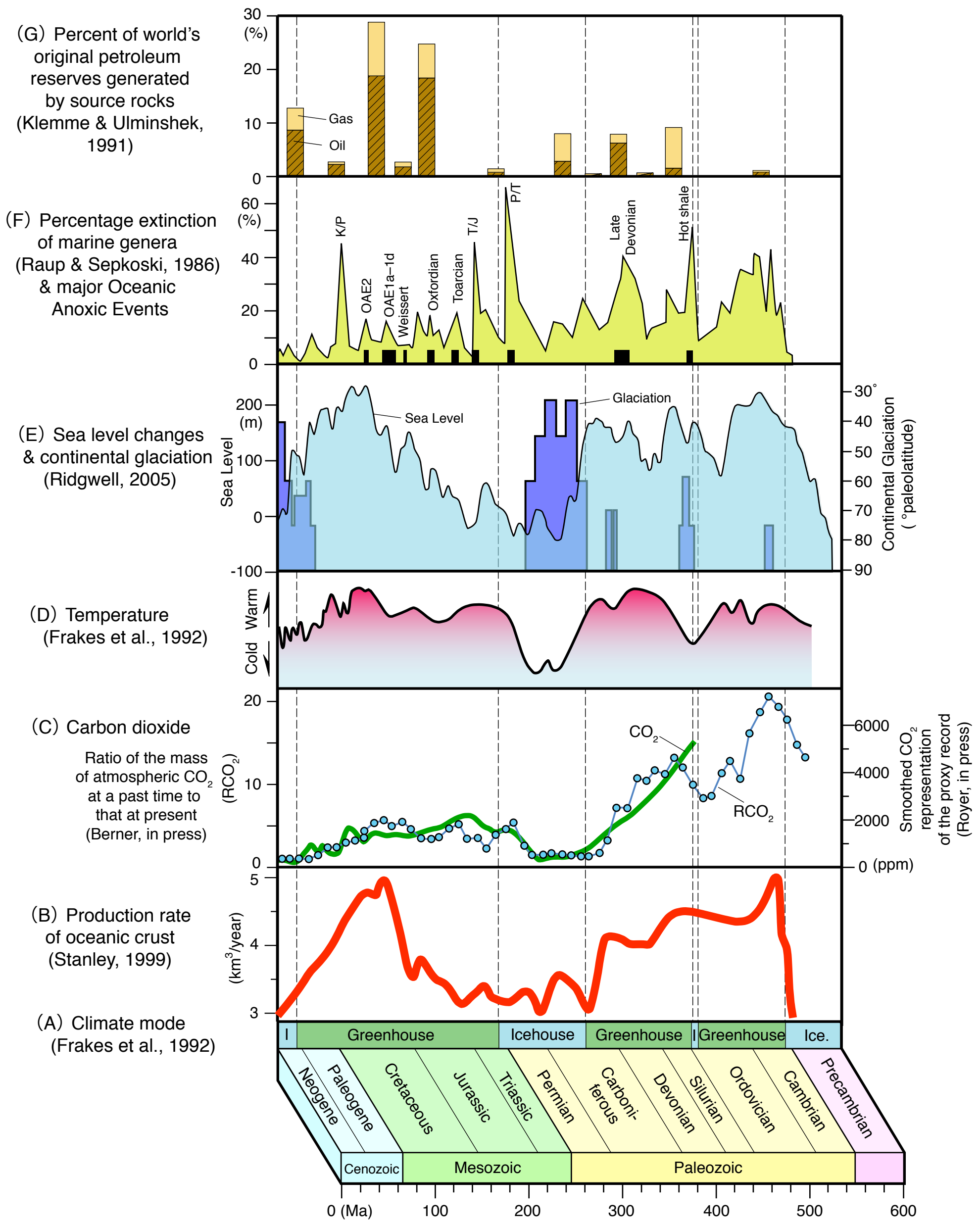
Figure 4: Cretaceous black shales intercalated in pelagic limestone sequence, central Italy. Provided by R. Coccioni.

Figure 5: Distribution of black shales and/or increased organic carbon sediments at OAE 2. Data are from Schlanger et al. (1987), Arthur et al. (1987; 1988),

Jenkyns, (1991), Thurow et al. (1992), Kassab and Obaidalla (2001), Wang et al. (2001), Lebedeva and Zverev (2003), Yurtsever et al. (2003), Coccioni and Luciani (2005), Fisher et al. (2005) and Takashima and Nishi unpublished data.

Figure 6: Representative models for black shale deposition. (A) stagnant ocean model, (B) oxygen minimum-layer model.

Figure 7: Vertical ocean temperature structure, reconstructed from oxygen isotopes, during (I) OAE 1b (Erbacher et al., 2001) and (II) OAE 2 (Huber et al., 1999) intervals at the Blake Nose, western North Atlantic.



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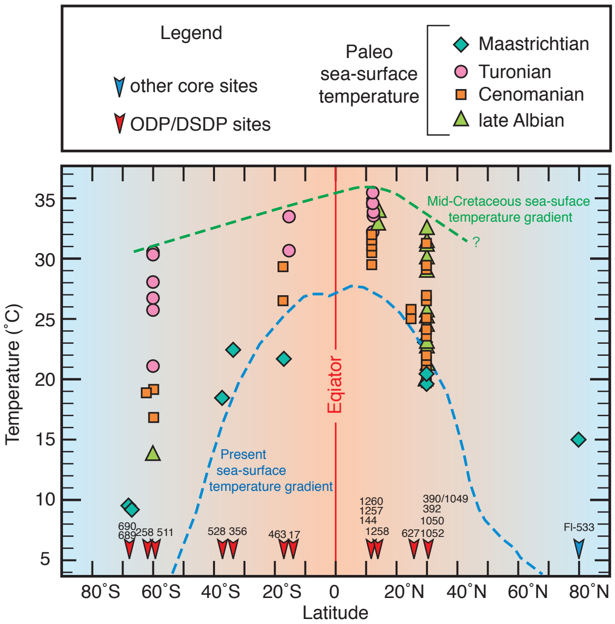

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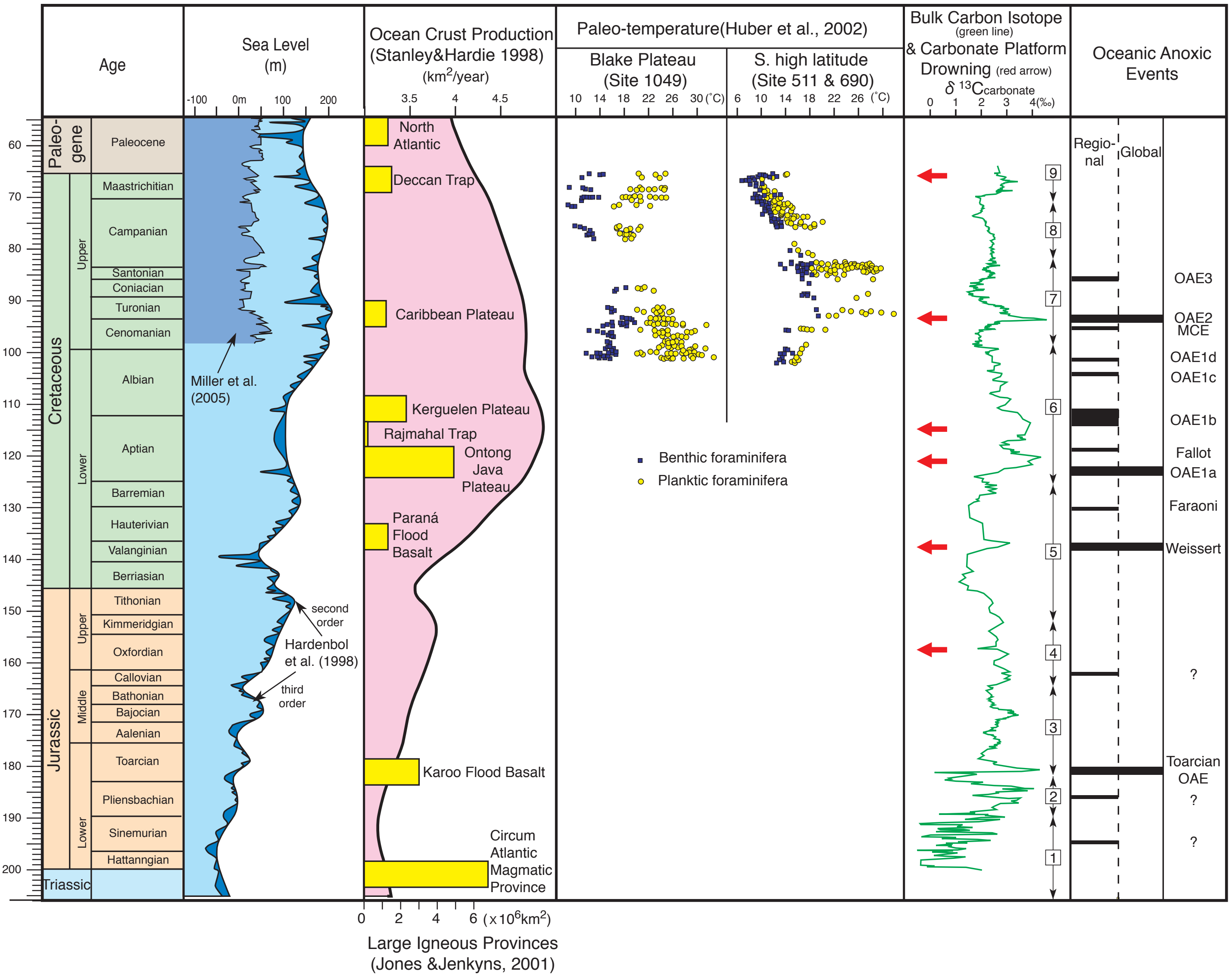
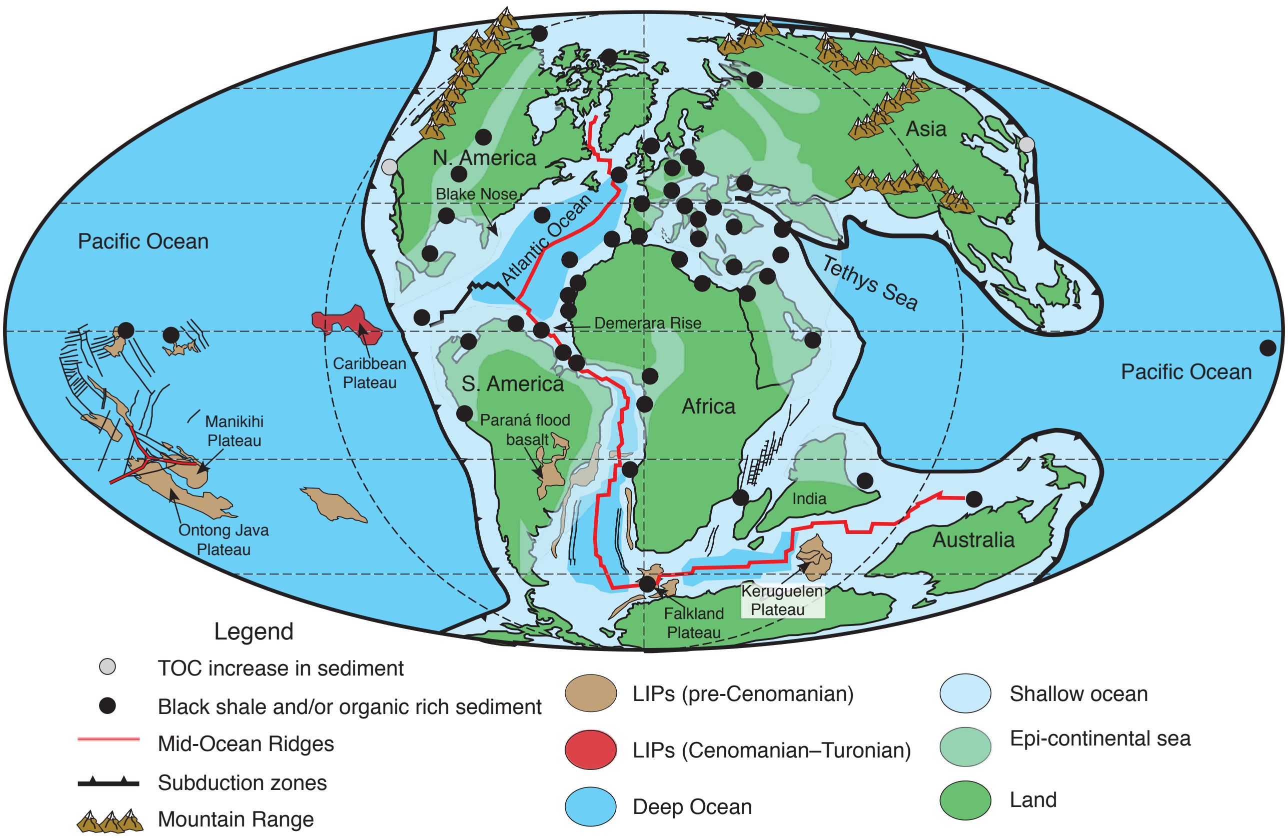


Fig. 3, Takashima et al.

Fig. 4, Takashima et al.



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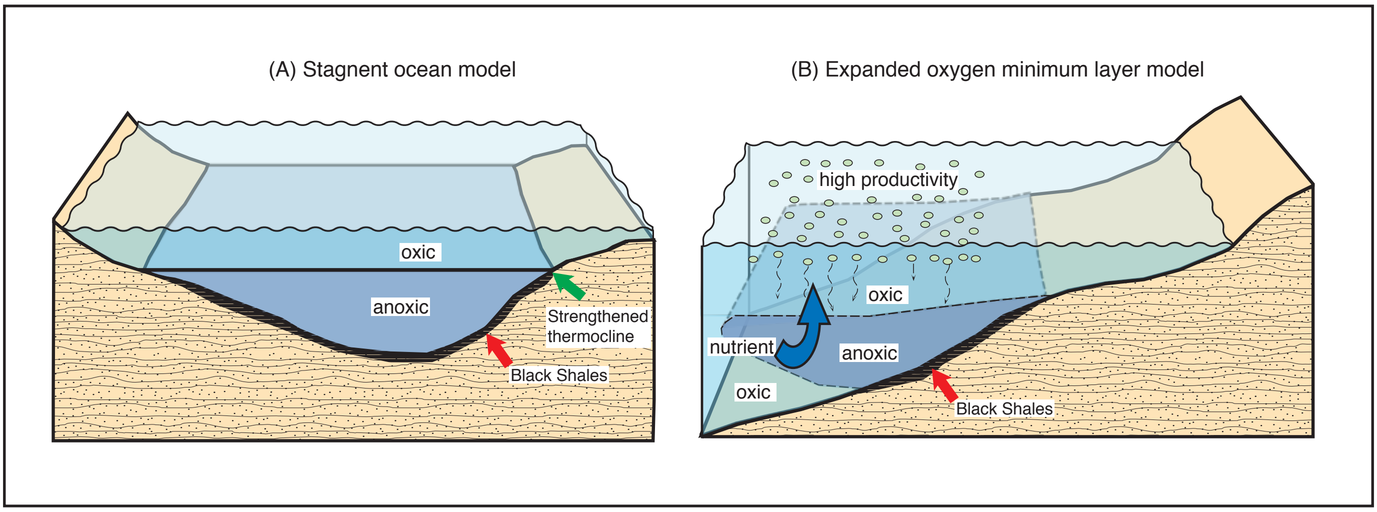

Fig. 6, Takashima et al.

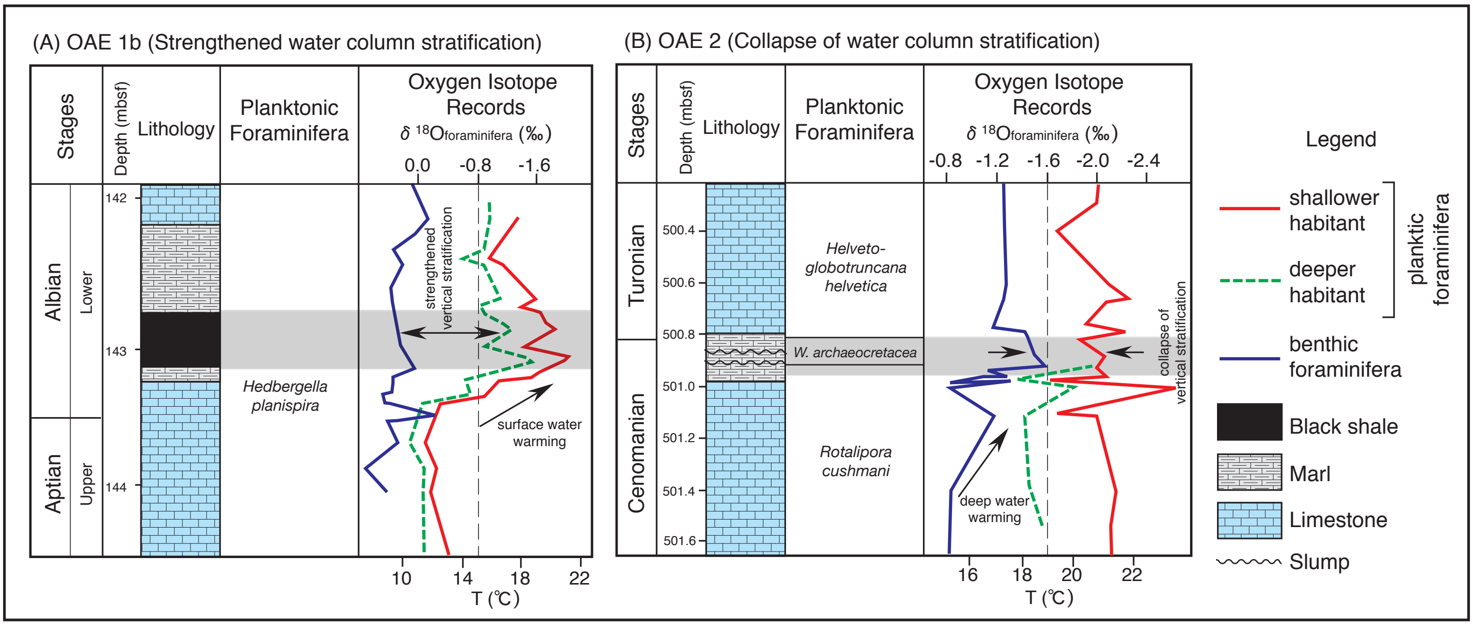

Fig. 7, Takashima et al.