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BIOTIC, PALEOCEANOGRAPHIC, AND STRATIGRAPHIC CHANGES ACROSS THE CRETACEOUS/PALEOGENE BOUNDARY

by

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A dissertation submitted to the

Graduate School-New Brunswick

Rutgers, The State University of New Jersey

In partial fulfillment of the requirements

For the degree of

Doctor of Philosophy

Graduate Program in Geological Sciences

Written under the direction of

Dr. Kenneth G. Miller

And approved by

New Brunswick, New Jersey

January, 2015

ABSTRACT OF THE DISSERTATION

Biotic, Paleoceanographic, and Stratigraphic Changes across the Cretaceous/Paleogene

Boundary

by SELEN ESMERAY-SENLET

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My dissertation investigates the sea-level changes, large perturbations in the oceanic carbon cycle, and enrichments and distribution of iridium (Ir) across the Cretaceous/Paleogene (K/Pg) boundary (~66 Ma). Biostratigraphic, lithofacies, and sequence stratigraphic analyses of the Haymana (Turkey) and Campo Pit (New Jersey, USA) sections show that the K/Pg boundary occurred during a transgressive systems tract, below a maximum flooding surface. Similar sea-level trends recorded in European and North African sections suggest that the K/Pg event occurred during a global sea-level rise.

Previous studies showed that the K/Pg mass extinction was associated with drastic perturbations in the oceanic carbon cycle, including a collapse in the carbon isotopic

 $(\delta^{13}C)$ gradient between planktonic and benthic foraminifera and a drop in bulk carbonate $\delta^{13}C$ values in deep-sea sites. The New Jersey paleoshelf recorded a ~2.5‰ $\delta^{13}C$ decrease in bulk carbonate, a ~0.8‰ $\delta^{13}C$ decrease in organic carbon, a collapse of the vertical $\delta^{13}C$ difference between planktonic and benthic foraminifera, and a significant decrease in organic carbon mass accumulation rates. A ~1.0‰ water column $\delta^{13}C$ gradient in planktonic foraminifera and a ~0.75‰ cross-shelf $\delta^{13}C$ gradient in benthic foraminifera indicates the presence of primary productivity in the early Danian. Comparison of deep Atlantic and Pacific sites shows a reduction in interbasinal $\delta^{13}C$ differences, most consistent with reduced export productivity.

Anomalous Ir concentrations occur at the K/Pg boundary in New Jersey, though the maximum Ir concentrations and the thickness of the sediments over which Ir is spread is highly variable. The shape of the Ir profiles and the maximum concentrations of Ir measured in the cores appear to be determined primarily by sedimentary and geochemical processes, especially bioturbation as also shown by a Lagrangian particle-tracking model of sediment mixing. The depth-integrated Ir inventory is very similar in the majority of the cores, indicating that the total Ir delivery at the time of the K/Pg event was spatially homogenous over this region. This provides additional evidence for redistribution of Ir after original deposition and shows that low to moderate Ir concentrations (~0.1 - 0.5 ppb) can be the product of the same Ir delivery as nearby higher peaks (~2.4 ppb).

DEDICATION

To my sunshine, my ocean

Turgay Senlet

ACKNOWLEDGEMENTS

My time at Rutgers has been the most challenging, yet the most wonderful time of my life. During this rewarding journey, there have been many people that helped my academic, professional, and personal growth and development, to whom I would like to express my deepest gratitude.

I would especially like to thank my advisor, Kenneth G. Miller, for being an inspiring, supportive, and fun advisor. His guidance and mentorship stands as the most influential factor on my advancement as a geoscientist. I admire his scientific knowledge, curiosity and endless enthusiasm.

I am thankful to James V. Browning for his endless support and help during my dissertation research. I deeply respect his professionalism and discipline. I appreciate his time for reading my dissertation diligently and valuable comments.

I feel very privileged to have the chance to work with Richard K. Olsson, the world expert on planktonic foraminifera. It was an honor to be one his students and follow up the research he has initiated.

I am thankful to James D. Wright for his time and guidance throughout my dissertation studies. I am especially grateful for the discussion we had for Chapter 4, which was enhanced by his brilliant ideas.

I would like to thank Richard Norris (Scripps Institute of Oceanography) for his valuable comments and suggestions on my thesis and his positive, encouraging attitude.

I am grateful to Robert Sherrell for allowing me to conduct Iridium measurements in his laboratory and for his valuable comments on Chapter 5. It was great to work with such a positive and helpful person. I would like to also thank Tracy M. Quan (Boone Pickens School of Geology) and Johan Vellekoop (Utrecht University) for their collaborations. I am thankful to my M.Sc. advisors Demir Altıner and Sevinç Özkan-Altıner (Middle East Technical University) for everything we shared in our academic and personal lives over the past years.

I have been very lucky to be a part of the Department of Earth and Planetary Sciences at Rutgers University. Every member of this department deserves my sincere appreciation. I would especially like to thank Carl Swisher and Jovani Reaves for their limitless help, and Gregory Mountain and Don Monteverde for the great conversations we had. I am thankful to current and former graduate students, with whom I shared wonderful times along the way: Catherine Beck, Ayda Shokoohi, Maria Makarova, Chris Lombardi, Cesar Sequeira, Dave Bord, Linda Martin, Zuhal Şeker, Sarp Karakaya, Tuçe Değirmenci, Turan İsçimen, and Deniz Küstü. They all made this educational journey enjoyable.

During my Ph.D., I had rewarding internship experiences at Chevron with the wonderful members of the Biostratigraphy Team. I would like to thank Alicia Kahn, Tom Dignes, Roger Witmer, Rome Lytton, Anthony Gary, Lawrence Febo, and Erik Anthonissen. I really enjoyed my U.S. industry experience with them and learned a lot.

My dear friends, Bahar Demirdirek, Gökçe Baykal, Gizem Karslı-Uzunbaş, Birnur Özbaş-Erdem, Gülşad Küçük, Gökhan Uzunbaş, Zafer Gezgin, have been like a family for me during the years I have been away from my home country. Thank you very much for the great years we shared together.

I would like to express my deepest gratitude to my beloved family; my mother, Asuman Esmeray, my father Mehmet Esmeray, and my sister and best friend Müge Esmeray for their endless support, prayers, and encouragements throughout this long process. Their presence and love give me a tremendous strength and I feel their warmth in my hearth every day despite of thousands of miles between us. My parents-in-law Rengin Senlet and Tevfik Senlet, my aunt Yasemin Özbek, and my dear friend Nil Özsancak deserve thanks for always believing in me and supporting me.

Last but not least, I would like to thank my husband, Turgay Senlet, for being such an encouraging, positive, and supportive partner. He is the person who deserves the most gratitude for this long journey. Every day we spend together is just joyful, happy, and peaceful despite all challenges we go through. I am grateful for all wonderful years we spent together and looking forward to the coming years that we will keep learning, growing, and laughing together.

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Chapter 1 Introduction to the Dissertation

1.1 Introduction

At the Cretaceous/Paleogene (K/Pg) boundary (~66 Ma) a global scale and abrupt mass extinction occurred, altering both marine and terrestrial ecosystems dramatically. As the most recent of the five major mass extinction events in the past 500 million years, the K/Pg mass extinction has the best fossil and geochemical record, providing an excellent case study for understanding how marine ecosystems respond to abrupt, large, and widespread perturbations. Although it has gained great attention from geoscientists, there are still controversies about the cause and the kill mechanism of the K/Pg mass extinction, as well as the nature of the recovery from the extinction. My motivation for studying the K/Pg boundary was to investigate global aspects of the K/Pg boundary with a multidisciplinary approach and elucidate outstanding questions. In my dissertation, I primarily focus on the large perturbations in the oceanic carbon cycle, stratigraphic position and the distribution of iridium, and sea-level changes across the K/Pg boundary.

In this chapter, I provide background on the K/Pg boundary event, explain the objectives of my dissertation with an outline of the chapters, and list the materials submitted or prepared to be submitted for publication.

1.2 Background on the Cretaceous/Paleogene Boundary

The global scale biotic turnover at the Cretaceous/Paleogene (K/Pg) boundary is one of the five largest mass extinctions in Earth's history (Raup and Sepkoski, 1982). A number of major fossil groups disappeared across the boundary including nonavian dinosaurs, marine and flying reptiles, ammonites and rudists; several other groups suffered major turnovers, including planktonic foraminifera, calcareous nannofossils, and land plants (Schulte et al., 2010 and the references therein).

More than thirty years ago, the discovery of anomalously high abundances of iridium (Ir) and other platinum group elements (PGEs) at several K/Pg boundary sections led to the hypothesis that an asteroid ~10 km in diameter collided with Earth causing many environmental perturbations (Alvarez et al., 1980; Smit and Hertogen, 1980). Subsequent recognition of impact related features such as shocked minerals (Bohor et al., 1987), impact spherules, and Ni-spinels (Smit and Kyte, 1984) and the discovery of the Chicxulub crater (165 – 200 km in diameter) in the Yucatan Peninsula, Mexico (Hildebrand et al., 1991; Morgan et al., 1997) substantiated the impact hypothesis. Large earthquakes, massive submarine landslides, and tsunamis from the Caribbean up to North America were caused by the impact (Bralower et al., 1998; Norris et al., 2000), and its hot ejecta caused at least regional if not global wildfires (Melosh et al., 1990; Goldin and Melosh, 2009; Robertson et al., 2013b).

The impact hypothesis has not been the only explanation for the K/Pg boundary mass extinction. It has been proposed that stratigraphic and micropaleontological data from the Gulf of Mexico show 3 impacts during the time interval spanning the boundary

with the Chicxulub impact predating the K/Pg boundary by ~300 kyr (Keller et al., 2003; 2007). This multiple impact hypothesis (Keller et al., 2003; 2007) has not gained support form other K/Pg boundary locations free from stratigraphic complexities of near-impact sites. Moreover, more than 350 K/Pg boundary sites known today show a distinct pattern of decreasing thickness of the ejecta deposits with increasing distance from the Chicxulub crater suggesting a unique source for the ejecta at the K/Pg boundary (Smit, 1999; Claeys et al., 2002; Schulte et al., 2010).

Emplacement of the large Deccan continental flood basalts in India spanning the K/Pg boundary has also been suggested as the cause of the K/Pg mass extinction. The activity of the large igneous province lasted ~800 kyr causing severe environmental effects due to the eruptive degassing of sulfur and carbon dioxide (Courtillot et al., 1986; Courtillot and Renne, 2003; Self et al., 2006; Chenet et al., 2007; 2009).

There is no consensus on the kill mechanism of the K/Pg mass extinction. Alvarez et al. (1980) suggested that impact ejecta might have caused global darkness leading a collapse in the food chain both in marine and terrestrial environments (the "impact winter" hypothesis). Toon et al. (1982) modeled ejecta related dust cloud in the atmosphere and concluded that the period of global darkness was 2 months to 1 year. Cessation of photosynthesis even for a short period of time can cause global extinction of phytoplankton taxa with life spans of weeks to months, also causing catastrophic extinction of marine autotrophs that are strongly dependent on daily photosynthetic output (Robertson et al., 2013a). Robertson et al. (2013a) suggested that two different mechanisms were responsible for marine (an impact winter) and terrestrial (a heat pulse and subsequent fires) environments (Goldin and Melosh, 2009; Robertson et al., 2013b).

Another possible kill mechanism is widespread acid rain that resulted from increased atmospheric nitric oxide, carbonic, and sulfuric acid (Toon et al., 1982; Kring, 2007). Models suggest that release of gases from the anhydrite (calcium sulfate) target rocks could have caused injection of sulfur vapor into the stratosphere, producing sulfuric acid rain (e.g., Pierazzo et al., 1998). Sulfuric acid rain might have had direct deadly biological impacts, as well as negative effects like extended darkness, global cooling, acidification, and increased weathering (Kring, 2007).

Mass extinction was catastrophic at the K/Pg boundary both in terrestrial and marine environments. Nonavian dinosaurs, rudist bivalves, ammonites and belemnites disappeared; some bird, marsupial and brachiopod species underwent drastic changes at the K/Pg boundary (D'Hondt, 2005; Schulte et al., 2010). Calcareous plankton suffered catastrophic extinctions, with the disappearance of more than 90% of the Cretaceous species of planktonic foraminifera (e.g., Olsson and Liu, 1993; Olsson et al., 1999) and calcareous nannofossils (e.g., Bernaola and Monechi, 2007). However, organic-walled, opportunistic dinoflagellates experienced minimal turnover (Brinkhuis et al., 1998), presumably because they employ a resting cyst stage as part of their life cycle. Benthic foraminifera also survived the K/Pg boundary (Culver, 2003), though the diversity of their assemblages decreased after the K/Pg boundary interpreted to be due to high environmental stress (Alegret and Thomas, 2005).

The K/Pg event is also associated with changes in primary productivity with evidence for decreased productivity in terrestrial and open ocean environments (D'Hondt, 2005). In the open ocean, surface to deep water carbon isotopic (δ^{13} C) gradients collapsed, indicating a decline in export productivity (Hsü et al., 1982; Zachos and

Arthur, 1986; Zachos et al., 1989). Benthic foraminiferal accumulation rates (Alegret and Thomas, 2007; Alegret et al., 2012) and biogenic barium (Hull and Norris, 2011) were also used as proxies for changes in productivity. In sites throughout the North Atlantic and Tethys, benthic foraminifera proxies show additional evidence for a reduction in export productivity (Alegret et al., 2001; 2004). However, proxies in the central Pacific Ocean and in some upwelling or Atlantic shelf sites show either no change or increased export productivity in the early Danian (Alegret and Thomas, 2009; Hull and Norris, 2011; Alegret et al., 2012). Full recovery to pre-extinction surface to deep water (δ^{13} C) gradients took approximately 3 million years after the K/Pg boundary (D'Hondt et al., 1998; Coxall et al., 2006).

1.3 Objectives and Outline of the Dissertation

One of the main topics I investigate in my dissertation is the change in the marine carbon cycle across the K/Pg boundary. The K/Pg mass extinction was associated with drastic perturbations in the oceanic carbon cycle as shown by a collapse in the δ^{13} C gradient between surface and deep waters recorded by planktonic and benthic foraminifera, and a significant drop in bulk carbonate δ^{13} C values (e.g., Zachos and Arthur, 1986). These changes suggest that there were drastic changes in primary productivity and/or export productivity to the deep ocean after the K/Pg boundary mass extinction (Zachos and Arthur, 1986; Stott and Kennett, 1989; Zachos et al., 1989; D'Hondt and Zachos, 1998; D'Hondt, 2005; Coxall et al., 2006). Hypotheses explaining the collapse in surface to deep ocean δ^{13} C differences after the K/Pg boundary have been

tested primarily in deep-ocean sites (Zachos and Arthur, 1986; D'Hondt and Zachos, 1998; Coxall et al., 2006; Alegret and Thomas, 2009; Alegret et al., 2012). However, most organic carbon burial occurs on continental margins (Premuzic et al., 1982), therefore, any hypothesis concerning large-scale changes in primary productivity and its export must account for changes recorded on the shelves. In Chapter 4, I test the hypotheses in a paleoshelf setting of New Jersey Coastal Plain (NJCP), USA. I present high-resolution δ^{13} C and oxygen (δ^{18} O) isotope records from multiple species of planktonic and benthic foraminifera, δ^{13} C changes of bulk and organic carbon, and the changes in the mass accumulation rates of total organic carbon between ~ 69 and 65 Ma. The updip and downdip comparison of the Ancora and Bass River coreholes, the most complete K/Pg boundary cores in the NJCP, allowed me to distinguish among the marine carbon cycle perturbation hypotheses because they recorded: 1) mixed layer to thermocline $\delta^{13}C$ gradients in planktonic foraminifera, 2) cross shelf $\delta^{13}C$ gradients in benthic foraminifera, and 3) differences in the rate of organic carbon burial across the K/Pg boundary.

In **Chapter 4**, I provide new age models for Ocean Drilling Program (ODP) Sites 1262 (Atlantic) and 1210 (Pacific) based on magnetostratigraphy and biostratigraphy, also supported by the Fe intensity counts independently. With the new age models, I evaluate the interbasinal deep-sea δ^{13} C differences between ocean basins as a proxy for changes in export productivity. Understanding the shelf response to environmental perturbations and its comparison to deep-sea responses will not only reveal important aspects of the K/Pg boundary, but also help us to understand the resilience and reorganization of carbon cycle to large perturbations, such as anthropogenic effects.

Another controversial topic related to the K/Pg boundary is sea-level change across the K/Pg boundary. It is widely accepted that the sea-level changes across the boundary were small compared to the Campanian/Maastrichtian boundary and later Paleogene sea-level fluctuations (Miller et al., 2005; Haq, 2014) and did not play an important role in the end-Cretaceous mass extinction (Hallam and Wignall, 1999). However, it is important to understand how sea level changed during the K/Pg transition to evaluate the local depositional changes of the basins, as well as global stratigraphy of the K/Pg boundary.

Sea-level studies analyzing sections from different geographic locations and paleoenvironments yielded very different interpretations about changes across the K/Pg boundary. Claeves et al. (2002) showed that in almost 100 shallow-marine sites globally, the number of sites showing regression versus no regression or transgression is almost the same. This variety in the sea-level interpretations is due to the different methodologies used and the fact that the sea-level changes across the K/Pg boundary were subtle, causing difficulties in evaluating the changes. I analyze sea-level changes across the K/Pg boundary in two sections using detailed lithostratigraphic and sequence stratigraphic approach. In Chapter 2, I delineate the K/Pg boundary in a carbonate – mixed siliciclastic section in the Haymana Basin, Central Anatolia, Turkey using planktonic foraminiferal biostratigraphy, determine the depositional history of the basin with a comprehensive microfacies analysis, and place the section within a sequence stratigraphic framework. In Chapter 3, I analyze the shoreface siliciclastic deposits at Campo Pit outcrop, New Jersey, USA with detailed lithostratigraphic and sequence stratigraphic analyses to test the two possible sea-level scenarios shown in the previous works in New

Jersey. I construct relative sea-level curves for both Haymana and Campo Pit sections, and compare them with sections elsewhere to understand the global sea-level changes across the K/Pg boundary (**Chapter 2**). In **Chapter 3**, I also provide a detailed review of sea-level studies across the K/Pg boundary from several sections in South America, North America, Europe, and North Africa.

Another major topic I investigate in my dissertation is the anomalous Ir concentration associated with the K/Pg boundary. An Ir anomaly has been found in more than 85 K/Pg boundary sites globally (Smit, 1999; Claeys et al., 2002) after its first discovery by Alvarez et al. (1980). Because Ir is highly depleted in Earth's crust, an anomalously high Ir concentration at the boundary is the basis for the extraterrestrial impact hypothesis (Alvarez et al., 1980; Smit and Hertogen, 1980). In Chapter 5, I investigate the Ir enrichments across the K/Pg boundary in 9 cores in the NJCP deposited in shallow marine environments at intermediate distances (~2500 km) from the Chicxulub crater. Outside the Gulf of Mexico, i.e., in the intermediate and distal sites from the Chicxulub crater, there is no correlation between the Ir concentration and distance from the impact site (Claeys et al., 2002). Original deposition of Ir can be altered due to remobilization by sedimentary processes including bioturbation, geochemical remobilization, or mass transport that can account for the difference in Ir concentrations even in geographically close sites (Sawlowicz, 1993; Claeys et al., 2002). I conducted Ir measurements and analyzed Ir anomaly profiles to understand how the remobilization of Ir by sedimentary processes affects the shape of Ir profiles. I used a sediment mixing model (Hull et al., 2011) to explain the shape of the Ir profiles. Furthermore, to understand whether the low to moderate Ir peaks with respect to higher anomalies

recorded can be attributed to bioturbation, geochemical remobilization, or redeposition, I analyzed the integrated Ir signal and calculated the total delivery of Ir.

Finally, in **Chapter 6**, I provide a summary of the conclusions of my dissertation research along with directions for future work.

1.4 Material Prepared for Publication

The chapters of this dissertation are in preparation or have been submitted for publication.

Chapter 2 was submitted for publication and is currently under review in the *Journal of Sedimentary Research* as "Planktonic foraminiferal biostratigraphy, microfacies analysis, sequence stratigraphy, and sea-level changes across the Cretaceous/Paleogene Boundary in the Haymana Basin, Central Anatolia, Turkey". This paper was constructed with the development and reinterpretation of the data that I collected during my M.Sc. thesis (Esmeray, 2008) and its comparison to new data from NJCP cores. I am the principal investigator and author in this paper and my co-authors are Sevinç Özkan-Altıner, Demir Altıner, and Kenneth G. Miller.

Chapter 3 is in preparation for publication as "Stratigraphy and sea-level changes across the Cretaceous/Paleogene Boundary in the Campo Pit Section, New Jersey, USA". This study is in collaboration with co-authors Kenneth G. Miller and James V. Browning, and I am the primary investigator and author of this work. Chapter 4 was submitted to *Paleoceanography* for publication and currently under review as "Evidence for the Living Ocean following the Cretaceous/Paleogene mass extinction". I co-author this paper with James D. Wright, Richard K. Olsson, Kenneth G. Miller, James V. Browning, and Tracy Quan, and am the primary investigator and author.

Chapter 5 is in preparation for publication as "Iridium profiles and delivery across the Cretaceous/Paleogene boundary in the New Jersey Coastal Plain". This study is in collaboration with Kenneth G. Miller, Robert M. Sherrell, Turgay Senlet, and Johan Vellekoop, and I am the primary investigator and author of this work.

- Alegret, L., and Thomas, E., 2005, Cretaceous/Paleogene boundary bathyal paleoenvironments in the central North Pacific (DSDP Site 465), the Northwestern Atlantic (ODP Site 1049), the Gulf of Mexico and the Tethys: The benthic foraminiferal record: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 224, p. 53–82, doi: 10.1016/j.palaeo.2005.03.031.
- Alegret, L., and Thomas, E., 2007, Deep-Sea environments across the Cretaceous/Paleogene boundary in the eastern South Atlantic Ocean (ODP Leg 208, Walvis Ridge): Marine Micropaleontology, v. 64, p. 1–17.
- Alegret, L., and Thomas, E., 2009, Food supply to the seafloor in the Pacific Ocean after the Cretaceous/Paleogene boundary event: Marine Micropaleontology, v. 73, p. 105– 116, doi: 10.1016/j.marmicro.2009.07.005.
- Alegret, L., Kaminski, M.A., and Molina, E., 2004, Paleoenvironmental recovery after the Cretaceous/Paleogene boundary crisis: evidence from the Marine Bidart section (SW France): Palaios, v. 19, p. 574–586.
- Alegret, L., Molina, E., and Thomas, E., 2001, Benthic foraminifera at the Cretaceous-Tertiary boundary around the Gulf of Mexico: Geology, v. 29, p. 891–894.
- Alegret, L., Thomas, E., and Lohmann, K.C., 2012, End-Cretaceous marine mass extinction not caused by productivity collapse: Proceedings of the National Academy of Sciences of the United States of America, v. 109, p. 728–732, doi: 10.1073/pnas.1110601109/-/DCSupplemental/ST01.doc.
- Alvarez, L., Alvarez, W., Asaro, F., and Michel, H., 1980, Extraterrestrial cause for the Cretaceous-Tertiary extinction: Science, v. 208, p. 1095–1108.
- Bernaola, G., and Monechi, S., 2007, Calcareous nannofossil extinction and survivorship across the Cretaceous–Paleogene boundary at Walvis Ridge (ODP Hole 1262C, South Atlantic Ocean): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 132–156, doi: 10.1016/j.palaeo.2007.02.045.
- Bohor, B., Modreski, P., and Foord, E., 1987, Shocked quartz in the Cretaceous-Tertiary boundary clays: evidence for a global distribution: Science, v. 236, p. 705–709.
- Bralower, T.J., Paull, C.K., and Leckie, R.M., 1998, The Cretaceous-Tertiary boundary cocktail: Chicxulub impact triggers margin collapse and extensive sediment gravity flows: Geology, v. 26, p. 331–334.
- Brinkhuis, H., Bujak, J.P., Smit, J., Versteegh, G., and Visscher, H., 1998, Dinoflagellate-based sea surface temperature reconstructions across the Cretaceous– Tertiary boundary: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 141, p. 67–83.
- Chenet, A., Quidelleur, X., Fluteau, F., Courtillot, V., and Bajpai, S., 2007, 40K-40Ar dating of the Main Deccan large igneous province: Further evidence of KTB age and short duration: Earth and Planetary Science Letters, v. 263, p. 1–15.
- Chenet, A.-L., Courtillot, V., Fluteau, F., Gérard, M., Quidelleur, X., Khadri, S.F.R., Subbarao, K.V., and Thordarson, T., 2009, Determination of rapid Deccan eruptions across the Cretaceous-Tertiary boundary using paleomagnetic secular variation: 2. Constraints from analysis of eight new sections and synthesis for a 3500-m-thick composite section: Journal of Geophysical Research, v. 114, p. B06103, doi: 10.1029/2008JB005644.
- Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of Chicxulub ejecta at the Cretaceous-Tertiary boundary (C. Koeberl & K. G. MacLeod, Eds.): Catastrophic Eevents and Mass extinctions: Impacts and Beyond: Geological Society of America Special Paper, v. 356, p. 55–68.
- Courtillot, V., and Renne, P., 2003, On the ages of flood basalt events: Earth and Planetary Science Letters, v. 335, p. 113–140.
- Courtillot, V., Besse, J., and Vandamme, D., 1986, Deccan flood basalts at the Cretaceous/Tertiary boundary?: Earth and Planetary Science Letters, v. 80, p. 361–374.
- Coxall, H.K., D'Hondt, S., and Zachos, J.C., 2006, Pelagic evolution and environmental recovery after the Cretaceous-Paleogene mass extinction: Geology, v. 34, p. 297– 300, doi: 10.1130/G21702.1.
- Culver, S., 2003, Benthic foraminifera across the Cretaceous-Tertiary (K-T) boundary: a review: Marine Micropaleontology, v. 47, p. 177–226.
- D'Hondt, S., 2005, Consequences of the Cretaceous/Paleogene mass extinction for marine ecosystems: Annual Review of Ecology, Evolution, and Systematics, v. 36, p. 295–317.
- D'Hondt, S., and Zachos, J.C., 1998, Cretaceous foraminifera and the evolutionary history of planktic photosymbiosis: Paleobiology, v. 24, p. 512–523.
- D'Hondt, S., Donaghay, P., Zachos, J.C., Luttenberg, D., and Lindinger, M., 1998, Organic carbon fluxes and ecological recovery from the Cretaceous-Tertiary mass extinction: Science, v. 282, p. 276–279, doi: 10.1126/science.282.5387.276.

- Esmeray, S., 2008, Cretaceous/Paleogene Boundary in the Haymana Basin, Central Anatolia, Turkey: Micropaleontological, Mineralogical and Sequence Stratigraphic Approach: M.Sc. Thesis, Middle East Technical University. Advisors: Demir Altıner and Sevinç Özkan-Altıner, 290 p.
- Goldin, T., and Melosh, H., 2009, Self-shielding of thermal radiation by Chicxulub impact ejecta: Firestorm or fizzle?: Geology, v. 37, p. 1135–1138.
- Hallam, A., and Wignall, P., 1999, Mass extinctions and sea-level changes: Earth-Science Reviews, v. 48, p. 217–250.
- Haq, B.U., 2014, Cretaceous eustasy revisited: Global and Planetary Change, v. 113, p. 44–58, doi: 10.1016/j.gloplacha.2013.12.007.
- Hildebrand, A., Penfield, G., Kring, D., Pilkington, M., and Camargo, Z., 1991, Chicxulub crater: a possible Cretaceous/Tertiary boundary impact crater on the Yucatan Peninsula, Mexico: Geology, v. 19, p. 867–871.
- Hsü, K., He, Q., McKenzie, J., Weissert, H., Perch-Nielsen, K., Oberhansli, H., Kelts, K., LaBrecque, J., Tauxe, L., and Krahenbuhl, U., 1982, Mass mortality and its environmental and evolutionary consequences: Science, v. 216, p. 249–256.
- Hull, P.M., and Norris, R.D., 2011, Diverse patterns of ocean export productivity change across the Cretaceous-Paleogene boundary: New insights from biogenic barium: Paleoceanography, v. 26, p. PA3205, doi: 10.1029/2010PA002082.
- Hull, P.M., Franks, P.J.S., and Norris, R.D., 2011, Earth and Planetary Science Letters: Earth and Planetary Science Letters, v. 301, p. 98–106, doi: 10.1016/j.epsl.2010.10.031.
- Keller, G., Adatte, T., Berner, Z., Harting, M., Baum, G.R., Prauss, M., Tantawy, A., and Stuben, D., 2007, Chicxulub impact predates KT boundary: New evidence from Brazos, Texas: Earth and Planetary Science Letters, v. 255, p. 339–356.
- Keller, G., Stinnesbeck, W., Adatte, T., and Stuben, D., 2003, Multiple impacts across the Cretaceous-Tertiary boundary: Earth-Science Reviews, v. 62, p. 327–363.
- Kring, D., 2007, The Chicxulub impact event and its environmental consequences at the Cretaceous-Tertiary boundary: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 4–21.
- Melosh, H.J., Schneider, N.M., Zahnle, K.J., and Latham, D., 1990, Ignition of global wildfires at the Cretaceous/Tertiary boundary: Nature, v. 343, p. 251–254.
- Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., and Pekar, S., 2005, The Phanerozoic record of global sea-level change: Science, v. 310, p. 1293–1298, doi: 10.1126/science.1116412.

- Morgan, J., Warner, M., Brittan, J., Buffler, R., Camargo, A., Christeson, G., Denton, P., Hildebrand, A., Hobbs, R., and Macintyre, H., 1997, Size and morphology of the Chicxulub impact crater: Nature, v. 390, p. 472–476.
- Norris, R.D., Firth, J., Blusztajn, J., and Ravizza, G., 2000, Mass failure of the North Atlantic margin triggered by the Cretaceous-Paleogene bolide impact: Geology, v. 28, p. 1119–1122.
- Olsson, R.K., and Liu, C., 1993, Controversies on the placement of Cretaceous-Paleogene boundary and the K/P mass extinction of planktonic foraminifera: Palaios, v. 8, p. 127–139.
- Olsson, R.K., Hemleben, C., Berggren, W.A., Huber, B.T., Editors and Members of the Paleogene Planktonic Foraminifera Working Group, 1999, Atlas of Paleocene Planktonic Foraminifera: Smithsonian Contributions to Paleobiology, v. 85, p. 252.
- Pierazzo, E., Kring, D.A., and Melosh, H.J., 1998, Hydrocode simulation of the Chicxulub impact event and the production of climatically active gases: Journal of Geophysical Research: Planets (1991–2012), v. 103, p. 28607–28625.
- Premuzic, E.T., Benkovitz, C.M., Gaffney, J.S., and Walsh, J.J., 1982, The nature and distribution of organic matter in the surface sediments of world oceans and seas: Organic Geochemistry, v. 4, p. 63–77, doi: 10.1016/0146-6380(82)90009-2.
- Raup, D., and Sepkoski, J., Jr., 1982, Mass extinctions in the marine fossil record: Science, v. 215, p. 1501–1503.
- Robertson, D.S., Lewis, W.M., Sheehan, P.M., and Toon, O.B., 2013a, K-Pg extinction patterns in marine and freshwater environments: The impact winter model: Journal of Geophysical Research: Biogeosciences, p. n/a–n/a, doi: 10.1002/jgrg.20086.
- Robertson, D.S., Lewis, W.M., Sheehan, P.M., and Toon, O.B., 2013b, K-Pg extinction: Reevaluation of the heat-fire hypothesis: Journal of Geophysical Research: Biogeosciences, v. 118, p. 329–336, doi: 10.1002/jgrg.20018.
- Sawlowicz, Z., 1993, Iridium and other platinum-group elements as geochemical markers in sedimentary environments: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 104, p. 253–270.
- Schulte, P., Alegret, L., Arenillas, I., Arz, J.A., Barton, P., Bown, P., Bralower, T.J., Christeson, G., Claeys, P., and Cockell, C. et al., 2010, The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary: Science, v. 327, p. 1214–1218.
- Self, S., Widdowson, M., and Thordarson, T., 2006, Volatile fluxes during flood basalt eruptions and potential effects on the global environment: A Deccan perspective: Earth and Planetary Science Letters, v. 248, p. 518–532.

- Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Sciences, v. 27, p. 75–113.
- Smit, J., and Hertogen, J., 1980, An extraterrestrial event at the Cretaceous–Tertiary boundary: Nature, v. 285, p. 198–200.
- Smit, J., and Kyte, F.T., 1984, Siderophile-rich magnetic spheroids from the Cretaceous– Tertiary boundary in Umbria, Italy: Nature, v. 310, p. 403–405.
- Stott, L.D., and Kennett, J.P., 1989, New constraints on early Tertiary palaeoproductivity from carbon isotopes in foraminifera: Nature, v. 342, p. 526–529.
- Toon, O.B., Pollack, J.B., Ackerman, T.P., Turco, R.P., McKay, C.P., and Liu, M.S., 1982, Evolution of an impact-generated dust cloud and its effects on the atmosphere: Special Papers-Geological Society of America, v. 190, p. 187–200.
- Zachos, J.C., and Arthur, M., 1986, Paleoceanography of the Cretaceous/Tertiary boundary event: inferences from stable isotopic and other data: Paleoceanography, v. 1, p. 5–26.
- Zachos, J.C., Arthur, M., and Dean, W., 1989, Geochemical evidence for suppression of pelagic marine productivity at the Cretaceous/Tertiary boundary: Nature, v. 337, p. 61–64.

Chapter 2

Planktonic Foraminiferal Biostratigraphy, Microfacies Analysis, Sequence Stratigraphy, and Sea-Level Changes across the Cretaceous/Paleogene Boundary in the Haymana Basin, Central Anatolia, Turkey

2.1 Abstract

The Cretaceous/Paleogene (K/Pg) boundary in the Haymana Basin, Central Anatolia, Turkey was delineated using planktonic foraminiferal biostratigraphy, microfacies analysis, and sequence stratigraphy. A ~29 m outcrop consisting of limestone and marl was measured and four planktonic foraminiferal biozones were identified spanning the boundary. Planktonic foraminiferal extinction across the K/Pg boundary is catastrophic and abrupt. The extinction level is overlain by a unit (Zone P0) showing an increase in echinoid fecal pellets and authigenic clay minerals such as glauconite suggesting low sedimentation rates in the early Danian. Ten microfacies types were identified indicating inner ramp to basinal paleoenvironments based on the sedimentological characteristics and micro- and macrofossil assemblages. Maastrichtian carbonates contain large benthic foraminifera, calcareous red algae, bryozoans, fragments of echinoderms and mollusks, and planktonic foraminifera. Overlying Maastrichtian -Danian silty marls and silty limestones have common planktonic and benthic foraminifera. Progradation of carbonates into the basin took place during the highstand systems tract and deposition of silty marl succession occurred during the transgressive systems tract. The K/Pg boundary is in the upper part of the transgressive systems tract, below a maximum flooding surface. Sequence stratigraphic analysis of a second section, Campo Pit, New Jersey, USA showed that the K/Pg boundary occurs within a transgressive systems tract in New Jersey as well, suggesting a global sea-level rise across the K/Pg boundary.

2.2 Introduction

The Cretaceous/Paleogene (K/Pg) boundary denotes one of the five largest mass extinctions in Earth history (Raup and Sepkoski, 1982), where a number of major groups disappeared, such as nonavian dinosaurs, marine and flying reptiles, ammonites, and rudists (Schulte et al., 2010 and the references therein). In the marine realm, the turnover of calcareous nannofossils and planktonic foraminifera was especially remarkable (e.g., Bernaola and Monechi, 2007; Olsson and Liu, 1993). More than 90% of Cretaceous planktonic foraminifera became extinct at the K/Pg boundary with three survivor species, *Guembelitria cretacea, Hedbergella holmdelensis* and *Hedbergella monmouthensis*, giving rise to the evolution of all Paleogene planktonic foraminifera (Olsson and Liu, 1993; Olsson et al., 1999).

The global stratotype section of the K/Pg boundary at El-Kef, Tunisia is characterized by a boundary clay, an Ir anomaly, and an increase in Ni-rich spinels (Ben Abdelkader et al., 1997). The presence of microtektites and shocked quartz is another important characteristic of K/Pg boundary sections in various locations in the world (Smit, 1999). However, none of these lithological or geochemical criteria alone is enough to identify the boundary and paleontological evidence is needed. The extinction level of Cretaceous planktonic foraminifera is a global horizon that is used for the correlation of the K/Pg boundary in marine sections worldwide (e.g., Luterbacher and Premoli Silva, 1964; Olsson and Liu, 1993; Olsson et al., 1999).

The objective of this study is to delineate the K/Pg boundary in the Haymana Basin (Fig. 2.1), a sedimentary basin in Central Anatolia, Turkey, using planktonic foraminiferal biostratigraphy, determine the depositional history of the basin with a comprehensive microfacies analysis, and place the section within a sequence stratigraphic framework. We also aim to develop a relative sea-level curve for the Haymana Basin and compare it with sections elsewhere, including new data from New Jersey, USA. In spite of numerous studies, there is still no consensus on the global sea-level changes across the K/Pg boundary (e.g., Claeys et al., 2002). Detailed microfacies analysis of the Haymana Basin and the stratigraphic comparison to sections in other geographic locations will help us to understand the global sea-level changes across the boundary. Understanding the depositional history and paleontologic characteristics of the Haymana Basin will be useful in future exploration of similar basins, and will help us to elucidate global aspects of the K/Pg boundary.

2.3 Geological Setting

The Haymana Basin, situated in Central Anatolia, is a fore-arc basin formed during the Late Cretaceous to late Eocene on the oceanic crust of the northern branch of Neo-Tethys (i.e., Izmir - Ankara suture zone, Fig. 2.1A). It was formed by the convergence and collision of the Eurasian plate to the north, the Gondwana plate to the south, and with the adjacent Sakarya continent (Fourquin, 1975; Şengör and Yılmaz, 1981; Görür et al., 1984; Koçyiğit, 1991; Koçyiğit et al., 1998). The basin is surrounded by the Sakarya continent to the north-northwest, the metamorphic Kırşehir massif to the east, and the Menderes-Taurid block to the south (Fig. 2.1A).

The existence of the calc-alkaline Galatean volcanics in the Pontides during the Paleogene and the ophiolithic basement show that the basin was developed on an accretionary wedge, which was active from the Late Cretaceous to the late Eocene (Şengör and Yılmaz, 1981; Görür et al., 1984; Koçyiğit, 1991). The arc activity in the Sakarya continent shows that the subduction was towards to the north (Fourquin, 1975; Şengör and Yılmaz, 1981) and the deformation in the basin continued until the late Pliocene (Koçyiğit, 1991).

The Haymana Basin consists of highly deformed and continuous Maastrichtian to upper Eocene sedimentary fill that is greater than 5 km thick (Ünalan et al., 1976; Görür, 1981). Deposition was dominated mostly by deep-marine flysch. The center of the basin mainly consists of turbidites, whereas towards the margins there are platform carbonates, continental red beds, and reefal build-ups with some volcanic intercalations (Yüksel, 1970; Görür, 1981; Koçyiğit and Lünel, 1987; Çiner, 1992). Our measured section represents the K/Pg transition within the Beyobasi Formation, which overlies shales of the Haymana Formation and underlies the limestones of Çaldağ Formation (Ünalan et al., 1976). It consists of shelf to basin limestones and marls and is located approximately 10 km southwest of the town of Haymana, which is situated 70 km southwest of Ankara, Turkey (Fig. 2.1B). Previous biostratigraphic studies indicated that the K/Pg boundary is conformable in the study area (Özkan-Altıner and Özcan, 1999).

2.4 Methods and Materials

A 29.41-m-thick section was measured in the Haymana Basin consisting of marls and limestones and 90 samples were collected throughout the section for biostratigraphic, lithostratigrapic, and microfacies analysis. Sixty samples (A1–A60) were collected from the section with a sampling interval of 10 –100 cm (Fig. 2.2). In order to sample the lowest occurrences of the earliest Danian forms and to be able to construct a detailed planktonic foraminiferal biozonation across the boundary, a 2.34-m-thick interval of the measured section including the boundary (between samples A48–A54) was trenched, resampled, and 30 additional samples (B1–B30) were collected with a sampling interval of 2–10 cm (Fig. 2.3).

Because the samples are indurated, several washing techniques were applied for the taxonomic and biostratigraphic studies. Neither conventional planktonic foraminifera washing techniques such as soaking the samples with hydrogen peroxide or acetic acid with different dilutions and durations, nor methods such as Knitter (1979) and Lirer (2000) yielded satisfactory results for limestone samples. Therefore, we developed a specific washing technique in order to obtain foraminiferal specimens. Approximately 25 g of sample was broken down into 2-5 mm³ pieces and soaked in a 250 ml solution of 50% diluted hydrogen peroxide for marl samples and half composed of 50% diluted hydrogen peroxide and half composed of 50% diluted acetic acid for limestone samples. The mixture was shaken with a magnetic splitter for 45 - 60 minutes, and then the samples were washed under running water through 425, 250, 125, and 63 µm sieves. In order to facilitate identification, picked samples were ultrasonically cleaned for 4 - 7 minutes. Thin sections of the samples were prepared in the Department of Geological Engineering, Middle East Technical University for mineralogical and microfacies analysis.

The taxonomic determinations of planktonic foraminifera are based on Robaszynski et al. (1984), Nederbragt (1991), and Premoli-Silva and Verga (2004) for the Late Cretaceous species and Olsson et al. (1999) for the early Paleogene species. Despite the poor to moderate preservation of the species, key morphological characteristics such as coiling mode, peripheral shape, arrangement and number of the chambers, presence or absence of keels, and sutural properties enabled us to construct a reliable biostratigraphy.

Carbonate rocks in the section were identified based on the Dunham (1962) and Embry and Klovan (1971) classifications of carbonate rocks For the mixed siliciclasticcarbonate rocks, the principles proposed in the Mount (1985) classification were utilized. Major fossil groups were determined using the descriptions and photographs in Flügel (2004), Scholle and Ulmer-Scholle (2003), and Horowitz and Potter (1971). For the determination of the depositional environments, microfacies models of Wilson (1975) and Flügel (2004) were used.

Microfacies analysis was performed by examining mineralogical components, macro-, and microfossil assemblages, and texture of the samples observed in thin sections and outcrop. Rock forming fossil groups like gastropods, pelecypods, echionoderms, bryozoans, hyaline large and smaller benthic foraminifera, agglutinated benthic foraminifera, and calcareous red and green algae were identified without taxonomic details in thin sections in order to determine the environment of deposition. Furthermore, echinoid fecal pellets from the $63 - 250 \mu m$ size fraction were counted using microsplitter and number of fecal pellets per gram of sample was determined.

A 6 m-thick section was measured at Campo Pit, Perrineville, New Jersey and 15 samples were collected for litho-, and sequence stratigraphic analyses. Samples were disintegrated using Calgon solution (5.5 g of sodium metaphosphate per 4 liters of water) and washed with tap water through a 63- μ m sieve. Dry weights before and after washing were noted to calculate the sand/mud ratio of the samples. After drying in an oven at 40°C, samples were sieved through 250, 150, 125, and 63 μ m sieves and each size fraction was examined separately for mineralogical analysis. Semi-quantitative cumulative percentage analysis of minerals was conducted by visual estimation for medium (> 250 μ m) and fine (< 250 μ m) sand fractions separately.

2.5 Results

2.5.1 Lithostratigraphy

The lowermost part of the section (samples A1–A23) consists of yellowish, highly fractured, bioclastic carbonates rich in hyaline and agglutinated benthic foraminifera, calcareous red algae, bryozoans, mollusks, and echinoderm fragments with minor amounts of Late Cretaceous planktonic foraminifera (Fig. 2.2). Above this ~7-m-thick

interval of packstone, grainstone, floatstone, and wackestone-packstone alternation, wackestone with planktonic organisms including planktonic foraminifera and calcispheres was deposited in a deeper water setting (samples A24–A26). Samples A27 and A30 (~ 2 m) are from a quartz-rich silty limestone with Cretaceous ammonites, large benthic foraminifera, calcareous red algae, bryozoans, mollusks and echinoderm fragments. The overlying unit is an 18-m-thick, dark gray to green silty marl that spans the K/Pg boundary (A30–A52/B17). The number of large benthic foraminifera, algae, mollusks, and echinoderm fragments decreases upsection and the number of planktonic foraminifera, agglutinated, and hyaline benthic foraminifera increases within the silty marl succession. The K/Pg boundary is placed between samples A50–A51 (B13–B14) upsection, in the transition from the silty marls to the silty marl-silty limestone alternations (Figs. 2.2, 2.3). Just above the boundary (B14, B15/A51), a remarkable increase in green clay minerals including glauconite, smectite, chlorite, and echinoid fecal pellets occurs (Figs. 2.2, 2.3). In the Danian, silty limestone-silty marl alternations occur above sample A52/B17 up to sample A60 (~3 m). Limestones in the uppermost part of the section contain silt and clay and are rich in planktonic and benthic foraminifera.

2.5.2 Biostratigraphy

The biostratigraphy of the measured section is based on planktonic foraminiferal datum events (Table 2.1). Fourteen genera and 47 species indicative of the Maastrichtian and 10 genera and 15 species indicative of the Danian were identified. Due to the

rareness and absence of coiled index taxa in the uppermost Maastrichtian, we used the Tethyan biozonation of Robaszynski and Caron (1995) and Robaszynski (1998) that is based on biserial (Heterohelicidae) families. This standard heterohelicid biozonation, although based on Mediterranean sections, is globally applicable and has been used by many authors in different K/Pg boundary localities where globotruncanids are absent or rare (e.g., Li and Keller, 1998; Obaidalla, 2005; Georgescu and Abramovich, 2008; Huber et al., 2008). For the lowermost Paleogene, we used the standard zonal scheme of Olsson et al. (1999) and Berggren and Pearson (2005).

We recognize four upper Maastrichtian to lower Danian biozones: the Pseudoguembelina hariaensis Zone, Zone P0 (Guembelitria cretacea), Zone Pa (Parvularugoglobigerina eugubina), and Zone P1a (Parasubbotina pseudobulloides) from older to younger (Table 2.1). In the lowermost part of the section (samples A1– A44), below the *P. hariaensis* Zone, carbonates and marls representing a shallow-water paleoenvironment have very few planktonic foraminifera to assign biozones. However, some upper Maastrichtian index taxa exist in this unzoned interval such as Planoglobulina acervulinoides (lowest occurrence in sample A36) and Racemiguembelina fructicosa (lowest occurrence in sample A41), whose lowest occurrences in the section seem to be facies controlled (Table 2.1). In addition, there is a \sim 2-m-thick covered interval in the outcrop below sample A40 where we could not sample (Fig. 2.2). This may have also caused us to miss the global first appearance datum of R. fructicosa (Fig. 2.4).

Pseudoguembelina hariaensis Partial-range Zone: The *P. hariaensis* Partialrange Zone (Robaszynski and Caron, 1995) is the interval from the first appearance

datum (FAD) of the nominate species to the extinction horizon of the Cretaceous planktonic foraminifera at the K/Pg boundary, such as *Globotruncana aegyptiaca*, Globotruncanita Globotruncana arca. stuarti, Globotruncanita stuartiformis, Globigerinelloides multispinus, Globigerinelloides subcarinatus, Racemiguembelina Rugoglobigerina hexacamerata. Rugoglobigerina fructicosa, macrocephala, Rugoglobigerina rugosa, and Pseudotextularia elegans. The P. hariaensis Zone in the measured section is ~ 5 m thick (samples A45–A50/B13) and lies within the silty marks rich in quartz and iron-oxide (Table 2.1, Fig. 2.4).

Zone P0 (*Guembelitria cretacea* **Partial-Range Zone**): This zone is the lowermost Danian zone and defined by the partial range of the nominate taxon between the LAD of Cretaceous genera like *Globotruncana*, *Globotruncanita*, *Contusotruncana*, *Globotruncanella*, *Rugoglobigerina*, *Racemiguembelina*, *Planoglobulina*, and the FAD of *Parvularugoglobigerina eugubina* (Olsson et al., 1999; Berggren and Pearson, 2005; Wade et al., 2011). It is represented in the section from samples B14 to B15. Early-evolved Danian forms observed in this interval are *Eoglobigerina eobulloides*, *Globanomalina archeocompressa*, *Zeauvigerina waiparaensis*, and *Globanomalina* aff. *planocompressa* (Table 2.1, Fig. 2.5). *Guembelitria cretacea* is also abundant. In this zone except these forms, which typically appear in P0, we also identified biserial forms like *Woodringina claytonensis* and *Chiloguembelina morsei*. According to the Atlas of Paleocene Planktonic Foraminifera (Olsson et al., 1999) biserial species only evolve in Zone Pα, but several other authors observed them in Zone P0 (Keller, 1988; Koutsoukos, 1996; Luciani, 2002).

We found G. aff. planocompressa in Zone P0 starting from the first Danian sample. It is a low trochospiral form with 3–3.5 chambers resembling the species first described by Luterbacher and Premoli-Silva (1964) as *Globigerina minutula* in the Gubbio section, Italy together with "Globigerina" eugubina. Later, Globigerina minutula was used as a zonal species for the early Danian after the G. cretacea Zone and before "Globigerina" fringa and "Globigerina" eugubina Zones (Smit, 1982). After that, the species has been described as an early Danian form by many authors with various different names (Keller, 1988; Huber, 1991; Liu and Olsson, 1992; Luciani, 1997; Obaidalla, 2005; Arenillas et al., 2006). Globigerina minutula was not included in the Atlas of Paleocene Planktonic Foraminifera (Olsson et al., 1999) because the type specimen was completely recrystallized and the wall structure is unknown. The species we described in the Haymana Basin from P0 to lower P1a mostly resembles Globanomalina planocompressa with its low, umbilical-extraumbilical aperture bordered by a narrow lip, axial compression, and smooth wall (Fig. 2.5.1a, 2.5.1b). Since G. planocompressa has 5 chambers in the last whorl, this form should be an earlier morphological variant of the species and shall be named as G. aff. *planocompressa*.

Besides the first Paleocene forms, some Cretaceous forms are also observed in the first Danian sample (B14) such as *Globotruncana arca*, *Globotruncana orientalis*, *Globotruncanita pettersi*, *Globotruncanita stuartiformis*, *Rugoglobigerina hexacamerata*, *Rugoglobigerina pennyi*, *Rugoglobigerina rugosa*, *Globigerinelloides prairiehillensis*, *Heterohelix globulosa*, *Heterohelix punctulata*, *Pseudotextularia elegans*, *Pseudotextularia nuttalli*, *Planoglobulina acervulinoides*, *Pseudoguembelina hariaensis*, *Laeviheterohelix glabrans*, *H. holmdelensis*, *H. monmouthensis*, and *G. cretacea*. All these Cretaceous forms occur only within the first 5 cm of the Danian strata, their relative abundance is lower than the occurrences in the Maastrichtian, and they do not exist further in the section. Therefore, we consider them as reworked species.

Among the Cretaceous fauna that we observed in the first Danian sample, only *G. cretacea* continues up in the section into Zone P1a (Table 2.1). *Guembelitria cretacea* is considered as a survivor and all microperforate Cenozoic species are phylogenetically related to *G. cretacea* based on wall structure and other morphologic characteristic (Olsson et al., 1999).

Zone Pa (Parvularugoglobigerina eugubina Taxon-Range Zone): The *Parvularugoglobigerina eugubina* Total Range Zone, also called Zone P α , is the interval characterized by the total range of the nominate taxon (Berggren et al., 1995; Olsson et al., 1999; Wade et al., 2011). Zone P α in the section occurs between samples B16–B18. It includes the following species: G. cretacea. G. aff. planocompressa, G. archeocompressa, Globoconusa daubjergensis, Woodringina hornerstownensis, W. claytonensis, E. eobulloides, Eoglobigerina edita, Subbotina trivialis, C. morsei, Z. waiparaensis, P. eugubina, Praemurica taurica, Praemurica pseudoinconstans, and P. pseudobulloides (Table 2.1, Fig. 2.5).

Zone P1a (*Parasubbotina pseudobulloides* **Partial-range Subzone**): Zone P1a is the *Parvularugoglobigerina eugubina-Subbotina triloculinoides* Interval Subzone. It is defined as the biostratigraphic interval between the LAD of *P. eugubina* and the FAD of *S. triloculinoides* (Olsson et al., 1999; Berggren and Pearson, 2005). The zone starts in sample B19, above the LAD of *P. eugubina* and continues to the top of the measured section (Table 2.1, Fig. 2.5). Important forms observed in this zone are: *G. cretacea, G.* *daubjergensis*, *W. hornerstownensis*, *W. claytonensis*, *E. eobulloides*, *E. edita*, *S. trivialis*, and *G. archeocompressa*.

2.5.3 Microfacies Analysis

The objective of the microfacies analysis is to integrate sedimentological and paleontological results to evaluate the depositional history of the section spanning the K/Pg boundary. We determined 10 microfacies (MF) types corresponding to inner ramp to basin environments (Fig. 2.6) using mineralogical components, macro-, and microfossil assemblages (Fig. 2.7), and texture of the samples observed in thin sections and outcrop. All microfacies types were deposited on an unrimmed shelf with no reef-derived components.

MF1: Bioclastic Packstone with Large Benthic Foraminifera and Calcareous Red Algae: Yellowish colored bioclastic packstone at the lowermost part of the section (samples A1–A9 and A12–A14) is characterized by abundant macro- and microfossils within a micritic matrix (Figs. 2.2, 2.8A-B). Large hyaline benthic foraminifera such as *Orbitoides, Lepidorbitoides, Siderolites, Sulcoperculina,* and *Hellenocyclina* are highly diversified. Agglutinated uncoiled benthic foraminifera and corallinacean and solenoporacean calcareous red algae are abundant. Mollusk fragments (pelecypoda, gastropoda, and cephalopoda), echinoderm fragments and spines, bryozoans, and calcareous green algae are also present. Planktonic foraminifera are present but rare. The size of the allochems is coarse and the percentage of the micritic matrix shows variability within the microfacies. In some samples sparitic calcite is also seen in low amounts. MF1 corresponds to an inner ramp environment, which is above the fair weather wave base (FWWB), and restricted to open marine conditions (Fig. 2.6). Large rotaliinid benthic foraminifera are rock forming in restricted and open marine environments in the platform interiors and inner ramps (Flügel, 2004). Furthermore, calcareous red algae occur only in the photic zone and in the platform margins and inner to mid-ramp settings (Flügel, 2004). Bryozoans, mollusks, and echinoderms are also seen in inner and midramp settings and platform interior to slope environments.

This microfacies is equivalent to the standard microfacies (SMF) 18-FOR (bioclastic grainstones and packstones with abundant benthic foraminifera or calcareous algae; "FOR" denotes abundance of foraminifera) and ramp microfacies (RMF) 13 (packstone with abundant larger foraminifera) of Flügel (2004). It is also equivalent to facies zone (FZ) 7 (platform interior, open marine) of Wilson (1975).

MF2: Grainstone with Large Benthic Foraminifera and Calcareous Red Algae: MF2 is characterized by abundant large benthic foraminifera and calcareous red algae within sparitic cement and identified in samples A10–A11 (Figs. 2.2, 2.8C-D). The most common large benthic foraminifera in MF2 are *Orbitoides*, *Lepidorbitoides*, *Omphalocyclus*, *Siderolites*, *Sulcoperculina*, and *Hellenocyclina*. Bryozoans and agglutinated benthic foraminifera are common and mollusk and echinoderm fragments are also present.

MF2 is similar to MF1 except the sparitic calcite content of the former. It represents a high-energy environment above the FWWB (Fig. 2.6). It can be considered equivalent to the SMF18 of Flügel (2004), which is composed of bioclastic grainstones and packstones with abundant benthic foraminifera or calcareous algae. It is also

equivalent to FZ7 of Wilson (1975), which corresponds to an inner ramp environment with restricted or open marine conditions.

MF3: Bioclastic Wackestone-Packstone with Benthic Foraminifera and Calcareous Red Algae: This microfacies type occur in the lower parts of the measured section in samples A15–A16, A18, and A23 (Fig. 2.2). It is similar to MF1 and MF2 in terms of faunal content. However, the amount of micrite and the size and the abundance of the allochems are lower in this microfacies (Fig. 2.8E-F). Hyaline benthic foraminifera and calcareous red algae are abundant. Agglutinated benthic foraminifera, echinoderm spines, bryozoans, and mollusk fragments are quite common and calcareous green algae and some calcispheres are also observed. Planktonic foraminifera are rare. Considering its texture and fossil content, MF3 belongs to an environment above the FWWB and corresponds to SMF18-FOR and RMF13 of Flügel (2004) like MF1 and MF2 (Fig. 2.6).

MF4: Bivalved Floatstone: MF4 is in samples A17 and A19–A22 and composed of large mollusks, especially bivalve fragments (Fig. 2.8G-H). The shell fragments exceeds 2 mm in size and constitute almost 50% of the rock embedded within a micritic matrix. In between the bivalve fragments there are planktonic and benthic foraminifera and micritic mud containing calcispheres. The size of the foraminifera and other bioclasts within the shell pieces is very small. Besides bivalves, foraminifera and calcispheres, echinoderms, bryozoans, calcareous red and green algae are quite common, whereas large hyaline benthic foraminifera are rare in MF4.

MF4 is very similar to the SMF12-Bs (Bs stands for bivalves) defined by Flügel (2004). The SMF12-Bs was defined as densely packed floatstones characterized by accumulations of one-type of shell concentration. Bivalve shell beds can be found in a

wide range of environments from platform interior settings/inner ramp to deep-marine settings (Flügel, 2004). Because the shell concentrations are seen within a micritic matrix abundant in pelagic organisms, we concluded that this facies was deposited in a mid-ramp environment between FWWB and storm wave base (SWB) (Fig. 2.6).

MF5: Wackestone with Planktonic Organisms: MF5 is in samples A24–A26 and characterized by its very fine micritic matrix containing common planktonic foraminifera and calcispheres. Besides the planktonic organisms, it also contains hyaline and agglutinated benthic foraminifera, calcareous red algae, mollusk fragments, bryozoans, and echinoderms (Fig. 2.9A-B).

In this facies, we identified common planktonic foraminifera in association with algae and benthic foraminifera indicating a deepening in the section. Although the number of planktonic foraminifera increases in this facies relative to the MF1–4, large benthic foraminifera (e.g., *Helenocyclina*) are still common. Middle to outer neritic hyaline benthic foraminifera (e.g., *Nodosaria*) also exist in MF5. *Nodosaria* in association with planktonic foraminifers indicate a deeper shelf environment (Geel, 2000). Hence, we concluded that MF5 was deposited below SWB in a mid-ramp to outer ramp setting (Fig. 2.6).

MF6: Quartz-Rich Silty Limestone with Benthic and Planktonic Foraminifera and Calcareous Red Algae: This facies was described as dark gray, silty limestone in the field and includes samples A27–A30, and A34–A35, and A38–A39 (Fig. 2.2). It is quartz-rich and silty; however, the calcium carbonate matrix is the dominant constituent. The facies is fossiliferous and contains agglutinated benthic foraminifera, hyaline smaller benthic foraminifera, planktonic foraminifera, bryozoan, mollusk fragments, echinoderm spines, and calcareous red algae (Fig. 2.9C-D). Benthic foraminifera and calcareous red algae are especially dominant. The number of middle to outer neritic smaller benthic foraminifera like *Nodosaria* increases in the facies relative to their abundance in the older levels in the measured section.

MF6 shows characteristics of proximal outer ramp settings (Fig. 2.6). The model established by El Gadi and Brookfield (1999) and the RMF model proposed by Flügel (2004) place the marly facies in the outer ramp settings, below SWB. In addition, Heldt et al. (2008) has defined the silty limestone facies with planktonic and small benthic foraminifera and shell fragments in proximal outer ramp settings.

MF7: Iron-Rich Silty Marl with Planktonic and Benthic Foraminifera: This facies was described as dark gray marl in the field. It is one of the most common MF types in the studied section and spans the Upper Cretaceous sediments towards to the K/Pg boundary beds in samples A30–A34, A36–A38, and A40–A48 (Figs. 2.2, 2.9E-F). Iron and clay minerals are very rich and most shells are filled with iron-oxide minerals. Quartz and feldspar grains are also common in this facies. The sizes of quartz grains are usually larger than those of quartz grains seen in MF6. Although there are some red algae and other bioclasts, the main fossils in MF7 are smaller benthic and planktonic foraminifera. This marly facies was deposited in low energy settings below SWB. The dominance of deeper-water benthic and planktonic foraminifera indicate a proximal outer ramp setting (Fig. 2.6).

MF8: Silty Limestone with Planktonic and Benthic Foraminifera: This facies found in the Danian has been observed as a yellowish-brownish limestone in the field and is represented by samples B20–B25, B28, B30, A56, A58, and A60 (Figs. 2.2, 2.3). This

silty limestone is rich in clay minerals and contains planktonic and small hyaline and agglutinated benthic foraminifera (Fig. 2.9G-H). The early Paleocene planktonic foraminifera are small and commonly filled with iron minerals.

Based on its textural and compositional properties, MF8 indicates deposition below SWB, in low-energy settings. It resembles MF6, but the greater abundance of planktonic foraminifera than benthic foraminifera, as well as the absence of calcareous algae indicates proximity to the basin. It should be also noted that the increase in the number of planktonic foraminifera in MF8 upsection might reflect the recovery of Danian planktonic foraminifera after the K/Pg mass extinction and might not be solely controlled by depositional settings. The number of planktonic foraminifera is slightly less in this facies than in MF9 and MF10 indicating that MF8 represents a shallower paleoenvironment. We consider MF8 as an outer ramp facies (Fig. 2.6).

MF9: Silty Marl with Planktonic and Benthic Foraminifera: This facies was described as greenish gray marl in outcrop and is found in the lowermost Danian. It is represented by samples B1–B13, B16–B19, B26–B27, B29, A55, A57, and A59 (Figs. 2.2, 2.3). MF9 has a marly character consisting of silt size siliciclastic particles and micrite, rich in planktonic foraminifera and agglutinated and smaller hyaline benthic foraminifera (Fig. 2.10A-B).

MF9 has similar characteristics to MF7 in terms of texture and composition. However, the main difference between MF9 and MF7 is the iron-oxide rich composition of the latter and the abundance of planktonic foraminifera in MF9 indicating a paleoenvironment below SWB (Fig. 2.6). Furthermore, MF9 has a greenish color due to the enrichment of clay minerals. **MF10:** Silty Marl with Clay Minerals and Fecal Pellets: MF10 corresponds to lowermost Danian beds in samples B14, B15/A51, and B17/A52 (Figs. 2.2, 2.3). It is a marly unit that has both micritic matrix and silt-size siliciclastic material and is characterized by benthic and planktonic foraminifera. Large agglutinated non-coiled benthic foraminifera are common. Hyaline deep-sea benthic foraminifera displaying various sizes are also present. Early Danian planktonic foraminifera in this microfacies are very small and often filled with iron-oxide minerals (Fig. 2.10C-D).

In MF10, quartz and feldspar grains and green clay minerals showing rounded to angular shapes are very abundant. There is a prominent increase in clay minerals including glauconite, smectite, and chlorite (Fig. 2.10E-F), and echinoid fecal pellets (Fig. 2.10G-I). The echinoid fecal pellets in the section are ~1 mm long, orange to brown ellipsoidal grains with smooth surface (Fig. 2.10G-I). They were assigned to Echinodermata, especially of Echinoidea by Voigt (1929) and previously described in the lowest Paleocene in Denmark and Sweden in large quantities (Brotzen, 1948; their pl. 2, fig. 2). Quantitative analysis of echinoid fecal pellets in both sections A and B shows that the number of pellets per gram of sample increases from 0 to12 in the first sample of Danian (A51 and B14; Figs. 2.2, 2.3). Some samples in section B (B17, B19, B21, B23; in P α and lower P1a) still contains some echinoid fecal pellets, but their number goes to 0 upsection (Fig. 2.3). MF10 represents a basin environment below the SWB similar to MF9 but slightly deeper (Fig. 2.6).

2.5.4 Sequence Stratigraphy and Sea-Level Changes

The measured section consists of three main sedimentary packages. The bottom of the section is composed of carbonate rocks (first package), the middle section is composed of a thick marly succession (second package), and the top is composed of carbonate-marl alternation (third package).

The first package consisting of carbonate rocks covers the upper Maastrichtian interval from sample A1 to A30 (Fig. 2.2). The MF types observed in this package are: bioclastic packstone with large benthic foraminifera and calcareous red algae (MF1); grainstone with large benthic foraminifera and calcareous red algae (MF2); bioclastic wackestone-packstone with benthic foraminifera and calcareous red algae (MF3); bivalved floatstone (MF4); and wackestone with planktonic organisms (MF5). The second package is composed of silty limestones and silty marls and covers the upper Maastrichtian to basal Danian interval from the sample A31 to A52/B17 (Figs. 2.2, 2.3). In this marly package we observe quartz-rich silty limestone with benthic and planktonic foraminifera (MF7), silty marl with planktonic and benthic foraminifera (MF7), silty marl with planktonic and benthic foraminifera (MF7), silty marl with planktonic and benthic foraminifera (MF9) – silty limestone (MF8) alternations in samples A53 and A60.

The basic approach for the sequence stratigraphic interpretation of the section lies behind the fact that carbonate rocks prograde into the basin when accommodation decreases and there is not enough space for carbonate growth. Carbonate growth is limited by the creation of accommodation and the deposition of carbonates takes place from a platform or an upslope top to the basin as "highstand shedding" (Emery and Myers, 1996). In the highstand systems tract (HST), when the carbonate production exceeds the rate of creation of accommodation space, carbonates are shed off the platform top to the slope and basin, which is called the "keep up" phase (Neumann and Macintyre, 1985). The first carbonate package in the section consisting of mainly packstones and grainstones was deposited during a HST and high energy carbonates of inner to mid-ramp (MF1-MF4) correspond to the keep-up phase. The carbonate source, which prograded into the basin during a HST, is considered to be the reefal and algal limestones of the Çaldağ Formation in the Haymana Basin.

Above the first carbonate succession containing packstones, grainstones and wackestones, wackestone with planktonic organisms such as planktonic foraminifera and calcispheres (MF5) was deposited during a deepening in the basin and a transition from mid-ramp to outer ramp (Fig. 2.6). Therefore, a transgressive surface (TS) is between the samples A23 and A24. Above the TS there is a package of quartz-rich silty limestones (MF6, samples A26–A30) containing planktonic foraminifera, benthic foraminifera, and ammonites indicating deepening and above this unit transgression continues into the silty marls.

Silty marls (second package) start in the measured section at sample A30. As more accommodation is created with a relative sea-level rise, progradation of carbonates ceases and the transgressive systems tract (TST) starts. Silty marls indicate a general transgressive pattern till sample A52/B17 corresponding to outer ramp to basin environment (MF7, MF9).

The MFS is placed between samples A52 and A53 in section A and at sample B17 in section B, where the change from the distal basin (MF10) to proximal basin occurs (MF9). Above the MFS, alternation of distal outer ramp silty limestones (MF8) and proximal basin silty marls (MF9) are observed (third package; Fig. 2.6). Above sample A52/B17, at sample A53, carbonates reappear in the measured section, indicating a HST (Figs. 2.2, 2.3). Progradation of the carbonates in the early Danian indicates the latest part of a relative sea-level rise. The K/Pg boundary is just below the maximum flooding surface (MFS), in the uppermost part of the TST.

We developed a relative sea-level curve (Fig. 2.11) for the section using our microfacies model (Fig. 2.6). We took the depth of inner ramp/mid-ramp boundary (FWWB) as between 10–15 m and mid-rap/outer ramp boundary (SWB) as 30 m (Burchette and Wright, 1992) and estimated the relative paleowater depth of each MF type accordingly. Although absolute paleowater depths we assigned to the MF types might not be accurate, their relative positions to each other give us an idea about how relative sea level changed across the K/Pg boundary. The paleowater depth estimates for the Haymana Basin indicated that maximum water depth was reached in MF10, just above the K/Pg boundary (Fig. 2.11) placing the boundary itself in a TST.

In order to test the sea-level changes across the K/Pg boundary in a different geographic location and compare it with the Haymana Basin, we studied a section in the Campo Pit, New Jersey (Fig. 2.11). The Campo Pit section was deposited in a shallow siliciclastic shelf environment sensitive to minor sea-level changes and shows a very similar history for sea-level changes across the K/Pg boundary, where the boundary is placed in a later phase of a TST, below a MFS (Fig. 2.11). In the section, white-

yellowish, medium to coarse-grained sandstone of the Redbank Formation is unconformably overlain by the yellowish-brownish, medium to coarse-grained sandstone of the Tinton Formation (Fig. 2.11). The uppermost Redbank Formation represents upper shoreface environment with planar laminations and abundant, vertical and horizontal Ophiomopha burrows. The overlying Tinton Formation at the Campo Pit shows a slight fining upward succession with increased mud content. The top of the Tinton Formation at Campo Pit is a reddish indurated unit, which is overlain by the K/Pg boundary event bed. The event bed consists of red and green mud with basal clay clasts containing impact spherules. At the base of the event bed, the lowest occurrence of Senoniasphaera *inornata*, a dinoflagellate marker species, indicates the earliest Danian (Miller et al., 2011). Above the event bed, there is clayey green glauconitic sand of the Hornerstown Formation. Authigenic green glauconite of the Hornerstown Formation indicates an offshore environment below storm wave base. A MFS is associated with the maximum percentage of glauconite. In the HST, above the MFS there is a gradual increase in quartz grains indicating a shallowing. For the paleowater depth curve of Campo Pit (Fig. 2.11), we took upper shoreface/middle shoreface boundary (FWWB) as between 10-15 m, middle shoreface/lower shoreface boundary as 20 m, and lower shoreface/offshore boundary (SWB) as 30 m (Browning et al., 2008).

2.6 Discussion

Planktonic foraminifera biostratigraphy indicates that the K/Pg transition is biostratigraphically complete in the Haymana section. In spite of the poor to moderate

preservation of the foraminifera, index species defining the upper Maastrichtian and lower Danian biozones were identified. The basal part of the measured section (samples A1–A44) cannot be constrained at the level of biozones, because of the rareness of the planktonic foraminifera due to the facies control.

The Haymana Basin records an abrupt and catastrophic mass extinction of planktonic foraminifera. At the boundary, all the large, ornate, keeled forms of the genera *Globotruncana*, *Globotruncanita*, *Globotruncanella*, *Rugoglobigerina*, *Racemiguembelina*, *Pseudotextularia* and *Pseudoguembelina* disappear. Above the extinction level, minute and delicate first Danian forms of the genera *Eoglobigerina*, *Globanomalina*, *Globoconusa*, and *Woodringina* appear.

In the first sample dated as Danian (B14), we encountered some Cretaceous taxa as well. Reworked specimens are difficult to identify unless they show differential preservation or large age difference with the rest of the faunal assemblage. The Cretaceous species in our lower Danian sample do not show distinct characteristics that may enable us to decide whether they are survivor or reworked species. However, except for *G. cretacea*, none of the species are seen up in the section above the first 5 cm of the Danian and they are much lower in relative abundance relative to the Cretaceous samples. Therefore, we interpret that only *G. cretacea* survived the boundary. In fact, detailed taxonomic studies of the Paleogene Planktonic Foraminifera Working Group (Olsson et al., 1999) concluded that *G. cretacea* is one of the survivors of the K/Pg boundary mass extinction with *H. holmdelensis* and *H. monmouthensis*.

Microfacies analysis indicated that deposition took place on an inner ramp (restricted and open marine conditions) with alternations of carbonates (packstones, grainstone, wackestones, and floatstones) to a basin with mixed siliciclastic-carbonate rocks (silty marls, silty limestones). The section is quite fossiliferous having large hyaline benthic foraminifera, deeper water hyaline, and agglutinated benthic foraminifera, mollusk shells, echinoderm spines, bryozoans, calcispheres, and corallinacean, and solenoporacean red algae. The presence, absence and abundances of these fossils were used in the determination of the environment of deposition in addition to textural properties of rocks.

For the sequence stratigraphic interpretation of the section, we considered overall geometry, depositional history and geologic setting of the whole basin in addition to the microfacies analysis of the measured section. Our results indicate that the K/Pg boundary is recorded in the latest part of a TST, below a MFS. Enrichment of the authigenic clay minerals (Fig. 2.10E-F) just above the boundary (MF10) and below the MFS might also indicate a relative rise in sea level.

Although, analysis of additional sections in the basin is necessary to have a more complete view on the environment of deposition and better constrain the sequence stratigraphy, the detailed microfacies analysis gives clues about the relative sea-level evolution. We constructed a relative sea-level curve for the Haymana Basin assigning paleowater depths to each MF types based on their relative positions to FWWB and SWB (Fig. 2.11). Our relative sea-level curve showed the TS and MFS clearly putting the K/Pg boundary in a late TST.

There is no consensus on global sea-level changes across the K/Pg boundary, likely due to the fact that the changes were subtle. The study of Vail et al. (1977) showed a major fall at the K/Pg boundary, whereas Haq et al. (1987) showed a 25 m sea-level rise

from the K/Pg boundary into the Danian. Donovan et al. (1988) showed a major fall in the Late Maastrichtian and a sea-level rise across the K/Pg boundary in Braggs section, Alabama. Olsson and Liu (1993) made important biostratigraphic revisions on the Braggs section (Donovan et al., 1988) and moved the boundary to a lower stratigraphic position. Accordingly, Olsson and Liu (1993) stated that sea level was at lowstand both before and after the boundary with a rise afterwards. Coarse-grained siliciclastic and limestone breccia deposits at the K/Pg boundary close to the impact site have been interpreted as pre K/Pg sea-level lowstand deposits (Donovan et al., 1988; Keller and Stinnesbeck, 1996; Keller, 2007). However, this interpretation has been challenged by the fact that there are impact-generated tsunami deposits in these regions (e.g. Smit, 1999) that have no relation to eustasy (e.g. Claeys et al., 2002).

European sections generally indicate a similar history for the sea level across the K/Pg boundary in a number of locations; a SB below the boundary, with the K/Pg boundary placed within the overlying TST (Apellaniz et al., 1997; Pujalte, 1998). Alegret et al. (2003) and Pardo et al. (1996) in Spain, and Pardo et al. (1999) in Kazakhstan indicated a sea-level rise across the K/Pg boundary. Global sea-level review of Keller and Stinnesbeck (1996) and the Tunisian record (Adatte et al., 2002) also showed a global sea-level rise across the K/Pg boundary into the Danian. We observe a very similar trend in the Haymana Basin, where the K/Pg boundary is placed in a late TST below a MFS.

Previous studies from New Jersey have provided important sea-level records from the Cretaceous to Paleogene. There are two major sea-level falls recorded, one being at the Campanian/Maastrichtian boundary (Miller et al., 2004) and the other being at the base of the planktonic foraminiferal Zone P1b (Olsson et al., 2002). Benthic foraminiferal biofacies (Olsson et al., 2002) and the backstripping results (Miller et al., 2005; Kominz et al., 2008) showed a minimal fall in sea level (< 25 m) associated with the K/Pg boundary, which is similar to the trend shown in the Haq curve (2014). However new data from the updip section, Campo Pit, show that the K/Pg boundary is within a sea-level rise in New Jersey (Fig. 2.11).

Detailed microfacies analysis of the Haymana Basin shows a very similar sealevel history not only to European and Tethyan sections (Keller and Stinnesbeck, 1996; Pardo et al., 1996; 1999; Apellaniz et al., 1997; Pujalte, 1998; Adatte et al., 2002), but also to the Campo Pit section, New Jersey, where the K/Pg boundary is in a TST (Fig. 2.11). This might indicate that the sea level in the Haymana Basin was controlled mainly by eustasy despite of tectonic activity in the Central Anatolia. We acknowledge that reference frames for sea-level change are impacted by processes of regional tectonics and mantle dynamics (e.g., Moucha et al., 2008; Müller et al., 2008); however, these mechanisms largely affect the stratigraphic record on longer than million year scales (e.g., Miller et al., 2005). Our observations of sea-level change in both Haymana and Campo Pit sections are much shorter term, thus, less likely impacted by these tectonic effects.

The increase in clay minerals like glauconite, smectite, and chlorite in the immediate aftermath of the K/Pg boundary in Zone P0 (sample A51/B15) might indicate an overall decrease in sedimentation rates, which is coupled with the drop in carbonate deposition mostly due to the mass extinction of carbonate producing planktonic foraminifera and calcareous nannoplankton (Fig. 2.10E-F).

Another remarkable feature of Zone P0 is that the prominent increase in echinoid fecal pellets (Fig. 2.10G-I). Our quantitative analysis indicates that echinoid fecal pellets appear in the first sample of the Danian just after the K/Pg boundary. They exist in the section up to lower P1a and then disappear again (Figs. 2.2, 2.3). Miller et al. (2010) reported an increase in echinoid fecal pellets in the lowermost Danian beds in several New Jersey Coastal Plain downdip (i.e., deeper water) sections as well and suggested that this increase can be used as a stratigraphic horizon for the correlation of the K/Pg boundary. Our observation in the Haymana Basin supports the idea that the increase in the number of echinoid fecal pellets indicates an important environmental change after the K/Pg mass extinction and can be used as a marker for the K/Pg boundary in addition to other criteria defining the boundary. The absence of the fecal pellet horizon at Campo Pit (this study) and the adjacent Buck Pit core (Miller et al., 2010) suggests a facies control on this horizon due to non-deposition or post-depositional alteration in shoreface settings, and its presence in deeper water environments (e.g., Haymana Basin (this study), deep New Jersey sections (Miller et al., 2010), and Swedish sections (Brotzen, 1948)).

2.7 Conclusions

A section spanning the K/Pg boundary was examined in the Haymana Basin, Central Anatolia, Turkey with detailed planktonic foraminiferal biostratigraphy, microfacies analysis, and sequence stratigraphy. The biostratigraphic record showed that the K/Pg boundary in the Haymana Basin is complete and the extinction of the planktonic foraminifera is abrupt and catastrophic, as has been shown by previous researchers in many other K/Pg sections globally. Zone P0 is characterized by the evolution of early Danian planktonic foraminifera and associated with an increase in clay minerals and echinoid fecal pellets; this increase seems to be a correlatable level to several New Jersey Coastal Plain sections. Our relative sea-level curve shows that the K/Pg boundary in the Haymana Basin was deposited during a sea-level rise below a MFS. Comparison of the section with the Campo Pit section, New Jersey and other sections from Europe and North Africa showed a similar sea-level rise across the K/Pg boundary.

2.8 Acknowledgments

We thank the Scientific Research Projects Coordination of the Middle East Technical University for providing financial support for this study. We would like to thank Mevlüt Ertan for his contribution to the improvement of the washing techniques, Zühtü Batı for reviews on the manuscript, Richard K. Olsson for discussions on the planktonic foraminiferal taxonomy and fecal pellets, James V. Browning for discussions on the Campo Pit section, and James D. Wright for comments. We are thankful to Brian Huber and Steve D'Hondt for reviews, and to the Associate Editor Elias Samankassou and the Editor Gene Rankey for comments.

- Adatte, T., Keller, G., and Stinnesbeck, W., 2002, Late Cretaceous to early Paleocene climate and sea-level fluctuations: the Tunisian record: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 178, p. 165–196.
- Alegret, L., Molina, E., and Thomas, E., 2003, Benthic foraminiferal turnover across the Cretaceous/Paleogene boundary at Agost (southeastern Spain): paleoenvironmental inferences: Marine Micropaleontology, v. 48, p. 251–279.
- Apellaniz, E., Baceta, J.I., Bernaola-Bilbao, G., Nunez-Betelu, K., Orue-Etxebarria, X., Payros, A., Pujalte, V., Robin, E., and Rocchia, R., 1997, Analysis of uppermost Cretaceous-lowermost Tertiary hemipelagic successions in the Basque Country (western Pyrenees); evidence for a sudden extinction of more than half planktic foraminifer species at the K/T boundary: Bulletin de la Societe Geologique de France, v. 168, p. 783–793.
- Arenillas, I., Arz, J.A., Grajales-Nishimura, J., Murillo-Muñetón, G., Alvarez, W., Camargo-Zanoguera, A., Molina, E., and Rosales-Domínguez, M., 2006, Chicxulub impact event is Cretaceous/Paleogene boundary in age: New micropaleontological evidence: Earth and Planetary Science Letters, v. 249, p. 241–257.
- Ben Abdelkader, O., Ben Salem, H., Donze, P., Maamouri, A., Méon, H., Robin, É., Rocchia, R., and Froget, L., 1997, The K/T stratotype section of El Kef (Tunisia): events and biotic turnovers: Geobios, v. 21, p. 235–245.
- Berggren, W.A., and Pearson, P., 2005, A revised tropical to subtropical Paleogene planktonic foraminiferal zonation: Journal of Foraminiferal Research, v. 35, p. 279–298.
- Berggren, W.A., Kent, D.V., Swisher, C., III, and Aubry, M.-P., 1995, A revised Cenozoic geochronology and chronostratigraphy, *in* Berg- gren, W. A., Kent, D. V., Swisher, C. C., III, Aubry, M.-P. and Hardenbol, J., (eds.), Geochronology, Time Scales and Global Stratigraphic Correlation: SEPM (Society for Sedimentary Geology) Special Publication, v. 54, p. 129–212.
- Bernaola, G., and Monechi, S., 2007, Calcareous nannofossil extinction and survivorship across the Cretaceous–Paleogene boundary at Walvis Ridge (ODP Hole 1262C, South Atlantic Ocean): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 132–156, doi: 10.1016/j.palaeo.2007.02.045.
- Brotzen, F., 1948, The Swedish Paleocene and Its Foraminiferal Fauna: Sweriges Geologiska Undersökning, v. 42, p. 1–140.

- Browning, J.V., Miller, K.G., Sugarman, P.J., Kominz, M.A., McLaughlin, P.P., Jr, Kulpecz, A.A., and Feigenson, M.D., 2008, 100 Myr record of sequences, sedimentary facies and sea level change from Ocean Drilling Program onshore coreholes, US Mid-Atlantic coastal plain: Basin Research, v. 20, p. 227–248, doi: 10.1111/j.1365-2117.2008.00360.x.
- Burchette, T.P., and Wright, V.P., 1992, Carbonate ramp depositional systems: In, B.W. Sellwood (Ed.), Ramps and Reefs. Sedimentary Geology, v. 79, p. 3–57.
- Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of Chicxulub ejecta at the Cretaceous-Tertiary boundary (C. Koeberl & K. G. MacLeod, Eds.): Catastrophic events and Mass extinctions: impacts and Beyond: Geological Society of America Special Paper, v. 356, p. 55–68.
- Çiner, A., 1992, Sediméntologie et stratigraphie séquentielle du bassin d'Haymana à l'Eocene moyen Turquie: PhD Thesis, l'Univ. Louis Pasteur, France, 190 p.
- Çiner, A., Deynoux, M., Ricou, S., and Kosun, E., 1996, Cyclicity in the Middle Eocene Cayraz Carbonate Formation, Haymana Basin, Central Anatolia, Turkey: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 121, p. 313–329.
- Donovan, A.D., Baum, G.R., Blechschmidt, G.L., Loutit, T.S., Pflum, C.E., and Vail, P.R., 1988, Sequence stratigraphic setting of the Cretaceous-Tertiary boundary in central Alabama. in Wilgus, C.K., Kendall, C.G.S.C., Posamentier, H.W., and Ross, C.A., eds., Sea-level changes: An integrated approach: Society of Economic Paleontologists and Mineralogists Special Publication, v. 42, p. 299–307.
- Dunham, R.J., 1962, Classification of carbonate rocks according to the depositional texture in: Ham, W. E. (ed.): Classification of carbonate rocks: A symposium, AAPG Memoir, v. 1, p. 108–171.
- Embry, A.F., and Klovan, J.E., 1971, A late Devonian reef tract on northeastern Banks Island: Northwest Territories: Bulletin of Canadian Petroleum Geology, v. 19, p. 730–781.
- Emery, D., and Myers, K.J., 1996, Sequence Stratigraphy: Blackwell Science, 297 p.
- Flügel, E., 2004, Microfacies of carbonate rocks: analysis, interpretation and application: Springer, 976 p.
- Fourquin, C., 1975, L'anatolie du Nord-Ouest, marge méridionale du continent europeén, histoire paléogéographique, tectonique et magmatique durant le Secondaire et Tertiare: Bull. Soc. Géol. France, v. 7, p. 1058–1070.
- Gadi, El, M.S.M., and Brookfield, M.E., 1999, Open carbonate ramp facies, microfacies and paleoenvironments of the Gramame Formation (Maastrichtian), Pernambuco-Paraiba Basin, Northeastern Brazil: Journal of South American Earth Sciences, v. 12, p. 411–433.

- Geel, T., 2000, Recognition of stratigraphic sequences in carbonate platform and slope deposits: empirical models based on microfacies analysis of Palaeogene deposits in southeastern Spain: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 155, p. 211–238.
- Georgescu, M.D., and Abramovich, S., 2008, A new serial Cretaceous planktic foraminifer (Family Heterohelicidae Cushman, 1927) from the Upper Maastrichtian of the equatorial Central Pacific: Journal of Micropalaeontology, v. 27, p. 117–123.
- Görür, N., 1981, Tuzgölü-Haymana havzasının stratigrafik analizi: TJK, İç Anadolunun Jeolojisi Simpozyumu, Ankara, v. 1, p. 60–66.
- Görür, N., Oktay, Y.F., Seymen, I., and Şengör, A.M.C., 1984, Paleotectonic evolution of the Tuzgölü Basin complex, central Turkey: Sedimentary record of the neo-tethyan closure. In: Dixon J. and Robertson A. H. F. (eds), The Geological Evolution of the Eastern Mediterranean. Spec. Pub: Geological Society of London, v. 17, p. 467–481.
- Haq, B., Hardenbol, J., and Vail, P.R., 1987, Chronology of fluctuating sea levels since the Triassic: Science, v. 235, p. 1156–1167.
- Heldt, M., Bachmann, M., and Lehmann, J., 2008, Microfacies, biostratigraphy, and geochemistry of the hemipelagic Barremian-Aptian in north-central Tunisia: Influence of the OAE 1a on the southern Tethys margin: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 261, p. 246–260.
- Horowitz, A.S., and Potter, P.E., 1971, Introductory Petrography of Fossils: Springer Verlag, Heidelberg, Berlin, 302 p.
- Huber, B.T., 1991, 25. Maestrichtian Planktonic Foraminifer Biostratigraphy and the Cretaceous/Tertiary Boundary At Hole 738C (Kerguelen Plateau, Southern Indian Ocean) (J. Barron & B. Larsen, Eds.): Proceedings of the Ocean Drilling Program, Scientific Results, v. 119, p. 451–465.
- Huber, B.T., MacLeod, K.G., and Tur, N.A., 2008, Chronostratigraphic framework for upper Campanian-Maastrichtian sediments on the Blake Nose (subtropical North Atlantic): Journal of Foraminiferal Research, v. 38, p. 162–182.
- Keller, G., 1988, Extinction, survivorship and evolution of planktic foraminifera across the Cretaceous/Tertiary boundary at El Kef, Tunisia: Marine Micropaleontology, v. 13, p. 239–263.
- Keller, G., 2007, The Chicxulub impact and K-T mass extinction in Texas: Bulletin of South Texas Geological Society, v. XLVII, p. 15–44.
- Keller, G., and Stinnesbeck, W., 1996, Sea-level changes, clastic deposits, and megatsunamis across the Cretaceous–Tertiary boundary: In: MacLeod, N., Keller, G. (Eds.), Cretaceous–Tertiary Mass Extinctions: Biotic and Environmental Change. Norton, London, p. 415–449.
- Knitter, H., 1979, Eine verbesserte Methode zur Gewinnung von Mikrofosilien aus harten, nicht schlammbaren Kalken: Geol. Bl. Nor-Bayern, v. 29, p. 182 – 186.
- Koçyiğit, A., 1991, An example of an accretionary forearc basin from Central Anatolia and its implications for the history of subduction of Neo-Tethys in Turkey: Geological Society of America Bulletin, v. 103, p. 22–36.
- Koçyiğit, A., and Lünel, T., 1987, Geology and tectonic setting of Alcı region, Ankara: Middle East Technical University Journal of Pure and Applied Sciences, v. 20, p. 35– 57.
- Koçyiğit, A., Özkan, S., and Rojay, B., 1998, Examples of the forearc basin remnants at the active margin of northern Neo-Tethys: development and emplacement ages of the Anatolian Nappe, Turkey: Middle East Technical University Journal of Pure and Applied Sciences, v. 3, p. 183–210.
- Kominz, M.A., Browning, J.V., Miller, K.G., Sugarman, P.J., Mizintseva, S., and Scotese, C., 2008, Late Cretaceous to Miocene sea-level estimates from the New Jersey and Delaware coastal plain coreholes: an error analysis: Basin Research, v. 20, p. 211–226, doi: 10.1111/j.1365-2117.2008.00354.x.
- Koutsoukos, E., 1996, Phenotypic experiments into new pelagic niches in early Danian planktonic foraminifera: aftermath of the K/T boundary event: Geological Society of London, Special Publications, v. 102, p. 319–335.
- Li, L., and Keller, G., 1998, Maastrichtian climate, productivity and faunal turnovers in planktic foraminifera in South Atlantic DSDP sites 525A and 21: Marine Micropaleontology, v. 33, p. 55–86.
- Lirer, F., 2000, A new technique for retrieving calcareous microfossils from lithified lime deposits: Micropaleontology, v. 46, p. 365–369.
- Liu, C., and Olsson, R.K., 1992, Evolutionary radiation of microperforate planktonic Foraminifera following the K/T mass extinction event: Journal of Foraminiferal Research, v. 22, p. 328–346, doi: 10.2113/gsjfr.22.4.328.
- Luciani, V., 2002, High-resolution planktonic foraminiferal analysis from the Cretaceous-Tertiary boundary at Ain Settara (Tunisia): evidence of an extended mass extinction: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 178, p. 299–319.
- Luciani, V., 1997, Planktonic foraminiferal turnover across the Cretaceous-Tertiary boundary in the Vajont valley (Southern Alps, northern Italy): Cretaceous Research, v. 18, p. 799–821.
- Luterbacher, P.H., and Premoli Silva, I., 1964, Biostratigrafia del limite cretaceoterziario nell'Appennino centrale: Rivista Italiana di Paleontologia, v. 70, p. 67–117.

- Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., and Pekar, S., 2005, The Phanerozoic record of global sea-level change: Science, v. 310, p. 1293–1298, doi: 10.1126/science.1116412.
- Miller, K.G., Sherrell, R., Browning, J.V., Field, M., Gallagher, W., Olsson, R.K., Sugarman, P.J., Tuorto, S., and Wahyudi, H., 2010, Relationship between mass extinction and iridium across the Cretaceous-Paleogene boundary in New Jersey: Geology, v. 38, p. 867–870, doi: 10.1130/G31135.1.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Kominz, M.A., Olsson, R.K., Feigenson, M.D., and Hernandez, J., 2004, Upper Cretaceous sequences and sea-level history, New Jersey coastal plain: Geological Society of America Bulletin, v. 116, p. 368– 393.
- Moucha, R., Forte, A.M., Mitrovica, J.X., Rowley, D.B., Quéré, S., Simmons, N.A., and Grand, S.P., 2008, Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform: Earth and Planetary Science Letters, v. 271, p. 101–108, doi: 10.1016/j.epsl.2008.03.056.
- Mount, J., 1985, Mixed siliciclastic and carbonate sediments: a proposed first- order textural and compositional classification: Sedimentology, v. 32, p. 435–442.
- Müller, R.D., Sdrolias, M., Gaina, C., Steinberger, B., and Heine, C., 2008, Long-Term Sea-Level Fluctuations Driven by Ocean Basin Dynamics: Science, v. 319, p. 1357– 1362, doi: 10.1126/science.1151540.
- Nederbragt, A.J., 1991, Late Cretaceous biostratigraphy and development of Heterohelicidae (planktic foraminifera): Micropaleontology, v. 37, p. 329 –372.
- Neumann, A.C., and Macintyre, I., 1985, Reef response to sea-level rise: keep-up, catchup or give-up. : In: Proceedings of the Fifth International Coral Reef Congress, Tahiti, v. 3, p. 105–110.
- Obaidalla, N., 2005, Complete Cretaceous/Paleogene (K/P) boundary section at Wadi Nukhul, southwestern Sinai, Egypt: inference from planktic foraminiferal biostratigraphy: Revue de Paléobiologie, Genève, v. 24, p. 201–224.
- Olsson, R.K., and Liu, C., 1993, Controversies on the placement of Cretaceous-Paleogene boundary and the K/P mass extinction of planktonic foraminifera: Palaios, v. 8, p. 127–139.
- Olsson, R.K., Hemleben, C., Berggren, W.A., Huber, B.T., Editors and Members of the Paleogene Planktonic Foraminifera Working Group, 1999, Atlas of Paleocene Planktonic Foraminifera: Smithsonian Contributions to Paleobiology, v. 85, p. 252.

- Olsson, R.K., Miller, K.G., Browning, J.V., Wright, J.D., and Cramer, B.S., 2002, Sequence stratigraphy and sea-level change across the Cretaceous-Tertiary boundary on the New Jersey passive margin: Catastrophic events and Mass extinctions: impacts and beyond,, p. 97–108.
- Özkan-Altıner, S., and Özcan, E., 1999, Upper Cretaceous planktonic foraminiferal biostratigraphy from NW Turkey: calibration of the stratigraphic ranges of larger benthonic foraminifera: Geological Journal.
- Pardo, A., Adatte, T., Keller, G., and Oberhansli, H., 1999, Paleoenvironmental changes across the Cretaceous-Tertiary boundary at Koshak, Kazakhstan, based on planktic foraminifera and clay mineralogy: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 154, p. 247–273.
- Pardo, A., Ortiz, N., and Keller, G., 1996, Latest Maastrichtian and Cretaceous-Tertiary Boundary Foraminiferal Turnover and Environmental Changes at Agost, Spain (N. McLeod & G. Keller, Eds.): Cretaceous-Tertiary Mass Extinctions: Biotic and Environmental Changes Norton, New York/London, p.139-171.
- Premoli Silva, I., and Verga, D., 2004, Practical manual of Cretaceous Planktonic Foraminifera. International School on Planktonic Foraminifera: 3. Course: Cretaceous. Verga & Rettori eds. Universities of Perugia and Milan, Tipografia Pontefelcino, Perugia (Italy), 283 p.
- Pujalte, V., 1998, Paleocene Strata of the Basque Country, Western Pyrenees, Northern Spain: Facies and Sequence Development in a Deep-Water Starved Basin: Mesozoic and Cenozoic Sequence Stratigraphy of European Basin, SEPM Special Publications, v. 60, p. 311–325.
- Raup, D., and Sepkoski, J., Jr, 1982, Mass extinctions in the marine fossil record: Science, v. 215, p. 1501–1503.
- Robaszynski, F., 1998, Planktonic foraminifera-Upper Cretaceous, Chart of Cretaceous Biochronostratigraphy in: P.C. de Graciansky, J. Hardenbol, P.R. Vail (Eds.), Mesozoic and Cenozoic Sequence Stratigraphy of European Basins: Society For Sedimentary Geology (SEPM), Special Publication, v. 60, p. 1–782.
- Robaszynski, F., and Caron, M., 1995, Foraminifères planctoniques du Crétacé: commentaire de la zonation Europe-Méditerranée: Bulletin de la Societe Geologique de France, v. 166, p. 681–692.
- Robaszynski, F., Caron, M., Donoso Gonzalez, M.J., Wonders, A.H., The European Working Group on Planktonic Foraminifera, 1984, Atlas of Late Cretaceous globotruncanids: Revue de Micropaléontologie, v. 26, p. 145–305.
- Scholle, P.A., and Ulmer-Scholle, D.S., 2003, A color guide to the petrography of carbonates rocks: grains, textures, porosity, diagenesis: AAPG Memoir 77, Tulsa, Oklahoma, USA, 474 p.

- Schulte, P., Alegret, L., Arenillas, I., Arz, J.A., Barton, P., Bown, P., Bralower, T.J., Christeson, G., Claeys, P., and Cockell, C., 2010, The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary: Science, v. 327, p. 1214– 1218.
- Smit, J., 1982, Extinction and evolution of planktonic foraminifera after a major impact at the Cretaceous/Tertiary boundary: Geological Society of America Special Papers, v. 190, p. 329–352.
- Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Sciences, v. 27, p. 75–113.
- Şengör, A.M.C., and Yılmaz, Y., 1981, Tethyan evolution of Turkey: a plate tectonic approach: Tectonophysics, v. 75, p. 181–241.
- Ünalan, G., Yüksel, V., Tekeli, T., Gönenç, O., Seyirt, Z., and Hüseyin, S., 1976, Haymana-Polatlı yöresinin (Güneybatı Ankara) Üst Kretase-Alt Tersiyer stratigrafisi ve paleocoğrafik evrimi: TJK Bülteni, v. 19, p. 159–176.
- Vail, P.R., Mitchum, R.M., Jr, and Thompson, S., III, 1977, Seismic Stratigraphy and Global Changes of Sea Level: Part 3. Relative Changes of Sea Level from Coastal Onlap. (C. E. Payton, Ed.): Seismic Stratigraphy — Applications to Hydrocarbon Exploration, Memoir vol. 26, American Association of Petroleum Geologists, v. 26, p. 63–81.
- Voigt, E., 1929, Die Lithogenese der Flach- und Tiefwassersedimente des jüngeren Oberkreidemeeres: Halle, Jahrbuch d. Halleschen Verbandes z. Erforschung d. mitteldeutschen Bodensch tze u. Verwertung, v. 8, p. 1–136.
- Wade, B.S., Pearson, P., Berggren, W.A., and Palike, H., 2011, Review and revision of Cenozoic tropical planktonic foraminiferal biostratigraphy and calibration to the geomagnetic polarity and astronomical time scale: Earth-Science Reviews, v. 104, p. 111–142.
- Wilson, J.L., 1975, Carbonate facies in geologic history: Springer-Verlag, New York, 469 p.
- Yüksel, S., 1970, Etude géologique de la region d'Haymana (Turquie Centrale): Thèse Fac. Sci. Univ. De Nancy, 179 p.

2.10 Figures and Tables



Fig. 2.1. A. Location of the Haymana Basin within the main structural feature of Turkey (Koçyiğit, 1991; Çiner et al., 1996) **B.** Generalized geological map of the Haymana Basin (modified from 1/500.000 scale Turkey Map). The location of the study area (39°24′20″ N, 32°27′55″ E) is indicated with a black rectangle.



Fig. 2.2. Lithostratigraphy, planktonic foraminiferal biozones, microfacies types, sequence stratigraphical interpretation, and echinoid fecal pellet counts of the measured section (samples A1–A60). The most abundant biogenic and abiogenic components are shown in red. Note the stratigraphic level of the detailed section (samples B1–B30) across the boundary, which is shown in Fig. 2.3. HST: highstand systems tract, TST: transgressive systems tract, TS: transgressive surface, MFS: maximum flooding surface.



Fig. 2.3. Lithostratigraphy, planktonic foraminiferal biozones, microfacies types, sequence stratigraphical interpretation, and echinoid fecal pellet counts of the high-resolution section across the K/Pg boundary (Samples B1–B30). The following samples represent the same stratigraphic horizon: A49=B3, A50=B13, A51=B15, A52=B17, A53=B24. Refer to the legend in Fig. 2.2. HST: highstand systems tract, TST: transgressive systems tract, TS: transgressive surface, MFS: maximum flooding surface.



Fig. 2.4. Scanning electron microscope images of Maastrichtian planktonic foraminifera species. Scale bar represents 100 µm. 1. Globotruncana arca CUSHMAN, sample no. B1, P. hariaensis Zone, a. spiral view, b. side view. 2. Globotruncana orientalis EL NAGGAR, sample no. B5, *P. hariaensis* Zone, **a.** spiral view, **b.** side view. **3.** Globotruncana dupeublei CARON et al., sample no. B1, P. hariaensis Zone, spiral view. 4. Globotruncana aegyptiaca NAKKADY, sample no. B2, P. hariaensis Zone, spiral view. 5. Globotruncana aegyptiaca NAKKADY, sample no. B1, P. hariaensis Zone, umbilical view. 6. Rugoglobigerina pennyi BRONNIMANN, sample no. B3, P. hariaensis Zone, spiral view. 7. Globotruncana stuarti de LAPPARENT, sample no. B5, P. hariaensis Zone, a. spiral view, b. side view. 8. Globotruncanita stuartiformis DALBIEZ, sample no. B5, P. hariaensis Zone, a. spiral view, b. side view. 9. Globigerinelloides subcarinatus BRONNIMANN, sample no. A46, P. hariaensis Zone, spiral view. 10. Pseudoguembelina hariaensis NEDERBRAGT, sample no. B2, P. hariaensis Zone, side view. 11. Pseudoguembelina hariaensis NEDERBRAGT, sample no. A46, P. hariaensis Zone, side view. 12. Heterohelix globulosa EHRENBERG, sample no. B2, P. hariaensis Zone, side view. 13. Heterohelix globulosa EHRENBERG, sample no. A46, P. hariaensis Zone, side view. 14. Pseudotextularia nuttalli VOORWIJK, sample no. B5, P. hariaensis Zone, side view. 15. Pseudotextularia nuttalli VOORWIJK, sample no. B5, P. hariaensis Zone, edge view. 16. Laeviheterohelix glabrans CUSHMAN, sample no. A46, P. hariaensis Zone, side view. 17. Laeviheterohelix glabrans CUSHMAN, B14, P. hariaensis Zone, edge view. 18. Pseudotextularia elegans RZEHAK, sample no. B3, P. hariaensis Zone, edge view. **19.** Planoglobulina acervulinoides EGGER, sample no. B1, P. hariaensis Zone, side view. 20. Planoglobulina carseyae PLUMMER, sample no. B2, P. hariaensis Zone, side view. 21. Racemiguembelina fructicosa EGGER, sample no. A46, P. hariaensis Zone, side view. 22. Racemiguembelina powelli SMITH and PESSAGNO, sample no. A46, P. hariaensis Zone, side view.



Fig. 2.5. Scanning electron microscope images of Danian planktonic foraminifera species. Scale bar represents 100 µm for 1–15 and 50 µm for 16–18. **1.** Globanomalina aff. planocompressa SHUTSKAYA, sample no. B14, Zone P0, a. spiral view, b. side view. 2. Eoglobigerina eobulloides MOROZOVA, sample no. B27, Zone P1a, spiral view. 3. Eoglobigerina eobulloides MOROZOVA, sample no. B19, Zone P1a, spiral view. 4. Eoglobigerina eobulloides MOROZOVA, sample no. B19, Zone P1a, umblical view. 5. Eoglobigerina edita SUBBOTINA, sample no. B27, spiral view, Zone P1a, spiral view. 6. Eoglobigerina edita SUBBOTINA, sample no. B27, spiral view, Zone P1a, spiral view. 7. Parvularugoglobigerina eugubina LUTERBACHER and PREMOLI-SILVA, sample no. B16, Zone P1a, spiral view. 8. Praemurica taurica MOROZOVA, sample no. B27, Zone P1a, spiral view. 9. Subbotina trivialis SUBBOTINA, sample no. B27, Zone P1a, spiral view. 10. Subbotina trivialis SUBBOTINA, sample no. B27, Zone P1a, spiral view. **11.** Subbotina trivialis SUBBOTINA, sample no. B17, Zone Pα, spiral view. 12. Subbotina trivialis SUBBOTINA, sample no. B27, Zone P1a, umbilical view. 13. Subbotina trivialis SUBBOTINA, sample no. B19, Zone P1a, umblical view 14. Parasubbotina pseudobulloides PLUMMER, sample no. B16, Zone Pa, spiral view. **15.** Parasubbotina pseudobulloides PLUMMER, sample no. B19, Zone P1a, umbilical view. 16. Guembelitria cretacea CUSHMAN, sample no. A60, Zone P1a, side view. 17. Woodringina claytonensis LOEBLICH and TAPPAN, sample no. B14, Zone P0, side view. 18. Woodringina claytonensis LOEBLICH and TAPPAN, sample no. B16, Zone $P\alpha$, side view.



Fig. 2.6. Model showing the paleoenvironment and lateral relationship of the microfacies types. Modified after ramp microfacies types model from Flügel (2004). Refer to the legend in Fig. 2.2. FWWB: fair weather wave base, SWB: storm wave base.



Fig. 2.7. Photomicrographs of the major fossil groups identified within the frame of the microfacies analysis. 1. Gastropoda shell (x6, sample no. A21). 2–4. Pelecypoda shell (2. x10, sample no. A1. 3. x15, sample no. A11. 4. x7, sample no. A4). 5–7. Echinodermata spine (5. x16, sample no A7. 6. x40, sample no. A19. 7. x50, sample no. A26).
8–10. Bryozoan shell (8. x25, sample no. A8. 9. x12, sample no. A17. 10. x20, sample no. A8). 11, 13–14. Hyaline large benthic foraminifera (11. x20, sample no. A1. 13. x15, sample no. A11. 14. x25, sample no. A1). 12, 15–19. Hyaline smaller benthic foraminifera (12. x18, sample no. A3. 15. x33, sample no. A2. 16. x52, sample no. A13. 17. x26, sample no. A1. 18. x12, sample no. A14. 19. x10, sample no. A25).
20–24. Agglutinated benthic foraminifera (20. x20, sample no. A16. 21. x35, sample no. A24. 22. x84, sample no. A22. 23. x38, sample no. A15. 24. x18, sample no. A6).
25–27. Calcareous red algae (25. x13, sample no. A2. 26. x45, sample no. A4. 27. x30, sample no. A2).



Fig. 2.8. Photomicrographs of the microfacies types. **A.** Bioclastic packstone with large benthic foraminifera and calcareous red algae (MF1), sample no. A1. **B.** Bioclastic packstone with large benthic foraminifera and calcareous red algae (MF1), sample no. A2. **C.** Grainstone with large benthic foraminifera and calcareous red algae (MF2), sample no. A10 **D.** Grainstone with large benthic foraminifera and calcareous red algae (MF2), sample no. A10. **E.** Bioclastic wackestone-packstone with benthic foraminifera and calcareous red algae (MF2), sample no. A10. **E.** Bioclastic wackestone-packstone with benthic foraminifera and calcareous red algae (MF2), sample no. A10. **E.** Bioclastic wackestone-packstone with benthic foraminifera and calcareous red algae (MF3), sample no. A18. **F.** Bioclastic wackestone-packstone with benthic foraminifera and calcareous red algae (MF3), sample no. A19. **H.** Bivalved floatstone (MF4), sample no. A19. (Abbreviations: hb: hyaline benthic foraminifera, ab: agglutinated benthic foraminifera, p: planktonic foraminifera, ra: calcareous red algae, ga: calcareous green algae, m: mollusk fragment, b: bryozoan, e: echinodermata spine. Scale bar is 0.25 mm for A, B, E, F, H and 0.50 mm for C, D, G).



Fig. 2.9. Photomicrographs of the microfacies types. A. Wackestone with planktonic organisms (MF5), sample no. A25. B. Wackestone with planktonic organisms (MF5), sample no. A25. C. Quartz-rich silty limestone with benthic and planktonic foraminifera and calcareous red algae (MF6), sample no. A38. D. Quartz-rich silty limestone with benthic and planktonic foraminifera and calcareous red algae (MF6), sample no. A38. E. Iron-rich silty marl with planktonic and benthic foraminifera (MF7), sample no. A37. F. Iron-rich silty marl with planktonic and benthic foraminifera (MF7), sample no. A37. G. Silty limestone with planktonic and benthic foraminifera (MF8), sample no. A54. H. Silty limestone with planktonic and benthic foraminifera (MF8), sample no. A60. (Abbreviations: hb: hyaline benthic foraminifera, ab: agglutinated benthic foraminifera, p: planktonic foraminifera, ra: calcareous red algae, ga: calcareous green algae, m: mollusk fragment, b: bryozoan, e: echinodermata spine, c: calcispheres, q: quartz. Scale bar is 0.25 mm).



Fig. 2.10. A. Silty marl with planktonic and benthic foraminifera (MF9), sample no. A59. **B.** Silty marl with planktonic and benthic foraminifera (MF9), sample no. A55. **C.** Silty marl with clay minerals and fecal pellets (MF10), sample no. A51. **D.** Silty marl with clay minerals and fecal pellets (MF10), sample no. A51. **E.** Clay minerals just above the K/Pg boundary, sample no. A51. **F.** Photomicrographs of the clay minerals, which were hand-picked from the residue of the washed samples, sample no. A51. **G.** Echinoid fecal pellet, sample no. A51. **H.** Photomicrographs of an echinoid fecal pellet, sample no. A51. **I.** Scanning electron microscope image of an echinoid fecal pellet, sample no. A51. (Abbreviations: hb: hyaline benthic foraminifera, ab: agglutinated benthic foraminifera, p: planktonic foraminifera, q: quartz, c: clay minerals. Scale bar is 0.25 mm for A, B, C, D, E, G and 100 μm for H, I).



Fig. 2.11. Lithostratigraphy, cumulative percentage diagram, and sequence stratigraphic interpretation of the Campo Pit section, New Jersey, USA and relative sea-level curve comparison of the section to Haymana Basin, Turkey. Constructions of the relative sea-level curves are explained in the text. Inset shows the geological map of New Jersey with the location of Campo Pit outcrop. Refer to the legend in Fig. 2.2 for the Haymana Basin section. LST: lowstand systems tract, HST: highstand systems tract, TST: transgressive systems tract, SB: sequence boundary, TS: transgressive surface, MFS: maximum flooding surface.

Contusotruncana contusa	Contusotruncana walfischensis	Globigerinelloides alvarezi	Globigerinelloides multispinus	Globigerinelloides prairiehillensis	Globigerinelloides sp.	Globiaerinelloides subcarinatus	Globotruncana aegovtiaca	Globotruncana arca	Globotruncana dupeublei	Globotruncana esnehensis	Globotruncana falsostuarti	Globotruncana hilli	Clobotrucona marial	Globotruncana onentalis	Giobotruncana sp.	Globotruncanella navanensis	Globotrancanella minuta	Globotruncanella petaloidea	City for the second sp.	Globotruncanita angulata Globotruncanita conica	Globotruncanita pettersi	Globotruncanita sp.	Globotruncanita stuarti	Globotruncanita stuartiformis	Hedbergella holmdelensis	Hedbergella monmouthensis	Heddergella sp.	Heterohelix labellosa	Heterohelix navarroensis	Heterohelix planata	Heterohelix punctulata	Heterohelix sp.	Laeviheterohelix dentata	Laeviheterohelix glabrans	Laeviheterohelix sp.	Planoglobulina acervulinoides	Planoglobulina carseyae	Pseudoguembelina costulata	Pseudoguembelina excolata	Pseudoguembelina hariaensis	Pseudoguembelina sp.	Pseudotextularia elegans	Pseudotextularia nuttalli	Pseudotextularia so.	Racemiguembelina fructicosa	Racemiquembelina powelli	Racemiguembelina sp.	Rugoglobigerina hexacamerata	Rugoglobigerina macrocephala	Rugoglobigerina milamensis	Rugoglobigerina pennyi	Rugoglobigerina rugosa	Rugoglobigerina sp.	Guembelitria cretacea	Chiloguembelina midwayensis	Chiloguembelina morsei	Ecolobiaerina ecolulloides	Globanomalina aff. planocompressa	Globanomalina archeocompressa	Globoconusa daubjergensis	Parasubbotina pseudobulloides	Parvularugoglobigerina eugubina	Praemurica pseudoinconstans	Preemunca taunca Subbotina trivialis	Woodringing clavtonensis	Woodringing homerstownensis	Zeauvigerina waiparaensis		SAMPLE NUMBER			BIOZONES	AGE	
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Table 2.1. Distribution chart of planktonic foraminifera in the Haymana section across the K/Pg boundary.

Chapter 3

Stratigraphy and Sea-Level Changes across the Cretaceous/Paleogene Boundary in the Campo Pit Section, New Jersey, USA

3.1 Abstract

Litho- and sequence stratigraphic analyses of an outcrop section at Campo Pit, Perrineville, New Jersey was conducted to evaluate the sea-level changes across the Cretaceous/Paleogene (K/Pg) boundary. There is a sequence boundary (SB) between the upper shoreface Redbank and middle to lower shoreface Tinton Formations (base of Navesink II sequence) in the late Maastrichtian. Above the SB there is a lag deposit assigned to the Redbank Formation and a transgressive systems tract (TST) starts above the lag deposits in the Tinton Formation. The TST continues across the K/Pg boundary into the lower Hornerstown Formation. The K/Pg boundary at Campo Pit is immediately above a reddish, indurated clayey sand unit of the uppermost Tinton Formation. The ~50cm-thick event bed above the K/Pg boundary consists of lower red and upper green clay with clay clasts and impact spherules. The event bed in the adjacent Buck Pit corehole records a 0.5 ppb iridium anomaly. The K/Pg boundary lies below the maximum flooding surface, which is associated with an increase in authigenic glauconite percentage in the lower Hornerstown Formation. Comparison of the section with previous records in the New Jersey Coastal Plain and several other K/Pg boundary sections in the world suggest that the K/Pg event occurred during a global sea-level rise.

3.2 Introduction

Sea-level changes across the Cretaceous/Paleogene (K/Pg) boundary were small compared to the Campanian/Maastrichtian boundary and later Paleogene sea-level fluctuations (Miller et al., 2005; Haq, 2014) and did not play an important role in the end-Cretaceous mass extinction (Hallam and Wignall, 1999). However, it is important to understand how sea level changed during the K/Pg transition to evaluate the local depositional evolutions of the basins, as well as global stratigraphy of the K/Pg boundary.

Sea-level studies from various locations worldwide have yielded different eustatic interpretations across the K/Pg boundary. Claeys et al. (2002) compiled sea-level studies from almost 100 global shallow-marine sites and compared the sites showing regression versus no regression or transgression (their Fig. 3). The compilation of Claeys et al. (2002) shows that the number of studies reporting regression versus no regression/transgression is almost the same. Various sea-level interpretations arise from the different methodologies used and the fact that the sea-level changes across K/Pg boundary were subtle.

New Jersey Coastal Plain (NJCP) results using onshore Ocean Drilling Program Leg 174AX coreholes (Fig. 3.1) indicated a minimal fall (< 25m) across the K/Pg boundary (Miller et al., 2005; Kominz et al., 2008), though outcrop sections yielded two possibilities for the sea-level changes: 1) a gradual fall across the boundary into the earliest Danian, and then a rise within the early Danian; or 2) a gradual rise starting in the late Maastrichtian and continuing into the early Danian. This study aims to test these two possibilities using detailed lithostratigraphic and sequence stratigraphic analyses of outcrops at Campo Pit, Perrineville, New Jersey (Fig. 3.1). The Campo Pit outcrop is an ideal section to test relative sea-level changes because it shows distinct facies variations across the K/Pg boundary due to its shoreface setting. This contrasts with middle to outer neritic sections such as Bass River and Ancora that are not sensitive to small sea-level variations at the K/Pg boundary due to their deeper water settings. We aim to develop a relative sea-level curve for the Campo Pit section and compare it with sections elsewhere to understand the global sea-level changes across the K/Pg boundary.

3.2.1 Geological Setting of the New Jersey Coastal Plain

The mid-Atlantic coast of the United States (New Jersey, Delaware, Maryland, North Carolina) is a classic passive continental margin formed after the rifting and seafloor spreading between North America and Africa in the Late Triassic to Early Jurassic (Grow and Sheridan, 1988). After the rifting stage, thermal cooling and subsidence, sediment loading and flexure dominated accommodation (Watts and Steckler, 1979; Steckler et al., 1999). The large sedimentary basin that underlies the continental shelf along the mid-Atlantic coast is called the Baltimore Canyon Through (Grow and Sheridan, 1988). Shallow-water limestones and shales dominated the sediments of Baltimore Canyon in the Jurassic. Until the mid-Cretaceous a barrier reef was formed in the region and during the Late Cretaceous to Paleogene siliciclastic and carbonate deposition took place with low accumulation rates (Poag, 1985). A major change from carbonate ramp deposition to siliciclastic deposition occurred in the middle Eocene to early Oligocene (Miller et al., 1997) and sedimentation rates continued to increase in the late Oligocene to Miocene (Poag, 1985; Miller et al., 1997). The NJCP forms the western margin of the Baltimore Canyon and consists of Lower Cretaceous to Holocene sediments that dip gently seaward (Olsson et al., 1988).

The NJCP region evolved through million-year scale depositional phases during the Early Cretaceous to Recent (Browning et al., 2008). General trends are: fluvial to marine deposition with deltaic influence in the early–mid Cretaceous; starved marine ramp from Maastrichtian to middle Eocene; starved siliciclastic shelf from late middle Eocene to Oligocene; marine shelf with wave-dominated delta influence in early and middle Miocene; and eroded coastal system from late Miocene to Recent (Browning et al., 2008).

3.2.2 Upper Cretaceous to Lower Paleogene Stratigraphy

The Maastrichtian and lower Paleocene strata are characterized by glauconite, carbonate, and clay deposition. Sequences consist of (1) a basal unconformity; (2) a thin lower glauconitic sand (transgressive systems tract/TST); and (3) a coarsening-upward regressive succession of medial silts and upper quartz sands (highstand systems tract/HST) (Sugarman et al., 1995; Miller et al., 2003). The only lowstand systems tract (LST) identified in the NJCP from this interval is at the base of the Navesink Formation in the Maastrichtian (Miller et al., 1999).

Early studies documented 5 - 8 Late Cretaceous transgressive-regressive cycles in the NJCP using outcrops and coreholes based on regular glauconitic beds, physical unconformities, and biostratigraphic hiatuses (Olsson, 1963; Owens and Gohn, 1985; Olsson et al., 1988; Olsson, 1991). These cycles were attributed to tectonic control on sedimentation (Owens and Sohl, 1969) or eustatic changes (Olsson et al., 1988; Olsson, 1991). Sugarman et al. (1995) integrated Sr-isotope stratigraphy, magnetostratigraphy, and calcareous nannoplankton biostratigraphy to provide chronology for the sequences. After integrating the data from the onshore ODP Leg 174AX (Bass River, Ancora, Millville, and Sea Girt coreholes, Fig. 3.1), Miller et al. (2003; 2004) identified 11 - 14 Upper Cretaceous sequences in the coastal plain and constraint the ages using biostratigraphy and Sr-isotopic stratigraphy, and genetically linked them to eustasy.

Maastrichtian and Danian strata are described in Owens and Sohl (1969), Sugarman et al. (1995), Olsson et al. (2002), Miller et al. (2004), and Landman et al. (2004). Sugarman et al. (1995) defined only one Maastrichtian sequence and termed it the Navesink sequence (Fig. 3.2). The Navesink Formation, the lowermost formation in the Navesink sequence, consists of carbonate rich foraminiferal clayey glauconites deposited in middle-shelf environments (Sugarman et al., 1995; Miller et al., 2004). The Navesink Formation unconformably overlies the Mount Laurel Formation of the Campanian Marshalltown sequence. The prominent Marshalltown/Navesink sequence boundary (SB) represents a major hiatus of 2.2 myr and is associated with a layer of phosphate pebbles (Miller et al., 2003; 2004). Above the SB a thin (~ 1 m) lag unit containing reworked sands of Mount Laurel lithology may represent a LST (Miller et al., 1999; 2004). The lag unit is overlain by a unit in which quartz content decreases up section, suggesting a deepening and assigned to a TST (Sugarman et al., 1995; Miller et al., 2004). The Redbank Formation, overlying the Navesink Formation, is divided into two members, the Sandy Hook Member below and the Shrewsbury Member above (Olsson, 1960) (Fig.

3.2). The Sandy Hook Member consists of fossiliferous clayey silt and fine sand representing distal to lower shoreface environments and the Shrewsbury Member is unfossiliferous, burrowed, medium to coarse quartz sand representing an upper to lower shoreface environment (Sugarman et al., 1995). The Tinton Formation, the uppermost Maastrichtian unit in New Jersey outcrops (but not subsurface), is generally an indurated clayey glauconitic quartz sand representing an inner shelf environment (Sugarman et al., 1995; Landman et al., 2004). Since the Tinton Formation is siderite cemented and iron oxide crusted it is difficult to observe the contact with the Redbank Formation; however, it is considered to be disconformable (Sugarman et al., 1995; Landman et al., 2004).

There is an alternative nomenclature for the area. Olsson (1960; 1963) named the more clay-rich downdip equivalent of the Navesink, Redbank, Tinton, and Hornerstown Formations as the New Egypt Formation (Fig. 3.2). Later on Olsson (1987) restricted the New Egypt Formation to only deep-water facies equivalent to the Tinton and Redbank Formations (Fig. 3.2). It consists of brownish-gray glauconitic clayey sand to sandy clay with siderite nodules (Landman et al., 2004; Miller et al., 2004). The upper Maastrichtian Redbank and Tinton Formations (the New Egypt Formation) are considered a HST in Sugarman et al. (1995) and Olsson et al. (2002). Later, Miller et al. (2004) described the upper Redbank Formation as HST, but indicated a TST in the Tinton Formation with a question mark (Fig. 3.2). Above the K/Pg boundary, lies burrowed, clayey glauconite sand of the Hornerstown Formation (Sugarman et al., 1995; Landman et al., 2004; Miller et al., 2004). High glauconite abundance gives a dark greenish color to the formation and thus it is also called "Greensand Marl" informally (Landman et al., 2004). The base of the Hornerstown Formation contains a very concentrated fossil bed known as the 'Main

Fossiliferous Layer' (MFL) at Sewell, New Jersey (Gallagher, 1993). There are two different opinions about the contact between the Hornerstown and the Tinton (the New Egypt) Formations. Sugarman et al. (1995) argued that the Hornerstown unconformably overlies the upper Maastrichtian units, whereas Olsson et al. (2002) and Miller et al. (2004) interpreted a conformable contact between these formations with a complete K/Pg boundary (Fig. 3.2). Miller et al. (2003; 2004) reconsidered the Maastrichtian sequences and identified a second SB at the Navesink/New Egypt Formation contact (Fig. 3.2). The Navesink I sequence is separated by this SB from the overlying Navesink II sequence. The Navesink II sequence continues across the K/Pg boundary into the lower Paleocene where there is a prominent unconformity within the lower Hornerstown Formation. Foraminiferal Zone P1b is absent at Bass River and not complete at Ancora (Olsson et al., 2002; Miller et al., 2004) indicating a hiatus of ~0.5 Myr associated with this unconformity.

3.2.3 K/Pg Boundary Sections in New Jersey

Several corehole and outcrop sections in New Jersey preserve the K/Pg boundary. Coreholes having the K/Pg boundary were drilled by onshore ODP Leg 174AX (Bass River, Ancora, Millville, Sea Girt), and U.S. Geological Survey (Fig. 3.1). The most complete K/Pg boundary section in the NJCP described so far is in the Bass River core (Olsson et al., 1997; 2002). Planktonic foraminiferal, nannofossil, and dinoflagellate biostratigraphy delineated the K/Pg boundary. A 6-cm-thick spherule layer is just above the boundary consisting of common shocked minerals and altered microtektites of 175 – 1100 µm diameter. The New Egypt Formation just below the spherule layer is assigned to the *Abathomphalus mayaroensis* foraminifera, *Palynodinium grallator* dinocyst, and *Micula prinsii* nannofossil zones corresponding to uppermost Maastrichtian. The Hornerstown Formation immediately above the spherule layer includes the lowermost Danian planktonic foraminifera Zones P0 and P α and also dinoflagellate index fossil *Senoniasphaera inornata* (Olsson et al., 1997; 2002). A large iridium (Ir) anomaly of 2.4 ppb is found at the base of the spherule layer and a smaller Ir anomaly of ~0.7 ppb at the top of the spherule layer in the basal Danian, where there are white clay rip up clasts containing Cretaceous fauna (Olsson et al., 1997; 2002; Miller et al., 2010).

Spherule impressions into Cretaceous sediments show that the sediments were soft and unlithified at the time the spherule layer was deposited by the fallout of impact ejecta and is not reworked. On the other hand, the white clay rip-up clasts in the basal Danian with reworked Cretaceous fossils might have been deposited by a mega-tsunami inferred to have been generated by the impact (Smit, 1999). However, it has also been suggested that the Florida and Bahama platforms prevented tsunamis from spreading from the Gulf of Mexico into the western North Atlantic (Norris et al., 2000). Alternatively, the white clay clasts might be explained by slope failure triggered by earthquakes created by the Chicxulub impact causing erosion and redeposition of outer shelf Cretaceous clays on the inner to middle shelf (Norris et al., 2000; Olsson et al., 2002). The clay clast layer above the boundary is found in many other onshore New Jersey ODP sites and outcrops such as Ancora, Millville, Sea Girt, and Sewell but they generally lack distinct spherule layers and shocked minerals reflecting the heterogeneity of preservation of the spherules on the paleoshelf (Miller et al., 2010). In several shallow NJCP cores (< 25 m) adjacent to outcrops of the K/Pg boundary (Tighe Park 1, Search Farm 1, Meirs Farm 1, and Buck Pit 1) (Fig. 3.1) Ir concentrations were measured (Miller et al., 2010). These sections have Ir anomalies of ~0.5 ppb showing that Ir is in place (i.e. correlatable with the extinction level) in clayrich sections and displaced in sandier sections up or down based on the redox conditions (Colodner et al., 1992). (Chapter 5 includes detailed discussion on Ir anomaly across the K/Pg boundary in NJCP cores).

Another remarkable observation in NJCP K/Pg boundary cores is the abrupt increase in echinoid fecal pellets above the extinction level. Miller et al. (2010) showed that in several K/Pg boundary sections the number of echinoid fecal pellets per gram of sample increases from 0 to 10–30 in the first sample of the Danian. The echinoid fecal pellets in the section are ~1 mm long, orange to brown ellipsoidal grains with smooth surface, occasionally with thin laminations (Fig. 3.3). They were assigned to Echinodermata, especially of Echinoidea by Voigt (1929) and previously described in the lowest Paleocene in Denmark and Sweden in large quantities (Brotzen, 1948); their pl. 2, fig. 2). We showed the same increase in echinoid fecal pellets (from 0 to 12 pellets/gram) across the K/Pg boundary in the Haymana Basin, Turkey (Esmeray-Senlet et al., in review; Chapter 2). We suggest that this increase might indicate an important environmental change after the K/Pg mass extinction and can be used as a global marker for the K/Pg boundary in addition to other criteria defining the boundary.

3.3 Methods

A 6.5-m-thick section at Campo Pit (Fig. 3.4) was sampled (15 samples) for litho-, and sequence stratigraphic analyses. Samples were disaggregated using Calgon solution (5.5 g of sodium metaphosphate per 4 liters of water) and washed with tap water through a 63- μ m sieve. Dry weights before and after washing were noted to calculate the sand/mud ratio of the samples. After drying in an oven at 40°C, samples were sieved through 250, 150, 125, and 63 μ m sieves and each size fraction was examined separately for mineralogical analysis. Semi-quantitative cumulative percentage analysis of minerals was conducted by visual estimation for the medium (> 250 μ m) and fine (< 250 μ m) sand fractions separately.

3.4 Results

3.4.1 Lithostratigraphy

The upper Redbank Formation at Campo Pit (0 –1.0 m; Figs. 3.4, 3.5) is consists of white to yellowish, medium to coarse grained, well-sorted sand, rich in vertical and horizontal *Ophiomorpha* burrows (Fig. 3.5). Samples 1 and 2 contain more than 95% angular quartz grains with few opaque heavy minerals (OHMs), and brown glauconite (Fig. 3.4). Most of the quartz grains are transparent, although yellow-brown stained quartz grains are also common. Iron staining is due to recent ground water activity. There is an unconformity above sample 2 at ~1.0 m represented by an irregular erosional surface (Fig. 3.4). Above the unconformity there is a lag deposit (1.0 –1.4 m; Sample 3) consisting of white to yellowish, medium to coarse grained, well-sorted sand with increased brown glauconite content compared to samples 1 and 2.

The lower Tinton Formation (1.4 - 3.85 m) consists of yellowish to brownish, medium to coarse grained, well-sorted sand with some laminations. Sample 4 (1.65 m)consists of almost entirely of transparent, angular quartz grains with very few OHMs. The amount of mud, fine quartz, and glauconite increases upsection in the Tinton Formation (see cumulative percentage plot, Fig. 3.4) and light green glauconite starts to appear (Sample 5–7). The uppermost part of the Tinton Formation (3.85 - 3.95 m) is a reddish, indurated unit. The K/Pg boundary is placed at the top of the indurated unit that is lithologically distinct from either the Tinton Formation below or The Hornerstown Formation above. The K/Pg boundary is overlain by a clayey unit, which we define as the event bed. The event bed (3.95 - 4.5 m) consists of a ~15 cm-thick bed of red clay (3.95 - 4.5 m)-4.1 m; Sample 8) and a ~40 cm-thick bed of green clay (4.1 - 4.5 m; Sample 9) with clay clasts containing impact spherules and spherule imprints (Fig. 3.6). Spherules are rarely preserved in the clay clasts, but white, spherical-elliptical imprints of spherules are very common (Fig. 3.6). Samples from the red clay are unfossiliferous. The lowest occurrence of *Senoniasphaera inornata*, a dinoflagellate marker species representing the earliest Danian, is at the base of the green clay (Miller et al., 2010), which is possibly a late first occurrence. Above the event bed the Hornerstown Formation (4.5 - 6 m; Sample)12 - 14) is a clayey, green glauconite sand with quartz grains. The Hornerstown Formation is overlain by yellowish, heavily burrowed, medium grained sand of the Vincentown Formation (6 - 6.5 m).

The nearby Buck Pit corehole shows an Ir anomaly of ~ 0.5 ppb in the green clay (Miller et al., 2010). Neither the Buck Pit core nor Campo Pit section have echinoid fecal pellets. The absence of the fecal pellets at Campo Pit and the adjacent Buck Pit core (Miller et al., 2010) might suggest facies control due to post-depositional alteration in shoreface settings.

3.4.2 Sequence Stratigraphy and Sea-Level Changes

The uppermost part of the Redbank Formation contains abundant *Ophiomorpha* burrows (Fig. 3.5). The substrate for *Ophiomorpha* burrows consists of shifting particles of clean, well-sorted sand and they are considered part of the *Skolithos* ichnofacies indicative of relatively high levels of wave or current energy (Catuneanu, 2006 and the references therein). Vertical and horizontal *Ophiomorpha* burrows and planar laminations in the Redbank Formation represent an upper shoreface environment above the fair weather wave base (FWWB) (Fig. 3.4).

There is a SB above the Redbank Formation (Fig. 3.4). The lag deposits of the Redbank Formation above the SB are interpreted as a LST. Above the lag deposits the Tinton Formation represents a gradual transition from middle shoreface to lower shoreface environments indicated by a slight fining upward succession with increased mud and glauconite content (see cumulative percentage plot; Fig. 3.4) interpreted to represent a TST. Transgression continued during the deposition of the reddish indurated unit of the uppermost Tinton Formation, the K/Pg boundary, and the event bed. Above the event bed, clayey green glauconitic sand of the Hornerstown Formation indicates an

offshore environment below storm wave base. Glauconite sand of the Hornerstown Formation consists of authigenic, dark green glauconite with high maturity (Wahyudi, 2010). Authigenic glauconite is common in the TST and the lower HST; its abundance shows an upward increase in the TST and then a decrease in the HST (Amorosi, 1995). A MFS is associated with the maximum percentage of glauconite in the Hornerstown Formation (4.75 m; Sample 12). In the HST, above the MFS a gradual increase in quartz indicates shallowing. For the paleowater depth curve of Campo Pit we took upper shoreface/middle shoreface boundary (FWWB) as between 10–15 m, middle shoreface/lower shoreface boundary as 20 m, and lower shoreface/offshore boundary (SWB) as 30 m (Fig. 3.4).

3.5 Discussion

3.5.1 Sea-Level Changes in New Jersey Coastal Plain

Previous studies (Olsson et al., 2002; Miller et al., 2004; Van Sickel et al., 2004; Miller et al., 2005; Kominz et al., 2008; Browning et al., 2008) identified three major sealevel falls during the late Cretaceous to early Paleogene: 1) at the Campanian/Maastrichtian boundary, between the Mount Laurel and Navesink Formations, base of the Navesink I sequence (Fig. 3.7; (Miller et al., 2004; 2005); 2) in the late Maastrichtian, between the Redbank and Tinton Formations, base of the Navesink II sequence (Fig. 3.7; (Miller et al., 2005); and 3) in the Danian, at the base of planktonic foraminiferal Zone P1b (Olsson et al., 2002), base of the Pa1a sequence (Harris et al., 2010).

The K/Pg boundary is within the Navesink II sequence. Benthic foraminiferal biofacies (Olsson et al., 2002) and the backstripping results from the Ancora, Bass River, Millville and Sea Girt sections (Miller et al., 2005; Kominz et al., 2008) showed a minimal fall in sea level (< 25 m) associated with the K/Pg boundary. Miller et al. (2005) showed a gradual fall during the entire Navesink II sequence. On the other hand, Kominz et al. (2008) observed a stepwise rise at the beginning of the Navesink II sequence until just prior to the K/Pg boundary and a gradual fall across the K/Pg boundary. All studies agreed that sea-level change at the boundary is minimal and did not cause a sequence boundary.

Previous studies provide two possibilities for the sea-level changes across the K/Pg boundary: 1) a gradual fall across the boundary into the lowermost Hornerstown Formation, and then a rise within the lower Hornerstown Formation; or 2) a gradual rise starting in the Tinton Formation and continuing into the lower Hornerstown Formation (Fig. 3.7). New data from the Campo Pit section are more consistent with the second possibility and show that the K/Pg boundary occurred during a sea-level rise in New Jersey.

3.5.2 Global Sea-Level Changes

Although studied by numerous authors in several K/Pg boundary sections worldwide, there is no consensus on global sea-level changes across the K/Pg boundary,
likely due to the fact that the changes were subtle. Sea-level changes across the boundary interpreted by Exxon Production Research Company (EPR) showed that the end of the Cretaceous was associated with a major fall, on the order of 150–200 m, which was followed by a gradual rise in the early Paleogene (Vail et al., 1977). Hag et al. (1987) subsequently showed a sharp global sea-level fall of 100 m well below the K/Pg boundary, followed by a rapid rise of 75 m, and another minor fall of 10–20. This minor fall is followed by a 25 m sea-level rise from the K/Pg boundary upwards. Even if the sea-level trends determined by EPR are correct, the amplitudes they suggest are too high for this time interval (Miller et al., 2005; Kominz et al., 2008). Another EPR study by Donovan et al. (1988) showed a major fall in the late Maastrichtian and a sea-level rise across the K/Pg boundary in Braggs section, Alabama. Olsson and Liu (1993) made important biostratigraphic revisions to the Braggs section (Donovan et al., 1988) and moved the boundary to a lower stratigraphic position. Accordingly, Olsson and Liu (1993) stated that sea level was at lowstand both before and after the boundary with a rise afterwards.

Coarse-grained siliciclastic and limestone breccia deposits at the K/Pg boundary close to the impact site such as the Gulf of Mexico, Mexico, Guatemala, the Caribbean, and Brazil have been interpreted as pre K/Pg sea-level lowstand deposits (Donovan et al., 1988; Keller and Stinnesbeck, 1996; Keller, 2007). However, this interpretation has been challenged by the fact that there are impact-generated tsunami deposits in these regions (Smit, 1999). The siliciclastic and limestone breccia deposits were likely deposited by debris flows and thus have no relation to eustasy (Claeys et al., 2002).

Keller and Stinnesbeck (1996) provided a global sea-level review across the K/Pg boundary with a wide range of marine paleoenvironments worldwide: near shore to neritic, middle and outer shelf, continental slope, and bathyal. Their quantitative studies of benthic and planktonic foraminifera, spores and pollen, dinoflagellates, and macrofossils showed a consistent pattern of rise across the K/Pg boundary.

European and Tunisian sections show similar sea-level changes across the K/Pg boundary in a number of locations. In the Basque Region, a SB below the boundary was identified, with the K/Pg boundary placed within the overlying TST (Apellaniz et al., 1997; Pujalte, 1998). Alegret et al. (2003) and Pardo et al. (1996) in Spain, and Pardo et al. (1999) in Kazakhstan showed a sea-level rise across the K/Pg boundary. The Tunisian record (Adatte et al., 2002) also showed a global sea-level rise across the K/Pg boundary into the Danian. Our detailed microfacies and sequence stratigraphic analyses work in the Haymana Basin, Turkey also indicated a sea-level rise across the K/Pg boundary in agreement with other European and Tunisian sections (Esmeray-Senlet et al., in review; Chapter 2). We observe a very similar trend in the Campo Pit section, where the K/Pg boundary is placed in a late TST below a MFS.

3.6 Conclusions

Lithostratigraphic analysis of the updip Campo Pit section showed that there is an event bed above the K/Pg boundary consisting of red and green clay layers with clay clasts and impact spherules. This level also shows a ~0.5 ppb Ir anomaly at the Buck Pit corehole adjacent to the Campo Pit outcrop. Sequence stratigraphic analysis and a sea-

level curve of the Campo Pit section based on lithofacies analysis indicate that a SB exists between the Redbank and Tinton Formations (base of the Navesink II sequence) and the K/Pg boundary lies in the TST. This trend across the K/Pg boundary is very similar to the sea-level records of European and North African sections suggesting that the K/Pg impact occurred during a global sea-level rise.

- Adatte, T., Keller, G., and Stinnesbeck, W., 2002, Late Cretaceous to early Paleocene climate and sea-level fluctuations: the Tunisian record: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 178, p. 165–196.
- Alegret, L., Molina, E., and Thomas, E., 2003, Benthic foraminiferal turnover across the Cretaceous/Paleogene boundary at Agost (southeastern Spain): paleoenvironmental inferences: Marine Micropaleontology, v. 48, p. 251–279.
- Amorosi, A., 1995, Glaucony and Sequence Stratigraphy: A Conceptual Framework of Distribution in Siliciclastic Sequences: Journal of Sedimentary Research, v. 65, p. 419–425.
- Apellaniz, E., Baceta, J.I., Bernaola-Bilbao, G., Nunez-Betelu, K., Orue-Etxebarria, X., Payros, A., Pujalte, V., Robin, E., and Rocchia, R., 1997, Analysis of uppermost Cretaceous-lowermost Tertiary hemipelagic successions in the Basque Country (western Pyrenees); evidence for a sudden extinction of more than half planktic foraminifer species at the K/T boundary: Bulletin de la Societe Geologique de France, v. 168, p. 783–793.
- Brotzen, F., 1948, The Swedish Paleocene and Its Foraminiferal Fauna: Sweriges Geologiska Undersökning, v. 42, p. 1–140.
- Browning, J.V., Miller, K.G., Sugarman, P.J., Kominz, M.A., McLaughlin, P.P., Jr, Kulpecz, A.A., and Feigenson, M.D., 2008, 100 Myr record of sequences, sedimentary facies and sea level change from Ocean Drilling Program onshore coreholes, US Mid-Atlantic coastal plain: Basin Research, v. 20, p. 227–248, doi: 10.1111/j.1365-2117.2008.00360.x.
- Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of Chicxulub ejecta at the Cretaceous-Tertiary boundary (C. Koeberl & K. G. MacLeod, Eds.): Catastrophic Eevents and Mass extinctions: Impacts and Beyond: Geological Society of America Special Paper, v. 356, p. 55–68.
- Colodner, D., Boyle, E., Edmond, J., and Thomson, J., 1992, Post-depositional mobility of platinum, iridium and rhenium in marine sediments: Nature, v. 358, p. 402–404.
- Donovan, A.D., Baum, G.R., Blechschmidt, G.L., Loutit, T.S., Pflum, C.E., and Vail, P.R., 1988, Sequence stratigraphic setting of the Cretaceous-Tertiary boundary in Central Alabama. In Wilgus, C.K., Kendall, C.G.S.C., Posamentier, H.W., and Ross, C.A., eds., Sea-level changes: An integrated approach: Society of Economic Paleontologists and Mineralogists Special Publication, v. 42, p. 299–307.
- Gallagher, W.B., 1993, The Cretaceous/Tertiary mass extinction event in the northern Atlantic Coastal Plain: The Mosasaur, v. 5, p. 75–155.

- Gradstein, F.M., Ogg, J.G., Schmitz, M., and Ogg, G., 2012, The Geologic Time Scale 2012 2-Volume Set: Elsevier BV, v. 1-2, p. 1176.
- Grow, A.J., and Sheridan, R.E., 1988, U.S. Atlantic continental margin; a typical Atlantic-type or passive continental margin: in, Sheridan, R.E., and Grow, J.A., Eds., The Atlantic continental margin, U.S.: Boulder, Colorado, Geological Society of America, Geology of North America, v. I, p. 1–7.
- Hallam, A., and Wignall, P., 1999, Mass extinctions and sea-level changes: Earth-Science Reviews, v. 48, p. 217–250.
- Haq, B.U., 2014, Cretaceous eustasy revisited: Global and Planetary Change, v. 113, p. 44–58, doi: 10.1016/j.gloplacha.2013.12.007.
- Harris, A.D., Miller, K.G., Browning, J.V., Sugarman, P.J., Olsson, R.K., Cramer, B.S., and Wright, J.D., 2010, Integrated stratigraphic studies of Paleocene–lowermost Eocene sequences, New Jersey Coastal Plain: Evidence for glacioeustatic control: Paleoceanography, v. 25, p. PA3211, doi: 10.1029/2009PA001800.
- Keller, G., 2007, The Chicxulub impact and K-T mass extinction in Texas: Bulletin of South Texas Geological Society, v. XLVII, p. 15–44.
- Keller, G., and Stinnesbeck, W., 1996, Sea-level changes, clastic deposits, and megatsunamis across the Cretaceous–Tertiary boundary: In: MacLeod, N., Keller, G. (Eds.), Cretaceous–Tertiary Mass Extinctions: Biotic and Environmental Change. Norton, London, p. 415–449.
- Kominz, M.A., Browning, J.V., Miller, K.G., Sugarman, P.J., Mizintseva, S., and Scotese, C., 2008, Late Cretaceous to Miocene sea-level estimates from the New Jersey and Delaware coastal plain coreholes: an error analysis: Basin Research, v. 20, p. 211–226, doi: 10.1111/j.1365-2117.2008.00354.x.
- Landman, N., Johnson, R., and Edwards, L., 2004, Cephalopods from the Cretaceous/Tertiary boundary interval on the Atlantic Coastal Plain, with a description of the highest ammonite zones in North America. Part 1. Maryland and North Carolina: American Museum Novitates, v. 3454, p. 1–64.
- Miller, K.G., Barrera, E., Olsson, R.K., Sugarman, P.J., and Savin, S.M., 1999, Does ice drive early Maastrichtian eustasy?: Geology, v. 27, p. 783–786.
- Miller, K.G., Browning, J.V., Pekar, S.F., and Sugarman, P.J., 1997, Cenozoic evolution of the New Jersey coastal plain: changes in sea level, tectonics, and sediment supply: In: Miller, KG, and Snyder, SW (Eds), 1997 Proceedings of the Ocean Drilling Program, Scientific Results, v. 150X, p. 361–373.

- Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., and Pekar, S., 2005, The Phanerozoic record of global sea-level change: Science, v. 310, p. 1293–1298, doi: 10.1126/science.1116412.
- Miller, K.G., Sherrell, R., Browning, J.V., Field, M., Gallagher, W., Olsson, R.K., Sugarman, P.J., Tuorto, S., and Wahyudi, H., 2010, Relationship between mass extinction and iridium across the Cretaceous-Paleogene boundary in New Jersey: Geology, v. 38, p. 867–870, doi: 10.1130/G31135.1.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Kominz, M.A., Hernandez, J., Olsson, R.K., Wright, J.D., Feigenson, M.D., and Van Sickel, W., 2003, Late Cretaceous chronology of large, rapid sea-level changes: Glacioeustasy during the greenhouse world: Geology, v. 31, p. 585–588.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Kominz, M.A., Olsson, R.K., Feigenson, M.D., and Hernandez, J., 2004, Upper Cretaceous sequences and sea-level history, New Jersey coastal plain: Geological Society of America Bulletin, v. 116, p. 368– 393.
- Norris, R.D., Firth, J., Blusztajn, J., and Ravizza, G., 2000, Mass failure of the North Atlantic margin triggered by the Cretaceous-Paleogene bolide impact: Geology, v. 28, p. 1119–1122.
- Olsson, R.K., 1987, Cretaceous stratigraphy of the Atlantic Coastal Plain, Atlantic Highlands of New Jersey: Geological Society of America Centennial Field Guide-Northeastern Section, p. 87–90.
- Olsson, R.K., 1991, Cretaceous to Eocene sea-level fluctuations on the New Jersey margin: Sedimentary Geology, v. 70, p. 195–208, doi: 10.1016/0037-0738(91)90141-Y.
- Olsson, R.K., 1960, Foraminifera of latest Cretaceous and earliest Tertiary age in the New Jersey Coastal Plain: Journal of Paleontology, v. 34, p. 1–58.
- Olsson, R.K., 1963, Latest Cretaceous and Earliest Tertiary Stratigraphy of New Jersey Coastal Plain: American Association of Petroleum Geologists Bulletin, v. 47, p. 643– 665.
- Olsson, R.K., and Liu, C., 1993, Controversies on the placement of Cretaceous-Paleogene boundary and the K/P mass extinction of planktonic foraminifera: Palaios, v. 8, p. 127–139.
- Olsson, R.K., Gibson, T.G., Hansen, H.J., and Owens, J.P., 1988, Geology of the northern Atlantic Coastal Plain: Long Island to Virginia: In: Sheridan, R.E, and Grow, J.A., Eds. The Atlantic continental margin, US: Boulder, Colorado, Geological Society of America, Geology of North America, v. I, p. 87–105.

- Olsson, R.K., Miller, K.G., Browning, J.V., Habib, D., and Sugarman, P.J., 1997, Ejecta layer at the Cretaceous-Tertiary boundary, Bass River, New Jersey (Ocean Drilling Program Leg 174AX): Geology, v. 25, p. 759–762.
- Olsson, R.K., Miller, K.G., Browning, J.V., Wright, J.D., and Cramer, B.S., 2002, Sequence stratigraphy and sea-level change across the Cretaceous-Tertiary boundary on the New Jersey passive margin: Catastrophic events and Mass extinctions: impacts and beyond, v. 356, p. 97–108.
- Owens, J.P., and Gohn, G.S., 1985, Depositional history of the Cretaceous Series in the U.S. Atlantic Coastal Plain: Stratigraphy, paleoenvironments, and tectonic controls of sedimentation: In Poag, C. W., ed., Geologic evolution of the United States Atlantic margin: New York, Van Nostrand Reinhold,, p. 25–86.
- Owens, J.P., and Sohl, F.N., 1969, Shelf and deltaic paleoenvironments in the Cretaceous-Tertiary formations of the New Jersey Coastal Plain: In Subitsky, S., ed., Geology of selected areas in New Jersey and eastern Pennsylvania and guide-book of excursions: New Brunswick, New Jersey, Rutgers University Press, p. 235–278.
- Pardo, A., Adatte, T., Keller, G., and Oberhansli, H., 1999, Paleoenvironmental changes across the Cretaceous-Tertiary boundary at Koshak, Kazakhstan, based on planktic foraminifera and clay mineralogy: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 154, p. 247–273.
- Pardo, A., Ortiz, N., and Keller, G., 1996, Latest Maastrichtian and Cretaceous-Tertiary Boundary Foraminiferal Turnover and Environmental Changes at Agost, Spain (N. McLeod & G. Keller, Eds.): Cretaceous-Tertiary Mass Extinctions: Biotic and Environmental Changes Norton, New York/London, p.139-171.
- Poag, W.C., 1985, Depositional history and stratigraphic reference section for central Baltimore Canyon trough: In Poag, CW (Ed), Geologic Evolution of the United States Atlantic Margin: New York (Van Nostrand Reinhold), p. 217–264.
- Pujalte, V., 1998, Paleocene Strata of the Basque Country, Western Pyrenees, Northern Spain: Facies and Sequence Development in a Deep-Water Starved Basin: Mesozoic and Cenozoic Sequence Stratigraphy of European Basin, SEPM Special Publications, v. 60, p. 311–325.
- Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Sciences, v. 27, p. 75–113.
- Steckler, M., Mountain, G.S., Miller, K.G., and Christie-Blick, N., 1999, Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping: Marine Geology, v. 154, p. 399–420.

- Sugarman, P.J., Miller, K.G., Bukry, D., and Feigenson, M.D., 1995, Uppermost Campanian–Maestrichtian strontium isotopic, biostratigraphic, and sequence stratigraphic framework of the New Jersey Coastal Plain: Geological Society of America Bulletin, v. 107, p. 19–37.
- Vail, P.R., Mitchum, R.M., Jr, and Thompson, S., III, 1977, Seismic Stratigraphy and Global Changes of Sea Level: Part 3. Relative Changes of Sea Level from Coastal Onlap. (C. E. Payton, Ed.): Seismic Stratigraphy — Applications to Hydrocarbon Exploration, Memoir vol. 26, American Association of Petroleum Geologists, v. 26, p. 63–81.
- Van Sickel, W., Kominz, M.A., Miller, K.G., and Browning, J.V., 2004, Late Cretaceous and Cenozoic sea-level estimates: backstripping analysis of borehole data, onshore New Jersey: Basin Research, v. 16, p. 451–465.
- Voigt, E., 1929, Die Lithogenese der Flach- und Tiefwassersedimente des jüngeren Oberkreidemeeres: Halle, Jahrbuch d. Halleschen Verbandes z. Erforschung d. mitteldeutschen Bodensch tze u. Verwertung, v. 8, p. 1–136.
- Wahyudi, H., 2010, Lithofacies and depositional environment spanning the Cretaceous-Paleogene boundary on the New Jersey coastal plain: M.Sc. Thesis, Rutger University, New Brunswick.
- Watts, A.B., and Steckler, M.S., 1979, Subsidence and eustasy at the continental margin of Eastern North America: American Geophysical Union, Maurice Ewing Symposium, v. 3, p. 218–234.

3.8 Figures



Fig. 3.1. Geological map of the New Jersey Coastal Plain showing coreholes that sampled the K/Pg boundary. ODP Leg 174AX and ODP Leg 150X cores are shown as red circles. Shallow cores are shown in blue circles.

B	e	Formations							Sequences and System Tracts								
Syste	Stag	Sugarman et al., 1995		Olsson et al., 2002			Landman et al., 2007	This Study (Campo Pit)		Sugarman et al., 1995		Olsson et al., 2002		Miller et al., 2004		This Study (Campo Pit)	
Paleogene	Danian	Hornerst	own	Hornerstown Hornerstown			Hornerstown (Burrowed Unit)	Hornerstown		rnerstown	scribed	Home.	Home Home	Not described		Not described	
							(Pinna Layer)	Event Bed		Ho	ot de			ink I	MFS?	ink I	MFS
Cretaceous	M a a s t r i c h t i a n	Tinton		Tinton		(\$	Tinton	Tinto	n	Tinton	Ž			Naves	TST?	Naves	TST LST
		Red Bank	Shrewsbury Member	unk	Shrewsbury Member	New Egypt (for downdip section:		Red Bank	Shrewsbury Member	Navesink	HST	Navesink	HST	Navesink I	HST	vesink I	HST
			Sandy Hook Member	Red B	Sandy Hook Member				Sandy Hook Member						MFS	Na	
		Navesink		Navesink		ink	Not described				TST	TST	TST		Not described		
	Campanian	Mount Laurel			Mount Laurel			Not described		Marshalltown	HST	Marshall.	Marshall Marshalltown				

Fig. 3.2. Stratigraphic columnar section of the Upper Cretaceous – early Paleogene formations and sequences of New Jersey (Sugarman et al., 1995; Olsson et al., 2002; Miller et al., 2004; Landman et al., 2007; and this study).



Fig. 3.3. Photomicrographs of echinoid fecal pellets picked from **A.** Bass River at 383.74 m (1259.0 ft), Hornerstown Fm., Danian. **B.** Meirs Farm at 11.83 m (38.8 ft.) Hornerstown Fm., Danian. **C.** Meirs Farm at 12.25 m (40.2 ft.), Hornerstown Fm., Danian. Scale bar represents 1 mm.



Fig. 3.4. Lithostratigraphy of the Campo Pit section showing measured section, lithologic units (Fm.: Formation), cumulative percentage lithology and mineralogy (CC: clay clasts, Q: quartz; G: glauconite; OHM: opaque heavy minerals), sequence stratigraphy (SB: sequence boundary, TS: transgressive surface; MFS: maximum flooding surface; LST: lowstand systems tract; TST: transgressive systems tract), paleodepth interpretation (USF: upper shoreface; MSF: middle shoreface; LSF: lower shoreface), and stages.



Fig. 3.5. a. Exposure of Uppermost Cretaceous and Paleogene formations. **b.** View of the reddish, indurated uppermost Tinton Formation and the clay unit above the K/Pg boundary. **c., d.** *Ophiomorpha* burrows in uppermost Redbank Formation.



Fig. 3.6. Impact spherules found in the event bed above the K/Pg boundary in the green clay unit (sample 9). Scale bar represents 0.5 mm.



Fig. 3.7. Estimated sea-level curve for the NJCP during the late Maastrichtian to early Paleogene. Ages are estimated based on Sugarman et al. (1995) using GTS-2012 time scale (Gradstein et al., 2012).

Chapter 4

Evidence for the Living Ocean following the Cretaceous/Paleogene Mass Extinction

4.1 Abstract

The Cretaceous/Paleogene (K/Pg) mass extinction was associated with a collapse in the carbon isotopic (δ^{13} C) gradient between planktonic and benthic foraminifera, along with a decrease in bulk carbonate $\delta^{13}C$ values. These perturbations have been explained by three hypotheses: global collapse of primary productivity (Strangelove Ocean); greatly reduced export but not primary productivity (Living Ocean); and persistent export productivity (Resilient Ocean). While previous studies have focused on deep-sea locations, we tested the hypotheses in the paleoshelf of New Jersey, where $\delta^{13}C$ values and organic carbon accumulation rates can help distinguish among the hypotheses. We also evaluated interbasinal deep-sea benthic foraminiferal $\delta^{13}C$ gradients between the Pacific (ODP Site 1210) and the Atlantic (ODP Site 1262) Oceans as a proxy for changes in export productivity. On the New Jersey shelf, the K/Pg boundary is associated with a ~2.5‰ δ^{13} C decrease in bulk carbonate, a ~0.8‰ δ^{13} C decrease in organic carbon, a collapse of the vertical δ^{13} C gradient between planktonic and benthic foraminifera, and a significant decrease in organic carbon mass accumulation rates. We interpret a ~1.0% water column $\delta^{13}C$ gradient in planktonic foraminifera and a ~0.75% cross-shelf $\delta^{13}C$

gradient in benthic foraminifera in the early Danian to reflect the presence of active primary productivity, supporting the Living Ocean hypothesis. Deep-sea sites show that the interbasinal deep-sea benthic foraminiferal δ^{13} C gradient was reduced after the mass extinction consistent with a reduction in global export productivity after the K/Pg boundary, though the ocean floor in high productivity regions continued to receive significant export productivity.

4.2 Introduction

The mass extinction at the Cretaceous/Paleogene (K/Pg) boundary resulted in the alteration of both marine and terrestrial ecosystems in addition to the disappearance of large groups such as non-avian dinosaurs, marine and flying reptiles, ammonites, and rudists (Schulte et al., 2010; Raup and Sepkoski, 1982). Calcareous plankton suffered catastrophic extinctions, with the disappearance of more than 90% of Cretaceous species of planktonic foraminifera (Smit, 1982; Olsson and Liu, 1993; Olsson et al., 1999) and calcareous nannofossils (e.g., Bernaola and Monechi, 2007).

The mass extinction was associated with drastic perturbations in the oceanic carbon cycle as shown by a collapse in the carbon isotopic (δ^{13} C) gradient between surface and deep waters recorded by planktonic and benthic foraminifera, and a significant drop in bulk carbonate δ^{13} C values (e.g., Zachos and Arthur, 1986). A δ^{13} C gradient between surface and deep waters develops from the sinking of organic carbon from the surface ocean and its oxidation at depth, called the biological pump (Broecker and Peng, 1982; Volk and Hoffert, 1985). The export of organic carbon from surface waters produces a vertical δ^{13} C gradient (Kroopnick, 1985) because 12 C is preferentially used by photosynthetic organisms, causing the δ^{13} C of organic carbon to be ~20% lower than the δ^{13} C of the dissolved inorganic carbon (DIC) in seawater (Degens, 1969). Therefore, the $\delta^{13}C$ of DIC in modern surface waters is 1 to 3‰ higher than the $\delta^{13}C$ of DIC in the deep oceans (Kroopnick, 1985). Previous studies have shown that the δ^{13} C difference between surface-dwelling planktonic and deep-sea benthic foraminiferal tests in the latest Cretaceous was $\sim 2\%$ (D'Hondt, 2005), which is similar to that measured in the modern oceans (Kroopnick, 1985). However, this difference apparently disappeared at the beginning of the Paleocene (D'Hondt, 2005; D'Hondt and Zachos, 1998; Zachos and Arthur, 1986; Zachos et al., 1989; Coxall et al., 2006; Stott and Kennett, 1989). The collapse of the surface to deep ocean $\delta^{13}C$ difference suggests that there were drastic changes in primary productivity and/or export productivity to the deep ocean after the K/Pg boundary mass extinction (D'Hondt, 2005; D'Hondt and Zachos, 1998; Zachos and Arthur, 1986; Zachos et al., 1989; Coxall et al., 2006; Stott and Kennett, 1989). Nonetheless, this interpretation has been called into question by recent studies (Alegret and Thomas, 2009; Alegret et al., 2012) that attribute the collapse to effects associated with the dramatic biotic turnover.

Three distinct hypotheses are invoked to explain the collapse in surface to deep ocean δ^{13} C difference after the K/Pg boundary. The *Strangelove Ocean* hypothesis posits that primary productivity sharply decreased or ceased immediately after the K/Pg boundary due to the extinction of primary producers (Broecker and Peng, 1982; Hsü et al., 1982). The *Living Ocean* hypothesis postulates that marine primary productivity did not decrease, but export productivity (total organic carbon that sank to the deep ocean via

the biological pump) was dramatically reduced for a few hundred thousand years and moderately reduced for millions of years (D'Hondt, 2005; D'Hondt and Zachos, 1998; Coxall et al., 2006; Sepulveda et al., 2009; D'Hondt et al., 1998). The Living Ocean hypothesis assumes that total biological productivity recovered rapidly, but the export productivity from the photic zone was reduced for millions of years because of a greatly altered ecosystem (D'Hondt, 2005; D'Hondt and Zachos, 1998). A third hypothesis argues that export productivity must have persisted at a sufficient level to support phytoplankton-dependent deep-sea benthic foraminifera that did not suffer significant extinctions (Alegret and Thomas, 2009; Alegret et al., 2012). An increase in the relative abundance of benthic foraminiferal buliminids coeval with a peak in benthic foraminiferal accumulation rates at two sites in the Pacific Ocean was interpreted to show food supply to the sea-floor after the K/Pg extinction remained high, in disagreement with both the Strangelove and Living Ocean hypotheses (Alegret and Thomas, 2009; Alegret et al., 2012). Biogenic barium and other proxies led Hull and Norris (2011) to conclude that there was either no change or an increase in export production in the central Pacific and in some upwelling or shelf Atlantic sites in the early Paleocene. We term this ocean model with relatively high food supply to the sea floor as the Resilient Ocean hypothesis. In the Resilient Ocean hypothesis, the large $\delta^{13}C$ decrease in bulk sediment (2) -3%) in open ocean sites and the collapse of the vertical gradient have been attributed to the turnover and vital effects in nannofossils and planktonic foraminifera (Alegret et al., 2012). Deep-ocean records indicate that the biotic recovery from the mass extinction and the redevelopment of a full δ^{13} C gradient between surface and deep waters took more than 3 Myr (D'Hondt and Zachos, 1998).

Although previous studies focused on deep-sea sections, we propose to test the hypotheses in a paleoshelf setting that provides important records regarding to the δ^{13} C changes across K/Pg boundary. Most organic carbon burial occurs on continental margins (Premuzic et al., 1982), and therefore, any hypothesis concerning large-scale changes in primary productivity and its export must account for changes recorded on the shelves. Each hypothesis has a unique prediction on the shelf in terms of surface waters, deep waters, and water column δ^{13} C gradients. The Strangelove Ocean predicts that there would be limited primary productivity and hence no carbon burial for an extended period. Therefore, in the Strangelove Ocean there would be no $\delta^{13}C$ gradient on the shelf between surface dwelling planktonic and benthic foraminifera independent of the depth of the site relative to the photic zone. The Living Ocean, which assumes primary productivity is present, predicts a δ^{13} C gradient in the water column among planktonic for a living at different depths. In the Living Ocean, $\delta^{13}C$ gradient between surface dwelling planktonic and benthic foraminifera would vary on the shelf depending on the location, i.e. the depth of the site relative to the photic zone. Shallower shelf sites, where the bottom intersects the photic zone or the deep layer is shallow, would have higher surface to bottom gradients and lower benthic foraminiferal δ^{13} C values than the deeper shelf sites. Therefore, a strong cross-shelf gradient among benthic foraminifera would support the Living Ocean hypothesis. On the other hand, the Resilient Ocean hypothesis predicts little change in organic carbon flux across the boundary in the shelf; hence, surface to bottom and water column gradients would be maintained.

Deep-sea benthic foraminiferal $\delta^{13}C$ gradients between the Atlantic and Pacific also help to test the three hypotheses, because they reflect both water mass differences and global export productivity from the biologic pump (Broecker and Peng, 1982). If the biologic pump shuts down (as predicted by the Strangelove and Living Oceans), then the δ^{13} C gradient between the Atlantic and Pacific Oceans due to circulation differences would be severely reduced or disappear (Stott and Kennett, 1989; Zachos et al., 1992).

Here, we present high-resolution $\delta^{13}C$ and oxygen ($\delta^{18}O$) isotope records from multiple species of planktonic and benthic foraminifera, δ^{13} C changes of bulk and organic carbon, and the changes in the mass accumulation rates of total organic carbon (MAR- C_{org}) from the New Jersey paleoshelf (Fig. 4.1) between ~ 69 and 65 Ma (where the K/Pg boundary is placed at 66.04 Ma) (Gradstein et al., 2012). The Ancora (Hole B) and Bass River coreholes drilled by Ocean Drilling Program Leg 174AX provide the most complete K/Pg boundary sections in the New Jersey paleoshelf (Fig. 4.2), representing shallower and deeper shelf environments, respectively. The updip and downdip comparison of the Ancora and Bass River coreholes allows us to distinguish among the hypotheses because they monitored: 1) mixed layer to thermocline $\delta^{13}C$ differences in planktonic foraminifera, 2) cross shelf δ^{13} C differences in benthic foraminifera, and 3) organic carbon burial rates across the K/Pg boundary. Our goal is to evaluate the changes in productivity in a paleoshelf setting, where the three hypotheses provide distinctly different and testable responses. We also aim to evaluate the interbasinal deep-sea $\delta^{13}C$ differences between Atlantic (ODP Site 1262) and Pacific Oceans (ODP Site 1210) to test the hypotheses and provide a comparison between shelf and deep-sea record of the δ^{13} C changes across the K/Pg boundary.

4.3 Methods

For the biostratigraphic analysis and species-specific stable isotope measurements, core samples (20 g) were disaggregated using Calgon solution (5.5 g of sodium metaphosphate per 4 liters of water) and washed with tap water through a 63- μ m sieve. After drying in an oven at 40°C, samples were sieved through 250, 150, 125, and 63 μ m sieves and each size fraction examined separately with a reflected-light microscope. Planktonic foraminifera identifications follow the taxonomy of Olsson et al. (1999) for Danian and Premoli Silva and Verga (2004) for the Maastrichtian.

Stable isotope measurements of bulk carbonate and foraminifera were conducted in the Stable Isotope Laboratory of the Department of Earth and Planetary Sciences at Rutgers University. Samples were loaded into a multi-prep device attached to a Micromass Optima mass spectrometer. The CaCO₃ was reacted in 100% phosphoric acid at 90°C for 800 seconds. Values were reported relative to the Vienna Pee Dee Belemnite (VPDB) using internal laboratory reference material that is routinely calibrated against NBS-18 and NBS-19 (Coplen et al., 1983). The long-term standard deviation on the internal laboratory standard is 0.05‰ and 0.08‰ for δ^{13} C and δ^{18} O measurements, respectively.

Bulk samples collected from Ancora and Bass River were dried at 40°C and homogenized using a ceramic mortar and pestle. Isotopic ratios for organic carbon and TOC were measured at the Boone Pickens School of Geology at the Oklahoma State University using a Thermo Finnigan Delta Plus XL isotope ratio mass spectrometer interfaced to a Costech ECS 4010 elemental analyzer (EA). Organic carbon isotope and concentration measurements were performed on 10-20 mg of sample, which was placed in silver capsules and decarbonated using hydrochloric acid, oven dried at 40°C, then wrapped in tin capsules. Carbon isotope delta values ($\delta^{13}C_{org}$) for Bass River are reported relative to VPDB and calculated using two isotopic standards, USGS40 and USGS24, whose values were calibrated using the L-SVEC/NBS-19 scale (Coplen et al., 2006). Replicates for carbon standards had a standard deviation of ±0.1‰. Replicates of Bass River sediment samples for $\delta^{13}C_{org}$ had a standard deviation of less than ±0.2‰. TOC was calculated from the EA chromatogram using an acetanilide standard curve; replicates of samples had a standard deviation of ±0.03 wt % for Bass River and ±0.07% for Ancora. Additional TOC measurements were conducted for Ancora samples at Rutgers University with Carlo Erba NA1500 series 2 EA with a similar analytical procedure.

Measurements for total nitrogen (TN) for both Ancora and Bass River samples were performed on a Eurovector EA attached under continuous flow to a GV Instruments IsoPrime isotope ratio mass spectrometer in the Stable Isotope Laboratory of the Department of Earth and Planetary Sciences at Rutgers University, using procedures described previously in Quan et al. (2008). Samples for total nitrogen measurements were performed on bulk sediment, and vanadium pentoxide was added as a catalyst. TN was calculated from the EA chromatogram using an acetanilide standard curve; replicate analyses of the nitrogen samples had a variation in TN of < 0.1%.

4.4 Results

4.4.1 New Jersey Paleoshelf

4.4.1.1 Biostratigraphy and Age Calibration

Upper Cretaceous to lower Paleocene strata at Ancora and Bass River consist of glauconitic clay and clayey glauconitic sand deposited in neritic paleoenvironments. Bass River provides a continuous record of deposition across the K/Pg boundary with preservation of impact spherules and a reworked Cretaceous clast layer (Olsson et al., 1997). Ancora lacks the spherule layer at the K/Pg boundary, though reworked clay clasts containing spherules occur immediately above the boundary (Miller et al., 1999).

The age model was constructed using planktonic foraminiferal datum events based on the chronology Huber (2008) for the Cretaceous and Olsson et al. (1999) and Berggren and Pearson (2005) for the Paleocene, which were calibrated to Geological Time Scale (GTS) 2012 (Gradstein et al., 2012) (Fig. 4.S1).

At Ancora, the interval 198.42 meters below land surface (mbls) to 194.22 mbls is defined as the *Abathomphalus mayaroensis* Zone equivalent with the Highest Occurrence (HO) of *Globotruncana linneiana* at 196.90 mbls (Fig. 4.3). The taxon *Abathomphalus mayaroensis* is absent probably due to ecologic exclusion in depths < 200 m. The interval between 194.22 mbls to 188.49 mbls was defined as the *Pseudoguembelina hariaensis/Plummerita hantkeninoides* Zone (Fig. 4.3). The absence of *Abathomphalus mayaroensis* did not allow us to differentiate between the *Pg. hariaensis* and *Pl. hantkeninoides* Zones. For Ancora, we used two datum events in the Maastrichtian, the HO of *Globotruncana linneiana* (68.37 Ma) and the HO of *G. gansseri* (66.49 Ma) (Fig. 4.3). The base of Zone P0 (66.04 Ma) was determined by the extinction level of Cretaceous planktonic for a miniferal taxa and the base of Zone P α was determined by the Lowest Occurrence (LO) of *Parvularugoglobigerina eugubina* (66.00 Ma). Previous studies indicated a hiatus between Zone Pa and Zone P1a (Miller et al., 1999). At 188.09 mbls within Zone Pla, a 2.6 cm-thick layer of spherules (microtektites), grains of glauconite, intermixed clay, and foraminifera was observed (Miller et al., 1999). The presence of glauconite grains and foraminifera in this layer was interpreted as an indication of the redeposition of the original microtektites and a short hiatus between Zones P α and P1a (Miller et al., 1999). The interval from 188.14 mbls to 186.84 mbls is assigned to Zone P1a (Fig. 4.3) and contains Eoglobigerina edita, Globoconusa daubjergensis, Guembelitria cretacea, Parasubbotina pseudobulloides, and Praemurica taurica. The base of Zone P1b was determined by the LO of Subbotina triloculinoides (65.25 Ma); in this zone E. edita, Gc. daubjergensis, Pa. pseudobulloides, Pr. taurica, and S. triloculinoides were identified. The ages of samples between each datum event were interpolated assuming a constant sedimentation rate (Fig. 4.S1). Because of the hiatus reported in (Miller et al., 1999) between Zone P α and Zone P1a, we extrapolated the age of the first sample of Zone P1a using the sedimentation rate of Zone P α assuming constant sedimentation rate and calculated the hiatus as 0.23 Myr.

For Bass River, the interval between 391.06 mbls to 384.12 mbls was defined as the *Pg. hariaensis/Pl. hantkeninoides* Zone (Fig. 4.3). In the *Pg. hariaensis/Pl. hantkeninoides* Zone we used two datum events for age calibration, the LO of *Pg. hariaensis* (67.30 Ma) and the HO of *Gansserina gansseri* (66.49 Ma). Zone P0 is defined as the interval between the extinction of Cretaceous planktonic foraminiferal taxa (66.04 Ma) and the LO of *P. eugubina* (66.00 Ma). Zone P0 at Bass River includes a 6cm-thick spherule (altered microtektite) layer with common shocked quartz grains and carbonate accretionary lapilli resulting from fallout of ballistic ejecta from the Chicxulub Crater in Yucatan, Mexico (Miller et al., 2010; 1998; Olsson et al., 1997). The LO of *P. eugubina* marks the base of Zone Pa (66.00 Ma) and the HO of the same taxon marks the base of Zone P1a (65.72 Ma). Planktonic foraminifera species that occur in Zone Pa and P1a include *Eoglobigerina eobulloides*, *E. edita*, *Pa. pseudobulloides*, *Pr. taurica*, *S. trivialis*, *Woodringina claytonensis*, and *Woodringina hornerstownensis*. The ages of samples between each datum event were interpolated assuming a constant sedimentation rate (Fig. 4.S1). The ages of the samples above the base of Zone P1a (interval 383.97 mbls to 383.87 mbls) were extrapolated using the sedimentation rate of Zone Pa. [Miller et al., 1998].

4.4.1.2 Paleowater Depth Estimates

Paleowater depth estimates were made by projecting the sites to a dip profile (A-A'; Fig. 4.1) from the outcrop adjacent to Medford, NJ through the Bass River corehole using structural contours on top of the Campanian Mount Laurel Formation (Fig. 4.1; (Zapecza, 1984)). Previous paleobathymetric studies in New Jersey have assumed Cretaceous to Eocene paleoslope gradients similar to the modern 1:1000 shelf gradients (Olsson and Nyong, 1984; Browning et al., 1997; Harris et al., 2010). 2-D backstripping of Steckler et al. (1999) allows modeling of paleogradients by accounting for the effects of compaction, loading and thermal subsidence. The oldest surface backstripped by

Steckler et al. (1999) on the New Jersey margin was the O1 sequence boundary (lower Oligocene; (Browning et al., 2013)), which yielded gradients beneath the coastal plain of 1:1000 to 1:500, with a best fit of 1:800 (Steckler et al., 1999; their Fig. 6a). Kominz and Pekar (2001) used 1:500 as the gradient for the Paleogene ramp profile on which succeeding Oligocene-Miocene clinoforms were built. The precise geometry of the Cretaceous paleoslope is not certain, with best estimates of ~1:750 (M. Steckler, written communication, 2014). Therefore, we used 1:750 in this study (Figs. 4.2, 4.S2) and also provided reconstructions for 1:500 and 1:1000 (Figs. 4.S3, 4.S4). The minimum water depths and thus the paleodepth projections are anchored by the Perrineville, Campo Pit outcrop deposited in a lower shoreface paleoenvironment below storm wave base with paleodepths of 25 ± 5 m (Esmeray-Senlet et al., in review). There is uncertainty of ±5 m in the relative placement of the coreholes; given the different end-member gradients. The errors in the absolute paleodepths are up to ± 20 m (Fig. 4.S2 provides paleowater depth estimates of all coreholes and outcrops on Fig. 4.1). Both Bass River (paleodepth of \sim 110±20m) and Ancora (paleodepth of 70±10 m) were located on the middle-outer shelf during the Late Cretaceous – early Paleogene transition (Fig. 4.2).

4.4.1.3 Paleohydrographic Reconstructions

Modern shelf hydrography provides insights into continental shelf processes that are applicable to the K/Pg paleoshelf. The deep chlorophyll maximum (DCM) occurs on the shelf in the lower photic zone (25 - 50 m) due to the balance of irradiance and nutrient supply from the deeper shelf. In the spring to summer, the water column becomes stratified with the shelf thermocline forming a boundary between nutrient-

depleted surface mixed layer above and the nutrient-rich deep layer below (Cullen, 1982; Fairbanks and Wiebe, 1980; Pingree et al., 1976; Hickman et al., 2009). During this seasonal stratification, the DCM is typically located at the base of the surface mixed layer within the shelf thermocline, and is strongly coupled to the nutricline (Cullen, 1982; Hickman et al., 2009; Sharples et al., 2001). On the modern western North Atlantic shelf, a seasonal thermocline develops between the surface mixed layer and deep layer, with primary productivity strongly concentrated in the stratified months (Glenn et al., 2008; Mouw and Yoder, 2005; Ryan et al., 1999). The thermocline and the DCM breaks down in the fall to winter under the influence of storms (Glenn et al., 2008). The deep layer (>60-100 m; ~ paleodepth of Ancora and Bass River) remains unmixed except during the most severe storms and remains essentially isothermal annually (Glenn et al., 2008). In a warmer world such as the Late Cretaceous – Paleogene greenhouse (Olsson et al., 2001), winds that drive the breakdown of the seasonal thermocline in this region would have been potentially weaker and stratification more pronounced (Hay and Floegel, 2012), and the DCM presumably stronger and perhaps a year-long feature. Paleolatitudes of this region were similar to today (~40°N) (Müller et al., 2008) to lower (~28°N) (Kopp et al., 2009). Considering all these factors, we estimate the depth of the DCM as between 25 -50 m for the New Jersey paleoshelf for the latest Cretaceous to Danian during the spring to summer bloom (Fig. 4.2), when most productivity occurs.

4.4.1.4 Stable Isotope Results

At Ancora and Bass River, we measured $\delta^{13}C$ of bulk and organic carbon, and $\delta^{13}C$ and $\delta^{18}O$ of monospecific planktonic and benthic foraminifera from several species

and genera (Figs. 4.3 – 4.5). We measured planktonic foraminiferal taxa *Rugoglobigerina rugosa*, *Globigerinelloides multispina*, *Heterohelix globulosa*, *Hedbergella holmdelensis* and *Guembelitria cretacea* for the Maastrichtian; *Eoglobigerina eobulloides*, *Subbotina trivialis*, *Subbotina triloculinoides*, *Globanomalina planocompressa*, *Praemurica pseudoinconstans*, and *Globoconusa daubjergensis* for the Danian. We analyzed the benthic foraminiferal taxa *Anomalinoides midwayensis*, *Anomalinoides acuta*, and *Gavelinella* spp. across the boundary, all of which survived the K/Pg extinction.

Previous studies showed a positive correlation between test size and δ^{13} C values in small (typically <150 µm) early Danian planktonic foraminifera (D'Hondt and Zachos, 1993; Norris, 1996; Birch et al., 2012). Therefore, we estimated vital effects in our species-specific planktonic foraminiferal δ^{13} C measurements for post-extinction taxa and for the species *G. cretacea* in the latest Maastrichtian (Supporting Information, Table 4.S1). Using size specific δ^{13} C measurements of planktonic foraminifera from ODP Site 1262 (Birch et al., 2012) and DSDP Site 528 (D'Hondt and Zachos, 1993), and taking 150 µm as a cutoff for size related metabolic overprint (Birch et al., 2012), we calculated the amount of vital effect varying between 0 to 0.4‰ (Table 4.S1). Our estimates are in agreement with *Bornemann and Norris* (2007), who showed that metabolic effects could cause 0.3 to 0.5‰ depletion of planktonic foraminiferal calcite equilibrium δ^{13} C values. We interpreted our results using the vital effect corrected δ^{13} C values (Figs. 4.3, 4.4, 4.6). However, the reconstructed δ^{13} C gradient of DIC in the early Danian remains even although applying this correction.

4.4.1.4.1 Paleoecology of Planktonic Foraminifera

We ranked the planktonic foraminifera as surface, intermediate, and deep layer dwellers using δ^{18} O values (Fig. 4.5). The δ^{18} O of calcite is dependent on the temperature of the water from which it is precipitated (Emiliani, 1954) and, therefore provides constraints on depth habitats of extinct planktonic foraminifera (e.g., D'Hondt and Zachos, 1993; Pearson et al., 2001; Wade and Pearson, 2008).

Guembelitria cretacea, one of the survivor species with *Hd. holmdelensis* and *Hd.* monmouthensis (Olsson et al., 1999; Olsson and Liu, 1993), records the lowest δ^{18} O values (Figs. 4.5, 4.6) indicating its preference to live in surface waters, as also shown in previous studies (D'Hondt and Zachos, 1993; Koutsoukos, 1996). The Maastrichtian planktonic taxon *R. rugosa* has also low δ^{18} O values and it is considered as a surfacedweller (Fig. 4.5) in agreement with previous studies (Boersma and Shackleton, 2012). *H. globulosa* has higher δ^{18} O values than *R. rugosa* and is considered an intermediatedweller and *Gg. multispina* and *Hd. holmdelensis* are identified as deep-dwellers based on the highest δ^{18} O values (Fig. 4.5).

In the Danian, the newly evolved planktonic foraminifera species *E. eobulloides*, *S. trivialis*, *S. triloculinoides*, *Pr. pseudoinconstans*, and *Gn. planocompressa* are considered intermediate-dwellers based on their δ^{18} O values (Fig. 4.5). *Subbotina* species were also considered cool temperature, intermediate depth species in previous studies (Norris, 1996; Birch et al., 2012). *Globoconusa daubjergensis*, which evolved from the Cretaceous survivor *G. cretacea* (Olsson et al., 1999), shows higher δ^{18} O values than other Danian planktonic foraminifera indicating a preference for cooler water (Fig. 4.5). Surface dwellers typically have the highest δ^{13} C due to photosynthesis in the photic zone, whereas intermediate and deep dwellers have lower δ^{13} C values due to export and oxidative regeneration of organic carbon (e.g., Pearson et al., 2001; Wade and Pearson, 2008; Fig. 4.6). The relationship between δ^{18} O and δ^{13} C is well expressed in the Maastrichtian, with an exception of the taxa, *Gg. multispina* (Fig. 4.6). We attribute the higher δ^{13} C values of deep-dweller *Gg. multispina* and *H. holmdelensis* than those of intermediate-dweller *H. globulosa* due to changes in the surface mixed layer related to seasonal differences. The relationship of δ^{13} C and δ^{18} O values in the Danian is completely the opposite (Fig. 4.6). In other words, the vertical gradients reversed in the early Danian, such that newly evolved planktonic foraminifera recorded lower δ^{13} C values than benthic foraminifera (Figs. 4.4, 4.6).

In the early Danian, planktonic foraminifera occupied a much narrower range of the water column than they did in the Maastrichtian (Figs. 4.4, 4.6). The newly evolved planktonic foraminifera record δ^{18} O values that are ~1‰ higher than the Maastrichtian taxa (Figs. 4.5, 4.6). This suggests that either surface waters were 5°C cooler (Shackleton, 1974) or the early Danian (P0 – P1b) planktonic foraminifera did not occupy the surface layer (Figs. 4.5, 4.6). The taxa with the lowest δ^{18} O values also have the lowest δ^{13} C values, indicating calcification in the DCM. This implies that the survivor and newly evolved taxa were grazers living within the DCM, and that symbiosis associated with surface dwellers had not evolved, as noted by previous studies (Norris, 1996; Birch et al., 2012).

4.4.1.4.2 Paleotemperature Estimates

We estimated paleotemperature of the water column using δ^{18} O values of benthic foraminifera and planktonic foraminifera living at different depths. Several paleotemperature equations have been developed to show the relation of oxygen isotopes with temperature and continental ice volume based on the original temperature equation proposed by Epstein et al. (1953). We used the below paleotemperature equation of Shackleton (1974), which is based on O'Neil (1969), where " $\delta^{18}O_c$ " stands for $\delta^{18}O$ of the carbonate and " $\delta^{18}O_w$ " for the $\delta^{18}O$ of seawater.

$$T = 16.9 - 4.38 * (\delta^{18}O_c - \delta^{18}O_w) + 0.10 * (\delta^{18}O_c - \delta^{18}O_w)^2$$

Today, the overall average isotopic composition of ocean water, in terms of VPDB is about -0.28‰ (Craig, 1965). Considering the ice-volume component to the δ^{18} O as 0.8 to 1‰ (Shackleton and Kennett, 1975), we used the δ^{18} O_w as -1.2‰ for an ice-free world during the K/Pg boundary transition. More recent calibrations (Kim and O'Neil, 1997; Bemis and Spero, 1998) yielded slightly lower values, but the relative differences between the surface and deep waters are essentially the same. Enrichments due to evaporation in subtropics would cause higher δ^{18} O_w, therefore our estimates indicate minimum paleotemperatures.

Paleotemperature estimates showed that surface water temperatures were ~23°C and the shelf bottom temperatures were ~16–17°C in the late Maastrichtian to early Danian (Fig. 4.2). Intermediate and deep dweller planktonic foraminifera indicate temperatures of ~21–22°C and ~18–19°C, respectively. The surface to bottom paleotemperature gradient (~7°C) was about the half of the present day's gradient (Glenn

et al., 2008) consistent with a lower meridional thermal gradient and slower ocean circulation (Hay and Floegel, 2012).

4.4.1.5 Carbon Isotope Results

Bulk sediment δ^{13} C values decreased by >2‰ across the K/Pg boundary at Bass River (Figs. 4.3, 4.4). The bulk isotope record at Ancora shows a ~1‰ decrease across the boundary (into Zone Pa), with a ~3‰ decrease by Zone P1a (Figs. 4.3, 4.4). Organic carbon δ^{13} C values at Bass River (Fig. 4.7) show a ~0.8‰ decrease at the K/Pg boundary.

During the Maastrichtian, a well-developed δ^{13} C gradient of ~1.5‰ persisted between surface-dwelling planktonic foraminifera and benthic foraminifera (Figs. 4.3, 4.4). The vertical trophic structure of planktonic foraminifera was well established with surface-dwellers (R. rugosa), intermediate-dwellers (H. globulosa), and deep-dwellers (*Gg. multispina* and *Hd. holmdelensis*), showing ~1.0% δ^{13} C gradient within the water column, resembling modern shelf environments. Across the boundary, planktonic for a miniferal δ^{13} C values decreased by ~2‰ at both Ancora and Bass River. Benthic for a for a show a gradual increase in the late Maastrichtian towards to the K/Pg boundary and a sharp decrease at the boundary. At Bass River a ~1‰ decrease of benthic foraminiferal δ^{13} C values is followed by a ~0.5‰ increase across the boundary. Ancora shows a $\sim 0.5\%$ drop across the boundary and the decreasing trend continues in the early Danian (Figs. 4.3, 4.4, 4.8). The vertical gradient between planktonic and benthic foraminifera collapsed across the boundary at both sites and the well-developed trophic structure of planktonic foraminifera exhibited by $\delta^{13}C$ gradients was also lost (Figs. 4.3, 4.4).

4.4.1.6 Total Organic Carbon and Nitrogen

We measured total organic carbon (TOC) across the K/Pg boundary at Ancora and calculated mass accumulation rates of TOC (MAR- C_{org}) (Figs. 4.8, 4.85) in order to evaluate changes in export productivity independent of isotopic tracers using the following formula:

Bulk density of the Ancora samples was measured and sedimentation rates were calculated using planktonic foraminiferal events and GTS-2012 ages (Gradstein et al., 2012). MAR-C_{org} shows a large decrease of ~0.7 g/cm²/kyr across the boundary and there was very low accumulation of organic carbon (~0.2 g/cm²/kyr) in the early Danian (Fig. 4.8).

In order to understand the source of the organic matter we measured total nitrogen (TN) percentages of the samples from Ancora and Bass River and analyzed carbon to nitrogen (C/N) ratios of the samples (Fig. 4.S6). Ancora and Bass River show C/N ratios that average 12 and 15, respectively and both indicate a marine source (Sampei and Matsumoto, 2001; Ohkouchi et al., 2003).

4.4.2 Atlantic – Pacific Interbasinal Deep-Sea Carbon Isotopic Gradients

We evaluated interbasinal deep-sea δ^{13} C gradients across the K/Pg boundary using the benthic foraminiferal δ^{13} C records from Site 1210 (ODP Leg 198, Shatsky Rise, North Pacific; 2574 m present and ~1500 – 2000 m paleodepth) and Site 1262 (ODP Leg 208, Walvis Ridge, South Atlantic; 4759 m present and ~2500 –3000 m paleodepth). Previously, Alegret et al. (2012) presented the benthic foraminiferal δ^{13} C records from these two sites using the astronomically tuned age models of Westerhold et al. (2008) showing that the benthic foraminiferal δ^{13} C gradient between these sites was maintained across the K/Pg boundary (see their Fig. 2B).

We used planktonic foraminifera (Bralower et al., 2002; Petrizzo et al., 2005; Zachos et al., 2004) and calcareous nannofossil biostratigraphy (Bralower, 2005; Zachos et al., 2004; Bernaola and Monechi, 2007), as well as magnetostratigraphy (Westerhold et al., 2008; Bowles, 2006) to revise the age calibrations of these two sites (Figs. 4.9, 4.10; Supporting Information). We also revised the iron-count tie points, which were used for the astronomical time scales of Westerhold et al. (2008) and Hilgen et al. (2010), for a 1.5 Myr time interval after the K/Pg boundary to correlate Site 1210 and Site 1262 in addition to biostratigraphic datum levels (Fig. 4.11). Our correlation with 14 tie-points (Fig. 4.11) is more compatible with the calcareous nannofossil and planktonic foraminiferal datum events than previous correlations (Westerhold et al., 2008; Hilgen et al., 2010) as also shown in the Shaw plot (Fig. 4.12). Some tie points, namely 1, 4, 8, and 12, are also projected onto the age-depth plots of sites in order to show their positions (Figs. 4.9, 4.10). Our new age models for the Sites 1210 (Fig. 4.9) and 1262 (Fig. 4.10) indicate that the deep-sea benthic foraminiferal $\delta^{13}C$ gradient between Atlantic and Pacific Oceans was reduced in the early Danian (Fig. 4.11).
4.5 Discussion

4.5.1 Planktonic Foraminiferal δ^{13} C Changes on the Shelf

Planktonic foraminiferal δ^{13} C values at Ancora show a ~1‰ gradient (-0.2 to -1.2‰) among the newly evolved taxa during Biozone P1a (Fig. 4.8). In the absence of early Danian surface dwellers, this reflects a gradient between intermediate and deepdweller species. We interpret this gradient as reflecting the presence of active mixed layer productivity on the continental shelf ~ 0.4 Myr after the mass extinction, supporting the Living Ocean or Resilient Ocean hypotheses. We cannot comment on the gradients during the earliest Danian (Biozones P0 and P α) because the only planktonic taxon available for measurement is *H. globulosa*.

We measured planktonic foraminiferal species *H. globulosa* across the K/Pg boundary at Ancora, including samples 120 kyr younger than the boundary (Fig. 4.8). *H. globulosa* is not considered a survivor species of the K/Pg mass extinction, and its morphological characteristics indicate that it did not give rise to any Cenozoic foraminifera (Olsson et al., 1999; Olsson and Liu, 1993), though some authors interpret it as a survivor species (Keller, 1988; Pardo and Keller, 2008). The δ^{13} C changes of the species follow the changes in the bulk sediment and benthic foraminiferal δ^{13} C values at Ancora (Fig. 4.8) and it is possible it survived at least 120 kyr and then became extinct as suggested by Pardo and Keller (2008). If so, it recorded δ^{13} C values of intermediate waters in the earliest Danian, which suggests an ~ 0.9‰ decrease across the K/Pg boundary. However, δ^{13} C and δ^{18} O cross plots show that *H. globulosa* has similar values in the late Maastrichtian and early Danian (Fig. 4.S7), perhaps indicating that *H*.

4.5.2 Cross-Shelf Benthic Foraminiferal δ^{13} C Gradients

Early Danian benthic foraminiferal δ^{13} C values at Ancora (~ 0.5‰) and Bass River (~1.25‰) indicate that a substantial δ^{13} C gradient (~ 0.75‰) developed across the continental shelf following the K/Pg event (Figs. 4.3, 4.4). This difference might be explained by two possibilities: 1) terrestrial organic carbon influx caused the shallower site (Ancora) to have lower values than the deeper site (Bass River); or 2) a cross-shelf δ^{13} C gradient existed between sites in the early Danian. Analyses of C/N ratios in both sites indicate a marine source across the boundary (Fig. 4.S6). Therefore, the difference in benthic foraminiferal δ^{13} C values is interpreted to reflect stratification in shelf bottom water δ^{13} C values between these sites such that the deeper site (Bass River) was far below the DCM and thus recorded higher δ^{13} C values than the shallower site (Ancora) (Figs. 4.2, 4.3).

The higher δ^{13} C values at Bass River indicate that either little or no organic carbon was remineralized in the bottom waters or organic carbon was not exported out of the DCM to the bottom at this site. However, benthic foraminiferal δ^{13} C values at Ancora were lower, indicating oxidation of organic carbon in the bottom waters and/or within the sediments. This pattern is interpreted to reflect the higher rate of regeneration of organic matter at the shallower site (Ancora) and the reduced export of organic carbon at the deeper site (Bass River). Neither the Strangelove nor the Resilient Ocean hypotheses can account for the differences in δ^{13} C gradient in bottom waters between Ancora and Bass River. However, the Living Ocean hypothesis predicts this gradient on a continental shelf setting providing evidence for the presence for an active mixed layer independent of the planktonic foraminiferal δ^{13} C values.

4.5.3 Deep-Sea Benthic Foraminiferal δ^{13} C Gradient

We test New Jersey paleoshelf implications by considering deep-sea benthic foraminiferal δ^{13} C differences between different ocean basins across the K/Pg boundary. The Strangelove and Living Ocean hypotheses predict that the interbasinal deep-sea benthic foraminiferal δ^{13} C gradient would collapse in the early Danian, whereas the Resilient Ocean hypothesis predicts that the gradient should remain relatively constant across the K/Pg boundary.

Here, we show that the interbasinal benthic foraminiferal δ^{13} C gradient after the K/Pg boundary strongly depends on the age models used to correlate the sites (Fig. 4.11). The age model of Westerhold et al. (2008), which was also used by Alegret et al. (2012), and age model of Hilgen et al. (2010) show that the benthic foraminiferal δ^{13} C differences was maintained after the K/Pg boundary between the South Atlantic and North Pacific (Fig. 4.11), requiring export production. However, our new age models for the sites based on magnetobiostratigraphy and biostratigraphy, also supported independently by the Fe-count correlation, shows that benthic foraminiferal δ^{13} C gradient was reduced (Fig. 4.11), indicating a decrease in global export productivity.

4.5.4 Resilient Productivity in the Living Ocean

The Resilient Ocean hypothesis explains the survivorship of deep sea benthic foraminifera across the K/Pg boundary by a persistent food supply to the ocean floor (Alegret et al., 2012; Alegret and Thomas, 2009). However, the significant drop in bulk and organic δ^{13} C, the low percentages and low accumulation rates of total organic carbon, and reverse gradient between planktonic and benthic foraminiferal δ^{13} C values after the K/Pg boundary are explained only by the Strangelove or Living Ocean hypotheses. Even though benthic foraminifera survived the K/Pg boundary (Culver, 2003), the diversity of benthic foraminiferal assemblages decreased after the K/Pg boundary in the Pacific and Atlantic Oceans as well as in Tethyan sections indicating high environmental stress (Alegret and Thomas, 2005). While Pacific Ocean sites (DSDP Site 465, ODP 1210) show an increase in benthic foraminiferal accumulation rates and infaunal taxa after the K/Pg boundary (indicative of a high food supply to the ocean floor), Southern Ocean (Site 690) and South Atlantic (Site 1262) sites show little change or a decrease in benthic foraminiferal accumulation rates and infaunal taxa percentages, interpreted as low export productivity after the K/Pg boundary (Alegret et al., 2012).

It has been shown that there is a spatial heterogeneity in the magnitude and direction of the change in export productivity among oceanic sites, even though the mechanism responsible for this heterogeneity is not clear (Hull and Norris, 2011). Geochemical proxies indicate that Atlantic sites except coastal and upwelling sites like Blake Nose and Fish Clay, Denmark show a decrease in export productivity, whereas some Pacific sites like Hess Rise (DSDP Site 465) and Shatsky Rise (ODP Site 1210)

indicate a brief increase in export productivity (Hull and Norris, 2011) in agreement with the benthic foraminiferal proxies (Alegret and Thomas, 2009). While the New Jersey paleoshelf data and our interbasinal deep sea δ^{13} C comparisons are most consistent with the Living Ocean hypothesis, we acknowledge that export productivity in some areas may have changed very little. This heterogeneity in response should not be interpreted as a weakness in our argument, but points to the complexity of reconstructing productivity variations globally.

4.5.5 Carbon Cycle Changes

Comparison of deep-sea records with New Jersey paleoshelf data have important implications related to the long- and short-term changes in the carbon cycle spanning the K/Pg boundary. During the late Maastrichtian, there was a long-term (~ 500 kyr, more than residence time of carbon) increase in benthic foraminiferal δ^{13} C values both in the deep-sea sections (Alegret et al., 2012) and in the New Jersey paleoshelf (Fig. 4.13). This indicates a gradual increase in the organic carbon burial relative to inorganic carbon towards to the K/Pg boundary or a change in the δ^{13} C of weathering input. After the K/Pg boundary, a long-term decrease in δ^{13} C of surface-dwelling planktonic foraminifera, deep-dwelling benthic foraminifera, and bulk sediments occurs in the deep-sea sections (Zachos and Arthur, 1986; Zachos et al., 1989) (Fig. 4.13). This could be attributed to a decrease in global organic carbon burial in the early Danian associated with productivity reduction in surface waters considering that there was not an important change in the weathering input across the K/Pg boundary (Kump, 1991).

The magnitude of decrease varies at different locations. For example, the shelf sections show a sharp benthic foraminiferal δ^{13} C decrease at the K/Pg boundary, and subsequently appear to follow the long-term deep-sea δ^{13} C decrease (Fig. 4.13). The sharp shelf benthic foraminiferal δ^{13} C decrease and lower δ^{13} C values of benthic foraminifera on the shelf than in the deep sea-sections can be attributed to at least three possible causes: 1) input of more terrestrial organic carbon input on the shelf; we reject this hypothesis based on C/N ratios that show a decrease across K/Pg boundary (Fig. 4.S6), suggesting less terrestrial organic matter input in the early Danian; 2) more organic carbon burial on the shelf; we reject this hypothesis based on our measurements that show a decrease in MAR-C_{org} (Fig. 4.8); and 3) shoaling of the oxygen minimum zone where organic carbon is regenerated. With a long-term decrease in the organic carbon burial shoaling of the oxygen minimum zone is predicted, and thus our data support this explanation.

Short-term (less than the residence time of carbon) global δ^{13} C seawater changes across the K/Pg boundary are more difficult to estimate. Deep-sea benthic foraminiferal δ^{13} C values show a short-term increase of 0.8‰ after a ~50 kyr lag (Fig. 4.13) (Alegret et al., 2012), which can be interpreted as reflecting a decrease in organic carbon oxidation at the sediment/water interface and deep water as a result of a reduction in surface water production (Kump, 1991; Kump and Arthur, 1999). Planktonic foraminifera δ^{13} C values show a short-term decrease of ~2‰ in deep sea locations across the K/Pg boundary transition and in the earliest Danian (Zachos and Arthur, 1986; Zachos et al., 1989). Bulk sediments also show a drop of 2–3‰ in deep-sea sections (Alegret et al., 2012) and a 2–2.5‰ decrease in New Jersey shelf cores. Both suggest reduction in organic carbon production in surface waters. However, the decrease in $\delta^{13}C_{org}$ across the K/Pg boundary transition in New Jersey shelf is only 0.8‰. This suggests that the bulk carbonate $\delta^{13}C$ change may overestimate seawater changes in $\delta^{13}C$, supporting Alegret et al.'s (2012) suggestion that the large bulk $\delta^{13}C$ decrease is an artifact of the turnover. However, changes in $\delta^{13}C_{org}$ are highly variable in different K/Pg boundary locations (Meyers, 1992). The difference in the inorganic and organic carbon can be associated with the local differences in the biosynthetic pathways, biochemical constitutions and species compositions that influence the organic matter (Meyers, 1992).

4.6 Conclusions

Integration of paleoshelf and deep-sea isotopic record provide constraints on global carbon budget change across the K/Pg boundary. Updip-downdip comparison of Ancora (paleodepth of 70 ± 10 m) and Bass River (paleodepth of $\sim 110\pm20$ m) provide a unique opportunity to test the existing hypotheses regarding to the perturbations of the carbon cycle after the K/Pg mass extinction. Detailed isotopic measurements on the shelf indicate a decrease in bulk carbonate and organic carbon δ^{13} C values, as well as mass accumulation rates of organic carbon. Similar to deep-sea isotope records, paleoshelf data shows that the vertical δ^{13} C gradient between planktonic and benthic foraminifera collapsed, the trophic structure of the planktonic foraminifera in the water column was lost, and no planktonic foraminifera occupied surface layer in the early Danian. All these perturbations seen both in the paleocontinental shelf and deep-sea locations can be attributed to reduced primary and/or export productivity after the K/Pg boundary event.

However, a ~1.0‰ water column δ^{13} C gradient in planktonic foraminifera and a ~0.75‰ cross-shelf δ^{13} C difference in benthic foraminifera in the early Danian measured in the shelf indicate the presence of active primary productivity, strongly supporting the Living Ocean hypothesis. Atlantic to Pacific interbasinal deep-sea δ^{13} C comparisons based on our new age model show a reduction in interbasinal benthic foraminiferal δ^{13} C differences after the K/Pg boundary, also most consistent with the Living Ocean hypothesis. In spite of this evidence supporting the Living Ocean hypothesis, uniformity in the change of export productivity in the world shelves and deep sea oceans cannot be expected; the ocean floor in highly productive regions like the equatorial Pacific may have continued to receive significant export productivity.

4.7 Acknowledgments

Data and figures supporting the manuscript are in Supporting Information. Samples from Ancora and Bass River coreholes were provided by the Integrated Ocean Drilling Program (IODP); cores are archived at the IODP Rutgers core repository. Supported by NSF grants EAR 070778 (Miller) and OCE-0961914 (Quan). We thank P. Falkowski for discussions and E. Thomas for a review of a draft of this manuscript.

- Alegret, L., and Thomas, E., 2005, Cretaceous/Paleogene boundary bathyal paleoenvironments in the central North Pacific (DSDP Site 465), the Northwestern Atlantic (ODP Site 1049), the Gulf of Mexico and the Tethys: The benthic foraminiferal record: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 224, p. 53–82, doi: 10.1016/j.palaeo.2005.03.031.
- Alegret, L., and Thomas, E., 2009, Food supply to the seafloor in the Pacific Ocean after the Cretaceous/Paleogene boundary event: Marine Micropaleontology, v. 73, p. 105– 116, doi: 10.1016/j.marmicro.2009.07.005.
- Alegret, L., Thomas, E., and Lohmann, K.C., 2012, End-Cretaceous marine mass extinction not caused by productivity collapse: Proceedings of the National Academy of Sciences of the United States of America, v. 109, p. 728–732, doi: 10.1073/pnas.1110601109/-/DCSupplemental/ST01.doc.
- Bemis, B.E., and Spero, H.J., 1998, Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations: Paleoceanography, v. 13, p. 150–160.
- Berggren, W.A., and Pearson, P., 2005, A revised tropical to subtropical Paleogene planktonic foraminiferal zonation: Journal of Foraminiferal Research, v. 35, p. 279–298.
- Bernaola, G., and Monechi, S., 2007, Calcareous nannofossil extinction and survivorship across the Cretaceous–Paleogene boundary at Walvis Ridge (ODP Hole 1262C, South Atlantic Ocean): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 132–156, doi: 10.1016/j.palaeo.2007.02.045.
- Birch, H.S., Coxall, H.K., and Pearson, P.N., 2012, Evolutionary ecology of Early Paleocene planktonic foraminifera: size, depth habitat and symbiosis: Paleobiology, v. 38, p. 374–390, doi: 10.5061/dryad.5p92t38g.
- Boersma, A., and Shackleton, N.J., 2012, Oxygen- and carbon-isotope variations and planktonic-foraminifer depth habitats, Late Cretaceous to Paleocene, Central Pacific, Deep Sea Drilling Project Sites 463 and 465: Initial Reports of the Deep Sea Drilling Project, v. 62, p. 513–526.
- Bornemann, A., and Norris, R.D., 2007, Size-related stable isotope changes in Late Cretaceous planktic foraminifera: Implications for paleoecology and photosymbiosis: Marine Micropaleontology, v. 65, p. 32–42, doi: 10.1016/j.marmicro.2007.05.005.
- Bowles, J., 2006, Data report: Revised magnetostratigraphy and magnetic mineralogy of sediments from Walvis Ridge, Leg 208: Proceedings of the Ocean Drilling Program, Scientific Results, v. 208, p. 1–24.

- Bown, J.A.L.A.P.R., 2005, Upper Cretaceous calcareous nannofossil biostratigraphy, ODP Leg 198 (Shatsky Rise, Northwest Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–60.
- Bralower, T.J., 2005, Data Report: Paleocene-Early Oligocene Calcareous Nannofossil Biostratigraphy, ODP Leg 198 Sites 1209, 1210, and 1211 (Shatsky Rise, Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1– 15.
- Bralower, T.J., Premoli Silva, I., and Malone, M.J., 2002, Site 1210: Proceedings of the Ocean Drilling Program, Initial Reports, v. 198, p. 1–89.
- Broecker, W.S., and Peng, T.H., 1982, Tracers in the sea: Lamont-Doherty Geological Observatory Palisades, NY, Columbia University Press, p. 620.
- Browning, J.V., Miller, K.G., and Olsson, R.K., 1997, Lower to middle Eocene benthic foraminiferal biofacies and lithostratigraphic units and their relationship to sequences, New Jersey Coastal Plain: Proceedings of the Ocean Drilling Program, Scientific Results, v. 156, p. 207–228.
- Browning, J.V., Miller, K.G., Sugarman, P.J., Barron, J., McCarthy, F.M.G., Kulhanek, D.K., Katz, M.E., and Feigenson, M.D., 2013, Chronology of Eocene-Miocene sequences on the New Jersey shallow shelf: Implications for regional, interregional, and global correlations: Geosphere, v. 9, p. 1434–1456, doi: 10.1130/GES00857.S1.
- Coplen, T.B., Brand, W.A., Gehre, M., Gröning, M., Meijer, H.A.J., Toman, B., and Verkouteren, R.M., 2006, New Guidelines for δ¹³C Measurements: Analytical Chemistry, v. 78, p. 2439–2441, doi: 10.1021/ac052027c.
- Coplen, T.B., Kendall, C., and Hopple, J., 1983, Comparison of stable isotope reference samples: Nature, v. 302, p. 236–238, doi: 10.1038/302236a0.
- Coxall, H.K., D'Hondt, S., and Zachos, J.C., 2006, Pelagic evolution and environmental recovery after the Cretaceous-Paleogene mass extinction: Geology, v. 34, p. 297– 300, doi: 10.1130/G21702.1.
- Craig, H., 1965, The measurement of oxygen isotope paleotemperatures (E. Tongiorgi, Ed.): In: Proceedings of the Spoleto Conference on Stable Isotopes in Oceanographic Studies and Paleotemperatures. Consiglio Nazionale delle Ricerche, Pisa,, p. 161– 182.
- Cullen, J.J., 1982, The deep chlorophyll maximum: comparing vertical profiles of chlorophyll a: Canadian Journal of Fisheries and Aquatic Sciences, v. 39, p. 791–803.
- Culver, S., 2003, Benthic foraminifera across the Cretaceous-Tertiary (K-T) boundary: a review: Marine Micropaleontology, v. 47, p. 177–226.

- D'Hondt, S., and Zachos, J.C., 1998, Cretaceous foraminifera and the evolutionary history of planktic photosymbiosis: Paleobiology, v. 24, p. 512–523.
- D'Hondt, S., and Zachos, J.C., 1993, On stable isotopic variation and earliest Paleocene planktonic foraminifera: Paleoceanography, v. 8, p. 527–547.
- D'Hondt, S., Donaghay, P., Zachos, J.C., Luttenberg, D., and Lindinger, M., 1998, Organic carbon fluxes and ecological recovery from the Cretaceous-Tertiary mass extinction: Science, v. 282, p. 276–279, doi: 10.1126/science.282.5387.276.
- Degens, E.T., 1969, Chapter 12: Biogeochemistry of stable carbon isotopes in Organic Geochemistry Methods and Results (G. Eglinton & M. T. J. Murphy, Eds.): Springer Verlag Berlin Heidelberg,, p. 304–328.
- Emiliani, C., 1954, Depth habitats of some species of pelagic foraminifera as indicated by oxygen isotope ratios: American Journal of Science, v. 252, p. 149–158.
- Epstein, S., Buchsbaum, R., Lowenstam, H.A., and Hurey, H.C., 1953, Revised carbonate-water isotopic temperature scale: Geological Society of America Bulletin, v. 64, p. 1315–1326.
- Esmeray-Senlet, S., Özkan-Altıner, S., Altıner, D., Miller, K. G., 2014, Planktonic foraminiferal biostratigraphy, microfacies analysis, sequence stratigraphy, and sealevel changes across the Cretaceous/Paleogene boundary in the Haymana Basin, Central Anatolia, Turkey: Journal of Sedimentary Research, in Review.
- Fairbanks, R.G., and Wiebe, P.H., 1980, Foraminifera and chlorophyll maximum: vertical distribution, seasonal succession, and paleoceanographic significance: Science, v. 209, p. 1524–1526.
- Glenn, S., Jones, C., Twardowski, M., Bowers, L., Kerfoot, J., Kohut, J., Webb, D., and Schofield, O., 2008, Glider observations of sediment resuspension in a Middle Atlantic Bight fall transition storm: Limnology and Oceanography, v. 53, p. 2180– 2196.
- Gradstein, F.M., Ogg, J.G., Schmitz, M., and Ogg, G., 2012, The Geologic Time Scale 2012 2-Volume Set: Elsevier B.V., p. 1176.
- Harris, A.D., Miller, K.G., Browning, J.V., Sugarman, P.J., Olsson, R.K., Cramer, B.S., and Wright, J.D., 2010, Integrated stratigraphic studies of Paleocene–lowermost Eocene sequences, New Jersey Coastal Plain: Evidence for glacioeustatic control: Paleoceanography, v. 25, p. PA3211, doi: 10.1029/2009PA001800.

- Hay, W.W., and Floegel, S., 2012, New thoughts about the Cretaceous climate and oceans: Earth-Science Reviews, v. 115, p. 262–272, doi: 10.1016/j.earscirev.2012.09.008.
- Hickman, A.E., Holligan, P.M., Moore, C.M., Sharples, J., Krivtsov, V., and Palmer, M.R., 2009, Distribution and chromatic adaptation of phytoplankton within a shelf sea thermocline: Limnology and Oceanography, v. 54, p. 525.
- Hilgen, F.J., Kuiper, K.F., and Lourens, L.J., 2010, Evaluation of the astronomical time scale for the Paleocene and earliest Eocene: Earth and Planetary Science Letters, v. 300, p. 139–151, doi: 10.1016/j.epsl.2010.09.044.
- Hsü, K., He, Q., McKenzie, J., Weissert, H., Perch-Nielsen, K., Oberhansli, H., Kelts, K., LaBrecque, J., Tauxe, L., and Krahenbuhl, U., 1982, Mass mortality and its environmental and evolutionary consequences: Science, v. 216, p. 249–256.
- Huber, B.T., MacLeod, K.G., and Tur, N.A., 2008, Chronostratigraphic framework for upper Campanian-Maastrichtian sediments on the Blake Nose (subtropical North Atlantic): Journal of Foraminiferal Research, v. 38, p. 162–182.
- Hull, P.M., and Norris, R.D., 2011, Diverse patterns of ocean export productivity change across the Cretaceous-Paleogene boundary: New insights from biogenic barium: Paleoceanography, v. 26, p. PA3205, doi: 10.1029/2010PA002082.
- Keller, G., 1988, Extinction, survivorship and evolution of planktic foraminifera across the Cretaceous/Tertiary boundary at El Kef, Tunisia: Marine Micropaleontology, v. 13, p. 239–263.
- Kim, S.-T., and O'Neil, J.R., 1997, Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates: Geochimica et Cosmochimica Acta, v. 61, p. 3461–3475.
- Kominz, M.A., and Pekar, S.F., 2001, Oligocene eustasy from two-dimensional sequence stratigraphic backstripping: Geological Society of America Bulletin, v. 113, p. 291– 304.
- Kopp, R.E., Schumann, D., Raub, T.D., Powars, D.S., Godfrey, L.V., Swanson-Hysell, N.L., Maloof, A.C., and Vali, H., 2009, An Appalachian Amazon? Magnetofossil evidence for the development of a tropical river-like system in the mid-Atlantic United States during the Paleocene-Eocene thermal maximum: Paleoceanography, v. 24, p. 1–17, doi: 10.1029/2009PA001783.
- Koutsoukos, E., 1996, Phenotypic experiments into new pelagic niches in early Danian planktonic foraminifera: aftermath of the K/T boundary event: Geological Society of London, Special Publications, v. 102, p. 319–335.
- Kroopnick, P.M., 1985, The distribution of ${}^{13}C$ of ΣCO_2 in the world oceans: Deep Sea Research, v. 32, p. 57–84.

- Kump, L.R., 1991, Interpreting carbon-isotope excursions: Strangelove oceans: Geology, v. 19, p. 299–302.
- Kump, L.R., and Arthur, M., 1999, Interpreting carbon-isotope excursions: carbonates and organic matter: Chemical Geology, v. 161, p. 181–198.
- Meyers, P.A., 1992, Changes in organic carbon stable isotope ratios across the K/T boundary: global or local control?: Chemical Geology: Isotope Geoscience section, v. 101, p. 283–291.
- Miller, K.G., Sherrell, R., Browning, J.V., Field, M., Gallagher, W., Olsson, R.K., Sugarman, P.J., Tuorto, S., and Wahyudi, H., 2010, Relationship between mass extinction and iridium across the Cretaceous-Paleogene boundary in New Jersey: Geology, v. 38, p. 867–870, doi: 10.1130/G31135.1.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Cramer, B.S., Olsson, R.K., de Romero, L., Aubry, M.-P., Pekar, S.F., Georgescu, M.D., Metzger, K.T., Monteverde, D.H., Skinner, E.S., Uptegrove, J., Mullikin, L.G., et al., 1999, Ancora Site (K. G. Miller, P. J. Sugarman, & J. V. Browning, Eds.): Proceedings of the Ocean Drilling Program, Initial Reports, v. 174AX (supplement), p. 1–65.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Olsson, R.K., Pekar, S.F., Reilly, J.T., Cramer, B.S., Aubry, M.-P., Lawrence, R.P., Curran, S.J., Metzger, J.M., Uptegrove, J., Bukry, D., Burckle, L.H., et al., 1998, Bass River Site (K. G. Miller, P. J. Sugarman, & J. V. Browning, Eds.): Proceedings of the Ocean Drilling Program, Initial Reports, v. 174AX, p. 5–43.
- Mouw, C.B., and Yoder, J.A., 2005, Primary production calculations in the Mid-Atlantic Bight, including effects of phytoplankton community size structure: Limnology and Oceanography, v. 50, p. 1232–1243.
- Müller, R.D., Sdrolias, M., Gaina, C., Steinberger, B., and Heine, C., 2008, Long-Term Sea-Level Fluctuations Driven by Ocean Basin Dynamics: Science, v. 319, p. 1357– 1362, doi: 10.1126/science.1151540.
- Norris, R.D., 1996, Symbiosis as an evolutionary innovation in the radiation of Paleocene planktic foraminifera: Paleobiology, v. 22, p. 461–480.
- O'Neil, J.R., 1969, Oxygen Isotope Fractionation in Divalent Metal Carbonates: The Journal of Chemical Physics, v. 51, p. 5547–5558, doi: 10.1063/1.1671982.
- Ohkouchi, N., Kuroda, J., Okada, M., and Tokuyama, H., 2003, Why Cretaceous black shales have high C/N ratios: Implications from SEM-EDX observations for Livello Bonarelli black shales at the Cenomanian-Turonian boundary: Earth Evolution, v. 1, p. 239–241.

- Olsson, R.K., and Liu, C., 1993, Controversies on the placement of Cretaceous-Paleogene boundary and the K/P mass extinction of planktonic foraminifera: Palaios, v. 8, p. 127–139.
- Olsson, R.K., and Nyong, E., 1984, A paleoslope model for Campanian-lower Maestrichtian foraminifera of New Jersey and Delaware: Journal of Foraminiferal Research, v. 14, p. 50–68.
- Olsson, R.K., Hemleben, C., Berggren, W.A., Huber, B.T., Editors and Members of the Paleogene Planktonic Foraminifera Working Group, 1999, Atlas of Paleocene Planktonic Foraminifera: Smithsonian Contributions to Paleobiology, v. 85, p. 252.
- Olsson, R.K., Miller, K.G., Browning, J.V., Habib, D., and Sugarman, P.J., 1997, Ejecta layer at the Cretaceous-Tertiary boundary, Bass River, New Jersey (Ocean Drilling Program Leg 174AX): Geology, v. 25, p. 759–762.
- Olsson, R.K., Wright, J.D., and Miller, K.G., 2001, Paleobiogeography of Pseudotextularia elegans during the latest Maastrichtian global warming event: Journal of Foraminiferal Research, v. 31, p. 275–282.
- Pardo, A., and Keller, G., 2008, Biotic effects of environmental catastrophes at the end of the Cretaceous and early Tertiary: Guembelitria and Heterohelix blooms: Cretaceous Research, v. 29, p. 1058–1073, doi: 10.1016/j.cretres.2008.05.031.
- Pearson, P., Ditchfield, P., Singano, J., Harcourt-Brown, K., Nicholas, C., Olsson, R.K., Shackleton, N.J., and Hall, M., 2001, Warm tropical sea surface temperatures in the Late Cretaceous and Eocene epochs: Nature, v. 413, p. 481–487.
- Petrizzo, M.R., Premoli Silva, I., and Ferrari, P., 2005, Data Report: Paleogene Planktonic Foraminifer Biostratigraphy, ODP Leg 198 Holes 1209A, 1210A, and 1211A (Shatsky Rise, Northwest Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–56.
- Pingree, R.D., Holligan, P.M., Mardell, G.T., and Head, R.N., 1976, The influence of physical stability on spring, summer and autumn phytoplankton blooms in the Celtic Sea: Journal of the Marine Biological Association of the United Kingdom, v. 56, p. 845–873, doi: 10.1017/S0025315400020919.
- Premoli Silva, I., and Verga, D., 2004, Practical manual of Cretaceous Planktonic Foraminifera. International School on Planktonic Foraminifera: 3. Course: Cretaceous. Verga & Rettori eds. Universities of Perugia and Milan, Tipografia Pontefelcino, Perugia (Italy), 283 p.
- Premuzic, E.T., Benkovitz, C.M., Gaffney, J.S., and Walsh, J.J., 1982, The nature and distribution of organic matter in the surface sediments of world oceans and seas: Organic Geochemistry, v. 4, p. 63–77, doi: 10.1016/0146-6380(82)90009-2.

- Quan, T.M., van de Schootbrugge, B., Field, M.P., Rosenthal, Y., and Falkowski, P.G., 2008, Nitrogen isotope and trace metal analyses from the Mingolsheim core (Germany): Evidence for redox variations across the Triassic-Jurassic boundary: Global Biogeochemical Cycles, v. 22, p. n/a–n/a, doi: 10.1029/2007GB002981.
- Raup, D., and Sepkoski, J., Jr, 1982, Mass extinctions in the marine fossil record: Science, v. 215, p. 1501–1503.
- Ryan, J.P., Yoder, J.A., and Cornillon, P.C., 1999, Enhanced chlorophyll at the shelfbreak of the Mid-Atlantic Bight and Georges Bank during the spring transition: Limnology and Oceanography, v. 44, p. 1–11.
- Sampei, Y., and Matsumoto, E., 2001, C/N ratios in a sediment core from Nakaumi Lagoon, southwest Japan-usefulness as an organic source indicator: Geochemical Journal, v. 35, p. 189–205.
- Schulte, P., Alegret, L., Arenillas, I., Arz, J.A., Barton, P., Bown, P., Bralower, T.J., Christeson, G., Claeys, P., and Cockell, C. et al., 2010, The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary: Science, v. 327, p. 1214–1218.
- Sepulveda, J., Wendler, J., Summons, R., and Hinrichs, K., 2009, Rapid resurgence of marine productivity after the Cretaceous-Paleogene mass extinction: Science, v. 326, p. 129–132.
- Shackleton, N.J., 1974, Attainment of isotopic equilibrium between ocean water and the benthonic foraminifera genus *Uvigerina*: isotopic changes in the ocean during the last glacial: Colloques Internationaux du C.N.R.S, v. 219, p. 203–209.
- Shackleton, N.J., and Kennett, J.P., 1975, Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation, oxygen and carbon isotope analyses in DSDP Sites 277, 279, and 281: in Kennett, J. P., Houtz, R. E., and others. (eds.), Initial Reports of the Deep Sea Drilling Project, U.S. Government Printing Office, Washington, D.C, v. 29, p. 743–755.
- Sharples, J., Moore, C.M., Rippeth, T.P., Holligan, P.M., Hydes, D.J., Fisher, N.R., and Simpson, J.H., 2001, Phytoplankton distribution and survival in the thermocline: Limnology and Oceanography, v. 46, p. 486–496.
- Smit, J., 1982, Extinction and evolution of planktonic foraminifera after a major impact at the Cretaceous/Tertiary boundary: Geological Society of America Special Papers, v. 190, p. 329–352.
- Steckler, M., Mountain, G.S., Miller, K.G., and Christie-Blick, N., 1999, Reconstruction of Tertiary progradation and clinoform development on the New Jersey passive margin by 2-D backstripping: Marine Geology, v. 154, p. 399–420.

- Stott, L.D., and Kennett, J.P., 1989, New constraints on early Tertiary palaeoproductivity from carbon isotopes in foraminifera: Nature, v. 342, p. 526–529.
- Volk, T., and Hoffert, M.I., 1985, Ocean carbon pumps: Analysis of relative strengths and efficiencies in ocean-driven atmospheric CO₂ changes (E. T. Sundquist & W. S. Broecker, Eds.): The Carbon Cycle and Atmospheric CO₂: Natural Variations Archean to Present, v. 32, p. 99–110.
- Wade, B.S., and Pearson, P.N., 2008, Planktonic foraminiferal turnover, diversity fluctuations and geochemical signals across the Eocene/Oligocene boundary in Tanzania: Marine Micropaleontology, v. 68, p. 244–255, doi: 10.1016/j.marmicro.2008.04.002.
- Westerhold, T., and Röhl, U., 2006, Data Report: Revised Composite Depth Records for Shatsky Rise Sites 1209, 1210, and 1211: Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–26.
- Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H., 2008, Astronomical calibration of the Paleocene time: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 257, p. 377–403.
- Zachos, J.C., and Arthur, M., 1986, Paleoceanography of the Cretaceous/Tertiary boundary event: inferences from stable isotopic and other data: Paleoceanography, v. 1, p. 5–26.
- Zachos, J.C., Arthur, M., and Dean, W., 1989, Geochemical evidence for suppression of pelagic marine productivity at the Cretaceous/Tertiary boundary: Nature, v. 337, p. 61–64.
- Zachos, J.C., Aubry, M.-P., Berggren, W.A., Ehrendorfer, T., Heider, F., and Lohmann, K., 1992, 54. Chemobiostratigraphy of the Cretaceous/Paleocene boundary at Site 750, Southern Kerguelen Plateau: Proceedings of the Ocean Drilling Program, Scientific Results, v. 120, p. 961–977.
- Zachos, J.C., Kroon, D., and Blum, P., 2004, Site 1262: Proceedings of the Ocean Drilling Program, Initial Reports, v. 208, p. 1–90.
- Zapecza, O.S., 1984, Hydrogeologic framework of the New Jersey coastal plain : U.S. Geological Survey Open-File Report, v. 84, p. 1–61.

4.9 Figures



Fig. 4.1. Geological map of New Jersey Coastal Plain with coreholes sampled the K/Pg boundary and drilled by ODP Leg 174AX, ODP Leg 150X, and K/Pg boundary drilling project. Ancora and Bass River coreholes are indicated in bold. Structural contours are on the top of the Campanian (Zapecza, 1984). A-A' is along the dip direction. Fig. 4.2 shows the cross section across A-A'.



Fig. 4.2. Paleoslope model and paleohydrography of the New Jersey shelf showing locations of Ancora and Bass River. Paleodepths appropriate for the latest Cretaceous to Danian were estimated assuming a paleoslope of 1:750 (Fig. 4.S2 shows the paleowater depths of other New Jersey coreholes indicated on Fig. 4.1. Figs. 4.S3 and 4.S4 show reconstructions for paleoslopes of 1:500 and 1:1000, respectively). Paleotemperature estimates based on δ^{18} O of planktonic and benthic foraminifera for different water columns at Ancora and Bass River are indicated (green color for the Maastrichtian and orange color for Danian). Temperature profile is estimated based on δ^{18} O data points indicated in the figure and fitted with a curve similar to modern GEOSECS station 113. The position of the deep chlorophyll maximum is based on modern shelf hydrography (Glenn et al., 2008; Ryan et al., 1999; Mouw and Yoder, 2005).



Fig. 4.3. δ^{13} C changes versus depth across the K/Pg boundary at the Ancora and Bass River coreholes. Right panels zoom into the K/Pg boundary transition and show details of Danian biozones. Lowest (\perp) and highest (\top) occurrences of planktonic foraminifera used in biozone determination and age calibration are shown. mbls: meter below land surface. Black: bulk sediments, red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera, and blue: benthic foraminifera. Legend indicates species key.



Fig. 4.4. δ^{13} C changes across the K/Pg boundary at the Ancora and Bass River coreholes. Ages are based on planktonic foraminiferal biozones calibrated to GTS-2012 (Gradstein et al., 2012). δ^{13} C values of *G. cretacea* and Danian planktonic foraminifera species except *Gn. planocompressa* are vital effect corrected (Supporting Information, Table 4.S1). Black: bulk sediments, red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera, and blue: benthic foraminifera. Legend indicates species key.



Fig. 4.5. δ^{18} O changes across the K/Pg boundary at the Ancora and Bass River coreholes. Ages are based on planktonic foraminiferal biozones calibrated to GTS-2012 (Gradstein et al., 2012). Red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera, and blue: benthic foraminifera. Legend indicates species key.



Fig. 4.6. δ^{13} C versus δ^{18} O values of planktonic and benthic foraminifera in Maastrichtian and early Danian (Zones P0 – P1b) from Ancora and Bass River coreholes. δ^{13} C values of *G. cretacea* and Danian planktonic foraminifera species except *Gn. planocompressa* are vital effect corrected (Table 4.S1). Red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera, and blue: benthic foraminifera. See legend on Fig. 4.5 for species key.



Fig. 4.7. δ^{13} C values of organic (δ^{13} C_{org}) and bulk (δ^{13} C_{bulk}) at Bass River. Note the planktonic foraminiferal biozones and the unconformity between Biozones P1a and P1c.



Fig. 4.8. δ^{13} C and MAR-C_{org} values at the Ancora corehole between 65.25 – 66.50 Ma. Black circles represent MAR-C_{org} values. Danian planktonic foraminiferal δ^{13} C values are vital effect corrected except the species *Gn. planocompressa* (Supporting Information, Table 4.S1). For the δ^{13} C values: black diamonds: bulk sediments, red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera, and blue: benthic foraminifera. See legend on Fig. 4.5 for species key.



Fig. 4.9. GTS-2012 ages (Gradstein et al., 2012) versus revised meters composite depth (rmcd) (Westerhold and Röhl, 2006) of ODP Leg 198 Site 1210 (North Pacific). Calcareous nannofossils biohorizons (red spheres) are from Bown (2005) and Bralower, (2005) for Maastrichtian and Danian, respectively. Planktonic foraminifera biohorizons (blue triangles) are from Bralower et al. (2002) and Petrizzo et al. (2005) for the Cretaceous and Paleocene, respectively. LO: Lowest Occurrence. HO: Highest Occurrence. For calcareous nannofossils: L.: Lithraphidites, C.: Ceratolithoides, M.: Micula, Cr.: Cruciplacolithus, Ch.: Chiasmolithus, S.: Sphenolithus. Nannofossil event LO Fasciculithus spp. is 2nd radiation of Fasciculithus, which is equivalent to the first appearance datum of F. ulii. For planktonic foraminifera: A.: Abathomphalus, P.: Parvularugoglobigerina, S.: Subbotina, Pr.: Praemurica, M.: Morozovella. All biohorizons are from Hole A except nannofossil taxa *Ch. danicus* s.s., which is from Hole B. Sedimentation rates are indicated on the figure. Representative (only 1, 4, 8, and 12) Fe-count tie points of this study (yellow stars) (Fig. 4.11) are projected onto the agedepth plot. There is no magnetostratigraphy data available for Site 1210 for this time interval.



Fig. 4.10. GTS-2012 ages (Gradstein et al., 2012) versus meters composite depth (mcd) of ODP Leg 208 Site 1262 (South Atlantic). Calcareous nannofossils biohorizons from Hole B (red spheres) are from Zachos et al. (2004). Calcareous nannofossils biohorizons from Hole C (pink spheres) are from Westerhold et al. (2008). All planktonic foraminifera biohorizons (blue triangles) are from Hole B (Zachos et al., 2004). LO: Lowest Occurrence. HO: Highest Occurrence. For calcareous nannofossils: *Cr.: Cruciplacolithus, Ch.: Chiasmolithus, S.: Sphenolithus, F.: Fasciculithus.* Nannofossil event LO *Fasciculithus* spp. is 2^{nd} radiation of *Fasciculithus,* which is equivalent to the first appearance datum of *F. ulii.* For planktonic foraminifera: *P.: Parvularugoglobigerina, S.: Subbotina, Pr.: Praemurica, M.: Morozovella.* Depths of magnetochrons are revised depths of Westerhold et al. (2008) from Bowles (2006). Sedimentation rates are indicated on the figure. Representative (only 1, 4, 8, and 12) Fe-intensity tie points of this study (yellow stars) (Fig. 4.11) are projected onto the age-depth plot.



Fig. 4.11. Correlation of ODP Leg 208 Site 1262 (South Atlantic) and ODP Leg 198 Site 1210 (North Pacific) using Fe-counts and biohorizons by (Westerhold et al., 2008), (Hilgen et al., 2010), and this study and comparison of benthic foraminiferal δ^{13} C values (Alegret et al., 2012) based on different age models. Age calibrations for Maastrichtian both for (Westerhold et al., 2008) and (Hilgen et al., 2010) are from (Alegret et al., 2012). Depths of Site 1262 and Site 1210 are meters composite depths and revised composite depths (Westerhold and Röhl, 2006), respectively. Tie-points (green lines) of (Westerhold et al., 2008), namely Pa1, Pa2, Pa3, and Pa are from their Fig. S1. Tie-points (green lines) of (Hilgen et al., 2010), namely 1, 2, and 3 are from their Fig. 2. Tie points of this study are named from 1 – 14. Lowest occurrences of calcareous nannofossil and planktonic foraminifera are indicated by black arrows (Westerhold et al., 2008; Zachos et al., 2004; Bralower, 2005; Petrizzo et al., 2005). For calcareous nannofossils: *Cr.: Cruciplacolithus, Ch.: Chiasmolithus. Cr. edwardsii = Ch. danicus* s.s., *Cr. tenius* s.l. = *Cr. intermedius.* For planktonic foraminifera: *P.: Parvularugoglobigerina, S.: Subbotina, Pr.: Praemurica.*



Fig. 4.12. Revised composite depth (rmcd) versus meter composite depth (mcd) graph of Site 1210 and Site 1262. Calcareous nannofossil and planktonic foraminiferal biohorizons (Westerhold et al., 2008; Zachos et al., 2004; Bralower, 2005; Petrizzo et al., 2005) and Fe-count tie points of this study, (Westerhold et al., 2008), and (Hilgen et al., 2010) are shown. For calcareous nannofossils: *Cr.: Cruciplacolithus, Ch.: Chiasmolithus*. For planktonic foraminifera: *P.: Parvularugoglobigerina, S.: Subbotina, Pr.: Praemurica*.

Site 1210 - Depth (rmcd)



Fig. 4.13. Benthic foraminiferal δ^{13} C record of ODP Sites 1262 and 1210 (Alegret et al., 2012), Bass River, and Ancora with the age model of this study. δ^{13} C_{org} record of Bass River is also shown. Dashed line in Ancora curve represents the hiatus between Biozones P α and P1a.

4.10 Supporting Information

4.10.1 Vital Effects on Carbon Isotopes

The δ^{13} C measured from species-specific planktonic foraminiferal tests may be out of equilibrium with ambient seawater in certain conditions, termed vital effects (Spero and Williams, 1989; Spero et al., 1991). Size related metabolic effects (Berger et al., 1978; Spero and Williams, 1989; Spero et al., 1991; Birch et al., 2012) and photosymbiosis (D'Hondt and Zachos, 1993; Norris, 1996; Birch et al., 2012; Bornemann and Norris, 2007) are the main reasons for these disequilibrium effects on the planktonic foraminifera. We evaluate possible vital effects for the interpretation of δ^{13} C gradient of DIC in early Paleocene. If there is departure from the equilibrium values in early Danian planktonic foraminifera measured, metabolic effects must have been the main reason for this because photosymbiosis after the K/Pg mass extinction developed only after 63.50 Ma among the newly evolved taxa (Norris, 1996; Birch et al., 2012).

The positive correlation between test size and δ^{13} C through ontogeny has been shown in very small (typically <150 µm) early Danian planktonic foraminifera (D'Hondt and Zachos, 1993; Norris, 1996; Birch et al., 2012). A series of size-specific stable isotopes measurements on early Paleocene planktonic foraminifera were conducted at ODP Site 1262 (Birch et al., 2012), DSDP 528 (D'Hondt and Zachos, 1993), and DSDP 577 (D'Hondt and Zachos, 1993). The results showed that in early Paleocene, δ^{13} C and test size are positively correlated in small tests <150 µm (Birch et al., 2012; D'Hondt and Zachos, 1993) and the δ^{13} C test size trajectory flattens out in tests >150 µm for almost all species (Birch et al., 2012). The same trend has also been observed in larger sizes (>150 μm) later in Paleogene only if the species interpreted as having dinoflagellate photosymbiosis (i.e., in *Praemurica* and *Morozovella*) (Birch et al., 2012).

We used size specific measurements of planktonic foraminifera from ODP Site 1262 (Birch et al., 2012) and DSDP Site 528 (D'Hondt and Zachos, 1993) in order to estimate the metabolic effects in our species-specific planktonic foraminiferal δ^{13} C measurements (Table 4.S1). It has been shown that for a certain planktonic foraminiferal taxa the amount of departure from the ambient seawater equilibrium values does not significantly change between different oceans (D'Hondt and Zachos, 1993). Taking 150 µm as a cutoff size for size related metabolic overprint (Birch et al., 2012), we calculated the amount of vital effect for post-extinction planktonic foraminiferal taxa and for the survivor species *G. cretacea* at Ancora and Bass River coreholes (Table 4.S1).

For the metabolic effect calculations, we used size-specific δ^{13} C measurements from ODP Site 1262 from Zone P1b (Birch et al., 2012) for species *E. eobulloides*, *S. trivialis*, S. *triloculinoides*, and *Pr. pseudoinconstans* and from DSDP Site 528 from Zone P1a (D'Hondt and Zachos, 1993) for species *Gc. daubjergensis*, and *G. cretacea* (Table 4.S1). Using the size-specific measurements for individual taxa from Sites 1262 and 528, we calculated the deviation from the equilibrium value for each species. We took the δ^{13} C measurement of the first size fraction >150 µm as the equilibrium value and calculated the offset for each size fraction assuming that specimens >150 µm had no vital effect in early Danian before the redevelopment of photosymbiosis (Birch et al., 2012). We could not calculate the vital effect of the Danian species *Gn. planocompressa*, since its size-specific δ^{13} C measurement has not been found in the literature. In our isotope measurements we used 63, 125, 150, and 250 μ m mesh sieves. If we did not have enough specimens of a species in one size fraction, we picked individuals from two different size fractions in order to have enough amount of material for the isotope measurements (Table 4.S1). For each sample we calculated the mean size of the measured size interval and determined the vital effect corresponding to that size from published data. If there were not measurements for that mean size in the published data, we interpolated/extrapolated the value in order to calculate vital effect (Table 4.S1). After finding the species-specific vital effect, we added the amount to our measured values to eliminate the depletion caused by the metabolic overprint on the small Danian planktonic foraminifera and latest Maastrichtian species *G. cretacea* (Table 4.S1). The amount of vital effect we estimated varies between 0 to 0.4 ‰ (Table 4.S1). This result is also in agreement with (Bornemann and Norris, 2007),who showed that metabolic effects could cause 0.3 to 0.5‰ depletion on planktonic foraminiferal calcite equilibrium δ^{13} C values.

4.10.2 Age Models for Site 1210 and Site 1262

Benthic foraminiferal δ^{13} C records from Ocean Drilling Program (ODP) Site 1210 (Leg 198, Shatsky Rise, North Pacific; 2574 m present and ~1500 – 2000 m paleodepth) and ODP Site 1262 (Leg 208, Walvis Ridge, South Atlantic; 4759 m present and ~2500 – 3000 m paleodepth) were presented by (Alegret et al., 2012). They used astronomically tuned age models of (Westerhold et al., 2008) for the sites and indicated that the benthic foraminiferal δ^{13} C gradient was maintained across the K/Pg boundary. However, we created new age models for the sites using planktonic foraminiferal (Bralower et al., 2002; Petrizzo et al., 2005; Zachos et al., 2004) and calcareous nannofossil biostratigraphy (Bralower, 2005; Bown, 2005; Bernaola and Monechi, 2007; Westerhold et al., 2008; Zachos et al., 2004), as well as magnetostratigraphy (Westerhold et al., 2008; Bowles, 2006) (Figs. 4.9, 4.10) from the most recently published data. Our age calibrations indicate that benthic foraminiferal δ^{13} C gradient between Site 1210 and Site 1262 decreased across the K/Pg boundary (Fig. 4.11).

For the age depth-plot of Site 1210 (Fig. 4.9), we used calcareous nannofossil (Bralower, 2005; Bown, 2005) and planktonic foraminifera (Bralower et al., 2002; Petrizzo et al., 2005) biohorizons for Maastrichtian and Danian. The ages of biohorizons are from GTS-2012 (Gradstein et al., 2012) and depths are revised composite depths (Westerhold and Röhl, 2006). All biohorizons are from Hole A except nannofossil taxa *Chiasmolithus danicus* s.s, which is from Hole B. There is no magnetostratigraphic data available for Site 1210 for the K/Pg boundary transition. Based on biostratigraphic horizons sedimentation rates are calculated as 7.8 m/myr for Maastrichtian and ~3 m/myr for Danian (Fig. 4.9).

For the age-depth plot of Site 1262 (Fig. 4.10), we used calcareous nannofossil biohorizons from Hole B (Zachos et al., 2004) and Hole C (Westerhold et al., 2008; Bernaola and Monechi, 2007; Hilgen et al., 2010), and planktonic foraminifera biohorizons from Hole B (Zachos et al., 2004). The ages of all biohorizons are from GTS-2012 (Gradstein et al., 2012) and depths are meters composite depth. Depth of magnetochrons for the Maastrichtian and Danian are from (Westerhold et al., 2008), which are revised from (Bowles, 2006). Based on biomagnetostratigraphy sedimentation

rates are calculated as 20.4 m/myr for the Maastrichtian and 4.7 - 9 m/myr for the Danian (Fig. 4.10).

Sites 1210 and 1262 were correlated using magnetic susceptibility, iron (Fe) counts, and nannofossil biohorizons by (Westerhold et al., 2008) (Fig. 4.11). A later study (Hilgen et al., 2010) revised the tie-points of (Westerhold et al., 2008) (Fig. 4.11). In this study, we evaluated the correlation of Fe-counts of these two studies (Westerhold et al., 2008; Hilgen et al., 2010) and revised the correlation for the 1.5 Myr time interval after the K/Pg boundary (Fig. 4.11). Our high-resolution correlation with 14 tie-points is more appropriate with the calcareous nannofossil and planktonic foraminiferal datum events than previous correlations (Westerhold et al., 2008; Hilgen et al., 2010) as shown in the Shaw plot (Fig. 4.12). Some tie points, namely 1, 4, 8, and 12, are also projected onto the age-depth plots of sites in order to show their positions (Figs. 4.9, 4.10).

- Alegret, L., Thomas, E., and Lohmann, K.C., 2012, End-Cretaceous marine mass extinction not caused by productivity collapse: Proceedings of the National Academy of Sciences of the United States of America, v. 109, p. 728–732, doi: 10.1073/pnas.1110601109/-/DCSupplemental/ST01.doc.
- Berger, W.H., Killingley, J.S., and Vincent, E., 1978, Stable isotopes in deep-sea carbonates: Box core ERDC-92, West Equatorial Pacific: Oceanologica Acta, v. 1, p. 203–216.
- Bernaola, G., and Monechi, S., 2007, Calcareous nannofossil extinction and survivorship across the Cretaceous–Paleogene boundary at Walvis Ridge (ODP Hole 1262C, South Atlantic Ocean): Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 132–156, doi: 10.1016/j.palaeo.2007.02.045.
- Birch, H.S., Coxall, H.K., and Pearson, P.N., 2012, Evolutionary ecology of Early Paleocene planktonic foraminifera: size, depth habitat and symbiosis: Paleobiology, v. 38, p. 374–390, doi: 10.5061/dryad.5p92t38g.
- Bornemann, A., and Norris, R.D., 2007, Size-related stable isotope changes in Late Cretaceous planktic foraminifera: Implications for paleoecology and photosymbiosis: Marine Micropaleontology, v. 65, p. 32–42, doi: 10.1016/j.marmicro.2007.05.005.
- Bowles, J., 2006, Data report: Revised magnetostratigraphy and magnetic mineralogy of sediments from Walvis Ridge, Leg 208: Proceedings of the Ocean Drilling Program, Scientific Results, v. 208, p. 1–24.
- Bown, J.A.L.A.P.R., 2005, Upper Cretaceous calcareous nannofossil biostratigraphy, ODP Leg 198 (Shatsky Rise, Northwest Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–60.
- Bralower, T.J., 2005, Data Report: Paleocene-Early Oligocene Calcareous Nannofossil Biostratigraphy, ODP Leg 198 Sites 1209, 1210, and 1211 (Shatsky Rise, Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1– 15.
- Bralower, T.J., Premoli Silva, I., and Malone, M.J., 2002, Site 1210: Proceedings of the Ocean Drilling Program, Initial Reports, v. 198, p. 1–89.
- D'Hondt, S., and Zachos, J.C., 1993, On stable isotopic variation and earliest Paleocene planktonic foraminifera: Paleoceanography, v. 8, p. 527–547.
- Glenn, S., Jones, C., Twardowski, M., Bowers, L., Kerfoot, J., Kohut, J., Webb, D., and Schofield, O., 2008, Glider observations of sediment resuspension in a Middle Atlantic Bight fall transition storm: Limnology and Oceanography, v. 53, p. 2180– 2196.
- Gradstein, F.M., Ogg, J.G., Schmitz, M., and Ogg, G., 2012, The Geologic Time Scale 2012 2-Volume Set: Elsevier B.V., p. 1176.
- Hilgen, F.J., Kuiper, K.F., and Lourens, L.J., 2010, Evaluation of the astronomical time scale for the Paleocene and earliest Eocene: Earth and Planetary Science Letters, v. 300, p. 139–151, doi: 10.1016/j.epsl.2010.09.044.
- Mouw, C.B., and Yoder, J.A., 2005, Primary production calculations in the Mid-Atlantic Bight, including effects of phytoplankton community size structure: Limnology and Oceanography, v. 50, p. 1232–1243.
- Norris, R.D., 1996, Symbiosis as an evolutionary innovation in the radiation of Paleocene planktic foraminifera: Paleobiology, v. 22, p. 461–480.
- Petrizzo, M.R., Premoli Silva, I., and Ferrari, P., 2005, Data Report: Paleogene Planktonic Foraminifer Biostratigraphy, ODP Leg 198 Holes 1209A, 1210A, and 1211A (Shatsky Rise, Northwest Pacific Ocean): Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–56.
- Ryan, J.P., Yoder, J.A., and Cornillon, P.C., 1999, Enhanced chlorophyll at the shelfbreak of the Mid-Atlantic Bight and Georges Bank during the spring transition: Limnology and Oceanography, v. 44, p. 1–11.
- Spero, H.J., and Williams, D.F., 1989, Opening the carbon isotope "vital effect" black box 1. Seasonal temperatures in the euphotic zone: Paleoceanography, v. 4, p. 593– 601.
- Spero, H.J., Lerche, I., and Williams, D.F., 1991, Opening the carbon isotope" vital effect" black box, 2, Quantitative model for interpreting foraminiferal carbon isotope data: Paleoceanography, v. 6, p. 639–655.
- Westerhold, T., and Röhl, U., 2006, Data Report: Revised Composite Depth Records for Shatsky Rise Sites 1209, 1210, and 1211: Proceedings of the Ocean Drilling Program, Scientific Results, v. 198, p. 1–26.
- Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H., 2008, Astronomical calibration of the Paleocene time: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 257, p. 377–403.
- Zachos, J.C., Kroon, D., and Blum, P., 2004, Site 1262: Proceedings of the Ocean Drilling Program, Initial Reports, v. 208, p. 1–90.





Fig. 4.S1. GTS-2012 ages (Gradstein et al., 2012) versus depth plots of Ancora and Bass River. Sedimentation rates are indicated on the figure. mbls: meters below land surface. LO: Lowest Occurrence. HO: Highest Occurrence. *Pg.: Pseudoguembelina, Gl.: Globotruncana, G.: Gansserina, P.: Parvularugoglobigerina, S.: Subbotina.*



Fig. 4.S2. Paleoslope model and paleohydrography of the New Jersey shelf showing locations of coreholes that sampled the K/Pg boundary including Ancora and Bass River. Paleodepths appropriate for the latest Cretaceous to Danian were estimated assuming a paleoslope of 1:750. Paleotemperature estimations based on δ^{18} O of planktonic and benthic foraminifera for different water columns at Ancora and Bass River are indicated (green color for the Maastrichtian and orange color for Danian). Temperature profile is estimated based on δ^{18} O data points indicated in the figure and fitted with a curve similar to modern GEOSECS station 113. The position of the deep chlorophyll maximum is based on modern shelf hydrography (Glenn et al., 2008; Ryan et al., 1999; Mouw and Yoder, 2005).



Fig. 4.S3. Paleoslope model and paleohydrography of the New Jersey shelf showing locations of coreholes that sampled the K/Pg boundary including Ancora and Bass River. Paleodepths appropriate for the latest Cretaceous to Danian were estimated assuming a paleoslope of 1:500. Paleotemperature estimations based on δ^{18} O of planktonic and benthic foraminifera for different water columns at Ancora and Bass River are indicated (green color for the Maastrichtian and orange color for the Danian). Temperature profile is estimated based on δ^{18} O data points indicated in the figure and fitted with a curve similar to modern GEOSECS station 113. The position of the deep chlorophyll maximum is based on modern shelf hydrography (Glenn et al., 2008; Ryan et al., 1999; Mouw and Yoder, 2005).



Fig. 4.S4. Paleoslope model and paleohydrography of the New Jersey shelf showing locations of coreholes that sampled the K/Pg boundary including Ancora and Bass River. Paleodepths appropriate for the latest Cretaceous to Danian were estimated assuming a paleoslope of 1:1000. Paleotemperature estimations based on δ^{18} O of planktonic and benthic foraminifera for different water columns at Ancora and Bass River are indicated (green color for the Maastrichtian and orange color for the Danian). Temperature profile is estimated based on δ^{18} O data points indicated in the figure and fitted with a curve similar to modern GEOSECS station 113. The position of the deep chlorophyll maximum is based on modern shelf hydrography (Glenn et al., 2008; Ryan et al., 1999; Mouw and Yoder, 2005).



Fig. 4.S5. A. Total organic carbon (TOC) values (black circles), bulk sediment δ^{13} C values (blue diamonds), planktonic foraminiferal biozones and core photographs of the Ancora corehole across the K/Pg boundary. Right panel zooms into 1.5 m interval spanning the boundary. mbls : meters below land surface. ftbls: feet below land surface. **B.** Black circles show the MAR-C_{org} changes and blue squares show the sedimentation rates at the Ancora corehole spanning the K/Pg boundary. Lowest (\perp) and highest (T) occurrences of planktonic foraminifera used in biozone determination and age calibration are shown. There is a short hiatus of 0.23 Myr between Zones Pa and P1a. Planktonic foraminiferal datum events at Ancora core. *Pg.: Pseudoguembelina, Pl.: Plummerita, Gl.: Globotruncana, G.: Gansserina, P.: Parvularugoglobigerina, S.: Subbotina.*



Fig. 4.86. Carbon to nitrogen ratios at Ancora and Bass River. Hiatuses at Ancora are between 188.15 mbls and 188.14 mbls (between Zones P α and P1a), and between 186.84 mbls and 186.69 mbls (between Zones P1b and P1c).



Fig. 4.S7. δ^{13} C versus δ^{18} O values of planktonic and benthic foraminifera in the Maastrichtian (Zone *Pseudoguembelina hariaensis/Plummerita hantkeninoides*) and early Danian (Zones P0 – P1b) from Ancora and Bass River coreholes. δ^{13} C values of *G. cretacea* and Danian planktonic foraminifera species except *Gn. planocompressa* are vital effect corrected (Table 4.S1). Red: surface-dwelling planktonic foraminifera, orange: intermediate-dwelling planktonic foraminifera, green: deep-dwelling planktonic foraminifera. Note planktonic foraminifera species *H. globulosa* in both panels. Legend indicates species key.

Table 4.S1. Vital effect calculation for the latest Maastrichtian and Danian planktonic foraminifera for Ancora and Bass River. mbls:meters below land surface. E.: Eobulloides, S.: Subbotina, Pr.: Praemurica, Gc.: Globoconusa, G.: Guembelitria, Pg.:Pseudoguembelina, Pl.: Plummerita.

Ancora B

E. eobulloides	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (Birch et al., 2012) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)
	186.84	65.25	P1b	63-150	106.50	0.40	-0.24	0.16
	187.15	65.35	P1a	63-150	106.50	0.40	-0.15	0.25
	187.88	65.60	P1a	63-150	106.50	0.40	-0.55	-0.15

Bircl	h et al., 201	12, ODP Site	1262, P1b	
Size Fraction (µm)	Mean Size (µm)	δ ¹³ C measured (‰)	Vital Effect calculated (‰)]
250-300	275.00			
212-250	231.00	1.86	0.00	
180-212	196.00	1.90	0.00	
150-180	165.00	1.70	0.00	Equilibrium
125-150	137.50	1.65	0.05	
106-125	114.50	1.39	0.31	
	106.50		0.40	Extrapolated
80-106	93.00			

				Size	Mean	Vital Effect	δ ¹³ C	δ ¹³ C	Birc
S. trivialis	Depth (mbls)	Age (Ma)	Biozone	Fraction (µm)	Size (µm)	(Birch et al., 2012) (‰)	measured (‰)	corrected (‰)	Size Fraction (µm)
	187.88	65.60	P1a	125-250	187.50	0.00	-0.58	-0.58	250-300

Birch	Birch et al., 2012, ODP Site 1262, P1b								
Size Fraction (µm)	$\begin{array}{c c} Mean & \delta^{13}C \\ n & Size & measured \\ (\mu m) & (\%) \end{array}$		Vital Effect calculated (‰)						
250-300	275.00								
212-250	231.00	1.91	0.00						
180-212	196.00	1.78	0.00						
150-180	165.00	1.63	0.00	Equilibrium					
125-150	137.50	1.54	0.09						
106-125	114.50	1.37	0.26]					
80-106	93.00]					

S. triloculinoides	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (Birch et al., 2012) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)
	186.84	65.25	P1b	125-250	187.50	0.00	0.21	0.21

Table 4.S1. Continued.

Birc	h et al., 201	2, ODP Site	1262, P1b	
Size Fraction (µm)	Mean Size (µm)	δ ¹³ C measured (‰)	Vital Effect calculated (‰)	
250-300	275.00			
212-250	231.00			
180-212	196.00	2.03	0.00	
150-180	165.00	1.93	0.00	Equilibrium
125-150	137.50	1.83	0.10	
106-125	114.50	1.72	0.21	
80-106	93.00	1.56	0.37]

Pr. pseudoinconstans	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (Birch et al., 2012) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)
	186.84	65.25	P1b	125-250	187.50	0.00	0.07	0.07

Birc				
Size Fraction (µm)	Mean Size (µm)	δ ¹³ C measured (‰)	Vital Effect calculated (‰)	
250-300	275.00			
212-250	231.00	1.88	0.00	
180-212	196.00	1.93	0.00	
150-180	165.00	1.82	0.00	Equilibrium
125-150	137.50	1.63	0.19	
106-125	114.50			
80-106	93.00			

Gc. daubjergensis	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (D'Hondt & Zachos,1993) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)	
	186.84	65.25	P1b	63-150	106.50	0.23	-0.44	-0.21	
	187.15	65.35	P1a	63-150	106.50	0.23	-0.59	-0.36	
	187.88	65.60	Pla	63-150	106.50	0.23	-1.42	-1.19	

D'Hondt	and Zacho	s, 1993, DSD	P Site 528, P1a		
Size Fraction (µm)	Mean Size (µm)	δ ¹³ C measured (‰)	Vital Effect calculated (‰)		
	150.00	1.61	0.00	Extrapolated	Equilibrium
106-125	115.50	1.39	0.22		
	106.50		0.23	Interpolated	
90-106	98.00	1.37	0.24		
75-90	82.50	1.22	0.39		
63-75	69.00	1.09	0.52		

Table 4.S1. Continued.

Bass River

	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (Birch et al., 2012) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)
E. eobuliolaes	383.92	65.56	Pla	63-150	106.50	0.40	0.14	0.54
	383.94	65.63	Pla	63-150	106.50	0.40	0.44	0.84
	383.95	65.67	Pla	63-150	106.50	0.40	0.03	0.43

	Bircl	h et al., 201	12, ODP Site	1262, P1b	
	Size Fraction (µm)	Size Mean Fraction Size (µm) (µm)		Vital Effect calculated (‰)	
	250-300	275.00			
	212-250	231.00	1.86	0.00	
	180-212	196.00	1.90	0.00	
-	150-180	165.00	1.70	0.00	Equilibrium
	125-150	137.50	1.65	0.05	
	106-125	114.50	1.39	0.31	
		106.50		0.40	Extrapolated
	80-106	93.00			

G. cretacea	Depth (mbls)	Age (Ma)	Biozone	Size Fraction (µm)	Mean Size (µm)	Vital Effect (D'Hondt & Zachos,1993) (‰)	δ ¹³ C measured (‰)	δ ¹³ C corrected (‰)		D'Ho Sizo Fract (µm
	384.15	66.04	Pg. hariaensis / Pl. hantkeninoides	63-125	94.00	0.21	0.33	0.54		
	384.24	66.06	Pg. hariaensis / Pl. hantkeninoides	63-125	94.00	0.21	0.89	1.10		125-1
	384.35	66.07	Pg. hariaensis / Pl. hantkeninoides	63-125	94.00	0.21	-0.47	-0.26][106-1
	384.60	66.11	Pg. hariaensis / Pl. hantkeninoides	63-125	94.00	0.21	0.36	0.57		90-1

D'Hondt	and Zacho				
Size Fraction (µm)	Mean Size (µm)	δ ¹³ C measured (‰)	Vital Effect calculated (‰)		
	150.00	1.74	0.00	Extrapolated	Equilibrium
125-150	137.50	1.68	0.06		
106-125	115.50	1.61	0.13		
90-106	98.00	1.54	0.20		
	94.00		0.21	Interpolated	
75-90	82.50	1.33	0.41		
63-75	69.00	1.37	0.37]	

Chapter 5

Iridium Profiles and Delivery across the Cretaceous/Paleogene Boundary in the New Jersey Coastal Plain

5.1 Abstract

We examined iridium (Ir) anomalies at the Cretaceous/Paleogene (K/Pg) boundary in siliciclastic shallow marine cores of the New Jersey Coastal Plain, USA that were deposited at intermediate distance (~2500 km) from the Chicxulub crater. Although closely spaced and generally biostratigraphically complete, the cores show heterogeneity in terms of preservation of the ejecta layers, maximum concentration of Ir measured (~ 0.1 ppb - 2.4 ppb), and total thickness of the Ir-enriched interval (11 - 119 cm). We analyzed the shape of the Ir profiles with a Lagrangian particle-tracking model of sediment mixing. Good fits between the mixing model and measured Ir profiles, as well as visible burrows in the cores, show that the shape of the Ir profiles was primarily determined by sediment mixing via bioturbation. In contrast, Tighe Park 1 and Bass River cores show remobilization of Ir by geochemical processes. There is a strong inverse relationship between the maximum concentration of Ir measured and the thickness of the sediments over which Ir is spread. We show that the depth-integrated Ir inventory is very similar in the majority of the cores, indicating that the total Ir delivery at time of the K/Pg event was spatially homogenous over this region. Cores having higher Ir maxima, but narrower peaks have similar total Ir delivery as cores with lower Ir maxima and diffused peaks;

this provides additional evidence that the original Ir deposition was scattered vertically across the K/Pg boundary due to sedimentary and geochemical processes.

5.2 Introduction

The discovery of anomalously high abundance of iridium (Ir) and other platinum group elements (PGEs) at the Cretaceous/Paleogene (K/Pg) boundary led the hypothesis that the Earth was hit by an asteroid ~10 km in diameter causing severe environmental disturbance (Alvarez et al., 1980; Smit and Hertogen, 1980). The impact hypothesis was supported by the subsequent discovery of shocked minerals (Bohor et al., 1987), impact spherules, and Ni-rich spinels (Smit and Kyte, 1984). Discovery of the buried Chicxulub crater, ~180 to 200 km in diameter in Yucatan Peninsula, Mexico (Hildebrand et al., 1991) also substantiated that the source of Ir and other PGEs is extraterrestrial.

An extraterrestrial source is not the only means of yielding high Ir concentrations at the K/Pg boundary. Deccan flood basalts in India spanning the K/Pg boundary have been suggested as a source for the Ir anomalies at the boundary (Officer and Drake, 1985). Emplacement of Deccan basalts took ~800 kyr during Chron C29r (Courtillot et al., 1986) and occurred in 3 phases, with the main pulse erupting ~340 kyr prior to the K/Pg boundary (Robinson et al., 2009), and ending at the K/Pg boundary (Chenet et al., 2007; Keller et al., 2008). The suggestion that the origin of anomalies in Ir and other platinum group elements (PGEs) are volcanic rather than extraterrestrial has been challenged by measurements of sedimentary PGE ratios, showing similarity to those of meteorites rather than terrestrial basalts (Evans and Chai, 1997; Evans et al., 1993; Koeberl, 2002). In addition, Sawlowicz (1993) and Shukla et al. (2001) proposed that the contribution of Deccan basalts is too small and local to explain the global inventory of Ir at the K/Pg boundary.

Iridium anomalies, mostly associated with ejecta layers (Smit, 1999; Schulte et al., 2010), have been recorded in more than 85 K/Pg boundary sites globally (Claeys et al., 2002; Schulte et al., 2010). The global occurrence of an Ir anomaly suggests that dust and vapor from the impacting bolide and target rock, rich in high-PGE meteoritic material, were transported to the stratosphere creating a homogeneous cloud encircling the Earth. Then, Ir-rich material rained down from the atmosphere on scales of months (Toon et al., 1982) and slowly settled through the water column (Claeys et al., 2002).

Outside the Gulf of Mexico, i.e., in the intermediate and distal sites from the Chicxulub crater, there is no correlation between the Ir concentration and distance from the impact site (Claeys et al., 2002). The original Ir-rich deposits can be redistributed due to remobilization by sedimentary processes including bioturbation, geochemical remobilization, or mass transport, which can account for the site to site differences in Ir concentrations (Sawlowicz, 1993; Claeys et al., 2002), as well as the shape of the Ir anomaly profiles (Hull et al., 2011). In some K/Pg boundary sites the Ir anomaly is concentrated in a thin (~1 cm) interval, whereas at other locations it spreads over as much as several meters of section (Smit, 1999; Claeys et al., 2002). Even geographically close sites show different maximum concentrations of Ir and/or different thicknesses over which the Ir enrichment is spread.

Shallow cores (<25 m) drilled adjacent to outcrops of the K/Pg boundary at several New Jersey Coastal Plain (NJCP) localities (Buck Pit 1, Tighe Park 1, Search

Farm 1, Meirs Farm 1, Inversand, and Fort Monmouth 3) and deeper cores drilled onshore by Ocean Drilling Program (ODP) 174AX (Ancora, Double Trouble, and Bass River) provide important constraints on the impact-related features across the K/Pg boundary (Fig. 5.1). Previously, a 6-cm-thick spherule layer immediately above the K/Pg boundary was reported at Bass River with reworked clay clasts and an Ir peak of 2.4 ppb (Olsson et al., 1997). Miller et al. (2010) documented Ir anomalies at Buck Pit 1, Search Farm 1, Meirs Farm 1, Tighe Park 1, and Bass River to investigate the stratigraphic relationship between the Ir anomalies to the extinction level. Updip sites yield lower Ir anomaly peak concentration (~0.5 ppb) compared to the downdip Bass River site and each core shows a different shape of Ir profile, despite being deposited in close proximity.

We conducted additional Ir measurements in the Ancora, Double Trouble, Inversand, and Fort Monmouth 3 sites (Fig. 5.1) in order to understand Ir concentrations and their potential mobility in the New Jersey sections. Here we address two main questions combining new data with the previous results. First, how did vertical redistribution of Ir by sedimentary processes like bioturbation or geochemical remobilization affect the shape of Ir profiles in the NJCP cores, deposited in shallow marine settings at intermediate distances (~2500 km) from the Chicxulub crater? Second, could the variations in peak Ir anomaly concentrations in NJCP cores, ranging from low to moderate, be attributed to bioturbation, geochemical remobilization, redeposition, or simply concentrations of background values? We analyze the shape of Ir profiles by modeling Ir anomalies under a range of mixing conditions with a Lagrangian advection– diffusion sediment mixing model (Hull et al., 2011) and compare the mixing model parameters with physical observations in the cores. Finally, we evaluate the similarities and differences in depth-integrated anomalies (total vertical flux) among the NJCP cores, and discuss the relevance of this quantity relative to the more-frequently-used peak Ir concentrations.

5.3 Analytical Techniques

Concentrations of Ir were measured using Sector Field Inductively Coupled Plasma Mass Spectrometry at the Institute of Marine and Coastal Sciences, Rutgers University. Pre-concentration and isolation of Ir from the sediment samples was done using NiS fire-assay technique modified after Ravizza and Pyle (1997). In this method, sediment samples were dried at 105°C overnight, and ~ 1 g subsample is finely ground and homogenized using an acid-cleaned agate mortar and pestle. The resulting powder was then mixed with pure Ni powder and sublimed sulfur (2:1 mass ratio), borax (2:1 ratio to sediment mass), and ¹⁹¹Ir enriched isotope spike prepared in 6.2N HCl. This mixture is then heated to 1000°C in a muffle furnace for 75 minutes to allow fusion. After fusion, glassy sample was broken to release bead of Ni:S containing scavenged Ir. Beads are then dissolved in 6.2N HCl at $190 - 200^{\circ}$ C on a hot plate, then filtered through cellulose 0.45 µm filters (Millipore HATF). Filters were then digested in concentrated HNO₃ in a 15 ml screw-cap Teflon vial (Savillex). Quantification of Ir concentrations was conducted by the method of isotope dilution, using the method of standard additions for verification of accuracy. Low procedural blanks (equivalent to 7 pg/g) combined with high sensitivity mass spectrometry provide a detection limit of $\sim 10 \text{ pg/g} = 0.01 \text{ ppb}$. This method yields excellent procedural reproducibility ($\pm 5\%$, 2σ) for even the lowest Ir concentrations (40 – 100 pg/g) found in background samples, allowing unambiguous determination of Ir anomalies.

5.4 Vertical Sediment Mixing and Mixing Model

Bioturbation is the mixing of the upper part of the sediment column by burrowing of benthic macrofauna, recycling the nutrients in the sediment mixed layer (Jumars et al., 1990). The rate and the extent of bioturbation have been investigated quantitatively using geologically instantaneous markers like impact ejecta and volcanic ashes (Glass, 1969; Ruddiman and Glover, 1972; Guinasso and Schink, 1975). Several techniques for modeling the movement of sediment particles across the sediment-water interface and into the permanent sedimentary record have been developed, as reviewed by Meysman (2003). Traditional sediment mixing models are advection–diffusion models, where the record of a sedimentary event is distributed within a mixed layer characterized by a mixing coefficient (i.e., diffusion coefficient) and advected into layers of sediments that are older than the event (Guinasso and Schink, 1975).

Hull et al. (2011) modified the one-dimensional Lagrangian particle-tracking model with depth-dependent eddy diffusivity of Tanaka and Franks (2008) to model Ir anomaly shapes across the K/Pg boundary. They used the mixing model to explain the observed Ir anomaly shape at North Pacific DSDP Site 577B, Shatsky Rise (their Fig. 5). The model fit the measured Ir profile at Site 577B, showing a peak concentration of 5.6 ppb Ir anomaly that is spread over a ~30-cm-thick bioturbated interval, with an r^2 of 0.95

(Hull et al., 2011). These authors showed that the one-dimensional Lagrangian particletracking model is successful in explaining Ir profile shapes, invoking only physical sediment mixing. In this study, we use the mixing model of Hull et al. (2011) to determine whether the anomaly profiles in this shallow water paleo-environment can similarly be explained by bioturbation, and if so what mixing parameters that best fit measured iridium anomalies in the NJCP cores.

In the one-dimensional Lagrangian particle-tracking model of Hull et al. (2011), the material is moved based on depth-dependent diffusivity with a non-deterministic approach (Fig. 5.2). In the model, vertical diffusivity (K_v) is modeled as a decreasing hyperbolic tangent (tanh) function that generates a vertical diffusivity profile that is continuously differentiable with depth (Ross and Sharples, 2004). Vertical diffusivity decreases with depth, causing a well-mixed layer at the top underlain by increasingly poorly mixed sediments. K_v decreases with depth according to:

$$K_{\nu}(z) = \frac{K_0}{2} \left[1 - \tanh\left(\frac{z - z_0}{z_{scale}}\right) \right]$$

where z_0 is the depth of the inflection point in the tanh profile, delineating the bottom of the well-mixed layer of sediment; z_{scale} is the e-folding scale for the tanh profile, determining the depth over which mixing asymptotes to zero; and K_0 is the maximum vertical diffusivity defining the upper limit of mixing in the tanh profile (not diffusivity at the sediment–water interface) (Fig. 5.2). Modeled iridium is moved vertically by vertical diffusivity as a Markov process, where the depth $z_{t+\Delta t}$ of a particle at time $t + \Delta t$ is a function of the depth (z_t) in the previous time step (t):

$$z_{t+\Delta t} = z_t + \frac{\partial K_v(z_t)}{\partial z} \Delta t + R \left[\frac{2K_v \left(z_t + \frac{1 \partial K_v}{2 \partial z} \Delta t \right) \Delta t}{r} \right]^{1/2} + w_s \Delta t$$

where R is a random process with a zero mean and a variance of r, where r = 1/3. R is selected from a uniform distribution ranging from -1 to 1. Sedimentation rate (w_s) causes the sinking of sediment out of the mixed layer continuously. We took the time step Δt as 0.2 years to account for the low sedimentation rates in the NJCP cores.

We simulated Ir deposition by the introduction of particles into the model at the sediment-water interface, with each particle representing an equal amount of Ir (Hull et al., 2011). 100,000 particles were injected into the model at the time of the event (K/Pg boundary) and mixed according to the different model parameterization. Particle numbers were summed with 2 cm deep increments (di) to obtain Ir concentrations.

Sedimentation rates were calculated using planktonic foraminiferal and/or dinocysts first and last appearances calibrated to the GTS-2012 time scale (Gradstein et al., 2012). Sedimentation rates calculated in the early Danian are 0.2 cm/kyr at Bass River, Ancora, and Inversand; 0.3 cm/kyr at Meirs Farm 1 and Double Trouble. For Search Farm 1 and Tighe Park 1 we do not have enough biostratigraphic information to calculate the sedimentation rates. Therefore, while running the model we tried a combination of different w_s possible (0.1, 0.2, 0.3, 0.4, and 0.5 cm/kyr) with other mixing parameters for each core. For all cores best model fit was obtained with sedimentation rates of 0.2 - 0.3 cm/kyr.

The combination of mixing parameters K_0 (0.75, 1, 1.5, 5, 10, 100, 250, 1000 cm2/kyr), z_0 (2, 5, 10, 20, 50, 100, 150 cm), and z_{scale} (1.5, 4, 9.5 cm) were simulated over a period of 1500 years using MATLAB to get the best fit for the measured Ir profiles. For each model the coefficient of determination (r^2) was calculated to determine how well the model fits the measured Ir profiles. Five models with the highest r^2 are shown in supplementary figures (Figs. 5.S1, 5.S2). The best combination of parameters was selected and model simulations were repeated 50 times to take the average, as individual model runs can vary due to the non-deterministic nature of the approach.

Ir concentrations measured in the NJCP cores are much lower that those of European (e.g., Stevns Klint, Denmark; Premovic et al., 2012) and Tunisian (e.g., El Kef, Tunisia; Ben Abdelkader et al., 1997) shallow-water sections. Although the model of Hull et al. (2011) does not take background levels into account, we consider background level of each core, since cores with low to moderate Ir anomalies yield low signal to noise ratios. For each core, to find the background level we took the average of the values measured above and below the anomalies excluding the data points of anomalously high Ir concentrations. The background levels vary between 0.04 ppb to 0.12 ppb. In the r^2 calculation we excluded data points less than or equal to the background level. We matched the lowest value of the model to the calculated background level.

5.5 Stratigraphy and Ir Anomalies

The NJCP cores were deposited in shallow shelf settings with paleowater depths of 20 ± 5 to 110 ± 20 m during the Late Cretaceous – early Paleogene transition (see

Chapter 4 for discussion). Ancora, Double Trouble, and Bass River are downdip cores with paleowater depths of ~70, 80, 100 m, respectively, allowing planktonic foraminiferal biostratigraphy. All other cores are shallower with no or low preservation of planktonic foraminifera, though dinocyst data allow confident placement of the K/Pg boundary. In addition to biostratigraphy, we counted echinoid fecal pellets (see Chapter 3 for details) that provide a distinct lowermost Danian stratigraphic horizon useful for correlation (Figs. 5.3 - 5.5).

All cores were deposited at a distance ~2500 km from the Chicxulub crater and are considered intermediate sites. Intermediate sites are described as having a distance of ~1000 to ~5000 km from the Chicxulub crater with a 2- to 10-cm-thick spherule layer overlain by a 0.2- to 0.5-cm-thick layer having PGE enrichments, shocked minerals, rip-up clasts, and Ni-rich spinels (Smit, 1999; Claeys et al., 2002; Schulte et al., 2010). The NJCP cores show differential preservation of impact related features such as spherule layer, clay clasts, and Ir enrichments (Figs. 5.3 - 5.5).

Bass River, the most complete and downdip core with a paleodepth of ~ 100 m, shows a biostratigraphically complete upper Maastrichtian to lower Paleogene succession (Fig. 5.3). The New Egypt Formation was assigned to *Abathomphalus mayaroensis* planktonic foraminifera, Palynodinium grallator dinocyst, and Micula prinsii nannofossil Zones, representing the uppermost Maastrichtian (Olsson et al., 1997; 2002). Immediately above the K/Pg boundary there is a 6-cm-thick spherule layer with common shocked quartz grains and carbonate accretionary lapilli (Olsson et al., 1997; 2002). Sediments above the spherule layer are assigned to planktonic foraminiferal Zone Pa including first appearance datum of planktonic foraminiferal taxa

Parvularugoglobigerina eugubina and the dinoflagellate taxa *Senoniasphaera inornata* (Olsson et al., 1997). A ~3-cm-thick layer of white clay rip-up clasts containing uppermost Cretaceous foraminifera and dinocysts overlies the spherule layer. There is a large Ir anomaly of 2.4 ppb at the base of the spherule bed and two modest Ir anomalies of 0.6 and 0.7 ppb straddling the top of the spherule bed associated with the clast layer.

Ancora (Hole B) is biostratigraphically complete across the K/Pg boundary (Fig. 5.4) with the presence of planktonic foraminiferal Zones P0 and P α in the earliest Danian (Miller et al., 1999). It contains a 2.6-cm-thick spherule layer with grains of glauconite, intermixed clay, and reworked foraminifera in planktonic foraminiferal Zone P1a (Miller et al., 1999). The presence of glauconite grains and reworked foraminifera in this layer was interpreted as an indication of the redeposition of the original microtektites and a short hiatus between Zones P α and P1a (Miller et al., 1999). There is bioturbation in the 50-cm-thick sediments overlying the boundary. At Ancora there is a low Ir anomaly with a maximum concentration of 0.13 ppb, spread over ~107 cm across the K/Pg boundary.

At Double Trouble, bioturbated clayey glauconite sand of the Upper Cretaceous Navesink Formation is overlain by the very dark green, clayey glauconite sand of the Hornerstown Formation (Fig. 5.4). A ~3.5 cm-thick calcareous shell layer is 50 cm above the K/Pg boundary, which is assigned to P0/P1a planktonic foraminiferal zone. At Double Trouble, a low Ir anomaly of 0.16 ppb was measured straddling the boundary and covering over ~119 cm.

At Search Farm 1 and Meirs Farm 1, basal Danian glauconite sands of the Hornerstown Formation overlie clayey glauconite sand of the Upper Maastrichtian New Egypt Formation (Fig. 5.5). The basal Danian is heavily burrowed at Meirs Farm 1, whereas the Upper Maastrichtian shows bioturbation at Search Farm 1. Although no distinct spherule layer exists, there is a 5-cm-thick clay clast layer at Meirs Farm 1. The clay clast at the K/Pg boundary is associated with a modest Ir anomaly of ~0.5 ppb (Miller et al., 2010), which coincides with an increase in echinoid fecal pellets.

The physical stratigraphy at Buck Pit 1 is slightly different from that in the other cores (Fig. 5.4). At Buck Pit 1, indurated quartz sand of the uppermost Cretaceous Tinton Formation is overlain by a ~40-cm-thick clay unit. The clay unit consists of a basal red clay and an overlying green clay. The Ir anomaly is spread over the clay unit with a peak of 0.4 ppb in the green clay associated with the lowest occurrence of *Senoniasphaera inornata*, a dinocyst marker for the base of the Danian (Miller et al., 2010). Although we did not find any impact spherules at Buck Pit 1 core, the adjacent K/Pg outcrop, Campo Pit, has numerous spherules embedded in clay clasts in the red and green clay (see Chapter 3 for details).

At Tighe Park 1, there is 20-cm-thick *Pinna* bed containing a diverse latest Cretaceous fauna including the ammonite *Discoscaphites iris* and the bivalve *Pinna laqueata* in the adjacent outcrop (Landman et al., 2007). Above the *Pinna* bed, the K/Pg boundary is associated with the lowest occurrence of *Damassadinium californicum* dinocyst species and a peak of echinoid fecal pellets (Fig. 5.3). The Ir anomaly of ~0.5 ppb occurs at the base of the *Pinna* bed, 20 cm below the K/Pg boundary (Miller et al., 2010) confirming the relationship observed between Ir and the *Pinna* bed in the adjacent outcrop in Agony Creek (Landman et al., 2007).

The Inversand Pit at Sewell is known for its well-preserved vertebrate fauna just above the K/Pg boundary. The bone bed in in the lowermost Hornerstown Formation, the Main Fossiliferous Layer (MFL), contains the remains of Cretaceous microfossils and macrofossils, including mosasaurs, fish, ammonites, and birds together with Paleogene microfossils (Gallagher, 1993; 2002; Landman et al., 2004; 2007). An Ir concentration of 0.3 ppb occurs in the middle of the 40-cm-thick MFL, which is ~50 cm above the Navesink/Hornerstown Formation contact (= K/Pg boundary in other cores) (Fig. 5.5). The absence of an Ir anomaly at the Hornerstown/Navesink formation contact indicates a short hiatus at the K/Pg boundary and reworking of Ir-rich sediments in the MFL, which consists of reworked Cretaceous and *in situ* Paleogene fauna.

The stratigraphy of Fort Monmouth 3 is similar to adjacent cores, having an increase in echinoid fecal pellets across the lithological contact of New Egypt/Hornerstown Formation contact (Fig. 5.5). However Ir concentrations across the boundary show background levels (0.04 - 0.08 ppb).

5.6 Iridium Profile Shapes

We applied the Lagrangian advection-diffusion mixing model to the NJCP cores, where Ir is scattered across the K/Pg boundary and bioturbation is visible (Ancora, Double Trouble, Buck Pit 1, Search Farm 1, Meirs Farm 1). Mixing model fits to the measured anomaly shapes are very good, with an r^2 of 0.81 - 0.96.

There is a strong correlation between mixing parameters obtained and physical observations. In the cores, in which Ir is dispersed over $\sim 107 - 119$ cm (Double Trouble, Ancora) z_0 (the bottom of the well-mixed layer) is 100 - 150 cm; cores with Ir dispersion of ~ 30 cm (Buck Pit 1, Search Farm 1) show z_0 values of 20 - 50 cm; and cores with Ir

dispersion of ~10 cm (Meirs Farm 1) show z_0 values of 20 cm (Figs. 5.6, 5.7). This indicates that Ir dispersion across the K/Pg boundary can be explained by physical mixing of sediments via bioturbation.

In the Tighe Park 1 and Bass River sections, the Ir peaks do not occur at the same stratigraphic level as the biostratigraphic K/Pg boundary (Fig. 5.3). Miller et al. (2010) discussed the relative position of the Ir anomaly and mass extinction level in the NJCP cores and concluded that Ir is *in situ*, i.e. at the same level of the biostratigraphic K/Pg boundary, in clay-rich sections, but remobilized geochemically in sandier sections like Tighe Park 1 and Bass River. At Bass River, the highest Ir anomaly occurs 6 cm below the appearance of Danian foraminifera and dinocysts, at the base of the spherule layer as a narrow peak. Because the deposition of Ir takes place slowly as a result of atmospheric fallout it should occur above the ballistic ejecta/spherule layer (Smit, 1999; Kring, 2007). Thus, the Ir enrichment at the base of the spherule layer at Bass River is interpreted as resulting from geochemical remobilization of Ir down section by 6 cm (Miller et al., 2010). Similarly, the Ir anomaly at Tighe Park 1 is a sharp peak at a redox boundary that occurs 20 cm below the appearance of first Danian dinocysts, at the base of the porous Pinna layer (Miller et al., 2010). Both the spherule layer at Bass River and the Pinna layer at Tighe Park 1 provided porous media for percolation of Ir, which accumulated in a 2- to 3-cm-thick zones as distinct and narrow peaks below these aquicludes (Miller et al., 2010) possibly due to differences in redox potential (Colodner et al., 1992).

At Bass River, a secondary Ir enrichment of ~0.7 ppb straddles the top of the spherule layer, where the original deposition of Ir should have occurred (Fig. 5.3). Also at Tighe Park 1, ~0.2 ppb Ir is dispersed in the heavily bioturbated clayey glauconite sand

of Hornerstown Formation just above the K/Pg boundary (Fig. 5.3). We applied the mixing model to these intervals, on the assumption that they represent the redistribution of particulate Ir by sediment mixing (Fig. 5.7). Model mixing parameters obtained for both Bass River and Tighe Park 1 are in line with the physical observations. The best model fit for Tighe Park 1 shows a z_0 of 100 cm, where the dispersion of Ir is ~30 cm; whereas as at Bass River with an Ir dispersion of ~11 cm z_0 is 10 cm.

5.7 Integrated Iridium Anomalies and Total Iridium Delivery

The NJCP sites show different Ir profiles across the K/Pg boundary in terms of maximum concentration of Ir measured, total thickness of the Ir-enriched interval, and background values of Ir (Table 5.1). However, there is a strong inverse relationship between the maximum concentration of Ir measured and the thickness of the Ir diffusion zone. The maximum Ir peak measured at Bass River is 2.4 ppb and the anomaly is concentrated within a 3-cm-thick layer (Fig. 5.3). Above this major peak a horizon exists, in which Ir enrichment of ~0.7 ppb straddles over 8 cm. Including these two peaks, the total thickness of the Ir enrichment interval at Bass River is 11 cm. Meirs Farm 1, Search Farm 1, Buck Pit 1, and Tighe Park 1 show moderate Ir peaks of ~0.5 ppb, which are spread over 10 – 30 cm (Figs. 5.4, 5.5). At Ancora and Double Trouble, where the thickness of the Ir-enriched interval reaches up to ~119 cm, there are much lower maximum concentrations; 0.13 and 0.16 ppb, respectively (Fig. 5.4). This suggests that Ir was diluted after the original time of deposition due to sedimentary processes like bioturbation and/or geochemical remobilization, causing lower peak Ir anomalies.

We evaluate the integrated Ir signal over the thickness of the Ir-enriched interval using the data points above the background levels (Table 5.1). Using the trapezoidal rule, a technique for approximating the definite integral, we calculated the area of the region that is bounded by the Ir anomalies in each core. The area under the curve, i.e., the integrated Ir signal indicates the total Ir delivery flux, independent of the distance over which it was diffused vertically in the core.

Results show that the integrated Ir anomaly at Ancora, Double Trouble, Buck Pit 1, Tighe Park 1, and Bass River are the same (0.06 ppb*m), indicating that the total Ir delivery to these cores were comparable (Table 5.1). Cores having higher Ir maxima, but narrower peak vs. lower Ir maxima and diffused peaks show the same total Ir delivery indicating that the original Ir deposited was scattered vertically across the K/Pg boundary due to sedimentary and geochemical processes.

Search Farm 1, Meirs Farm 1, and Inversand show the same integrated Ir values, though lower than the other cores (0.02 ppb*m). At Inversand, the Ir-rich horizon is situated in the MFL, 50 cm above the K/Pg boundary, and is reworked. Although the Ir peak is associated with the K/Pg boundary at Search Farm 1 and Meirs Farm 1, a short hiatus at the boundary is also possible causing lower integrated Ir anomalies in these two cores. White clasts containing Maastrichtian planktonic foraminifera and outer neritic benthic foraminiferal assemblage were interpreted as rip-ups due to a tsunami that eroded the outer continental shelf and upper slope after a slope failure triggered by earthquakes generated by the Chicxulub impact (Norris et al., 2000; Olsson et al., 2002). This tsunami was distinguished from the impact-generated mega-tsunami that reworked sediments in the Gulf of Mexico, since Florida and Bahama platforms prevented tsunamis from

spreading from the Gulf of Mexico into the western North Atlantic (Norris et al., 2000). This can also explain the heterogeneity of the preservation of the spherules on the NJCP (Miller et al., 2010).

5.8 Conclusions

The shape of the Ir profiles across the K/Pg boundary and the maximum Ir concentrations measured are highly variable in shallow shelf cores of NJCP, although they were deposited geographically very close to each other. The shape of the Ir profiles and the concentrations measured seem to appear primarily determined by sedimentary and geochemical processes, especially bioturbation. Evidence supporting sediment mixing are burrows visible in the cores and good fits between mixing model and measured Ir profiles.

Using the Lagrangian advection-diffusion model (Hull et al., 2011), we were able to show the extent of mixing effects on the Ir records quantitatively across the K/Pg boundary at 7 shallow-water cores in the NJCP. Modeling and fitting the model to the measured iridium profiles showed that Ir anomaly shapes could generally be explained by sediment mixing. Hull et al. (2011) showed that the model is successful to explain Ir profiles in a distal deep-sea site (Site 577B) with a less complicated mixing history. Our modeling effort in shallow water cores in an intermediate distance from the impact site showed that the model is also applicable in the sites with much more complicated mixing histories. However, the mixing model should be applied carefully taken the background values into account, where there is low signal-to-noise ratio with low to moderate Ir concentrations. Furthermore, for complex profiles, the model occasionally fails to fit secondary peaks.

Our evaluation showed that the integrated Ir signal in the majority of cores is the same in the NJCP, strongly suggesting that the total Ir delivery to this region was the same at time of K/Pg event. Although having received comparable amount of Ir, the sites show different maximum concentration of Ir, which is inversely proportional to the total thickness of the Ir-enriched interval. This provides additional evidence of diffusion of Ir after the original time of deposition.

It has been often thought that low Ir concentrations at the K/Pg boundary shows the incompleteness of the boundary or might only be reflecting the accumulation of cosmic dust due to low sedimentation rates (Sawlowicz, 1993). Nevertheless, the findings of this study show that modest Ir concentrations are a product of a single global cause, in this case the Chicxulub impact.

- Alvarez, L., Alvarez, W., Asaro, F., and Michel, H., 1980, Extraterrestrial cause for the Cretaceous-Tertiary extinction: Science, v. 208, p. 1095–1108.
- Ben Abdelkader, O., Ben Salem, H., Donze, P., Maamouri, A., Méon, H., Robin, É., Rocchia, R., and Froget, L., 1997, The K/T stratotype section of El Kef (Tunisia): events and biotic turnovers: Geobios, v. 21, p. 235–245.
- Bohor, B., Modreski, P., and Foord, E., 1987, Shocked quartz in the Cretaceous-Tertiary boundary clays: evidence for a global distribution: Science, v. 236, p. 705–709.
- Chenet, A., Quidelleur, X., Fluteau, F., Courtillot, V., and Bajpai, S., 2007, 40K-40Ar dating of the Main Deccan large igneous province: Further evidence of KTB age and short duration: Earth and Planetary Science Letters, v. 263, p. 1–15.
- Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of Chicxulub ejecta at the Cretaceous-Tertiary boundary (C. Koeberl & K. G. MacLeod, Eds.): Catastrophic Eevents and Mass extinctions: Impacts and Beyond: Geological Society of America Special Paper, v. 356, p. 55–68.
- Colodner, D., Boyle, E., Edmond, J., and Thomson, J., 1992, Post-depositional mobility of platinum, iridium and rhenium in marine sediments: Nature, v. 358, p. 402–404.
- Courtillot, V., Besse, J., and Vandamme, D., 1986, Deccan flood basalts at the Cretaceous/Tertiary boundary?: Earth and Planetary Science Letters, v. 80, p. 361–374.
- Evans, N., and Chai, C., 1997, The distribution and geochemistry of platinum-group elements as event markers in the Phanerozoic: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 132, p. 373–390.
- Evans, N., Gregoire, D., Grieve, R., Goodfellow, W., and Veizer, J., 1993, Use of platinum-group elements for impactor identification: Terrestrial impact craters and Cretaceous-Tertiary boundary: Geochimica et Cosmochimica Acta, v. 57, p. 3737– 3748.
- Gallagher, W.B., 1993, The Cretaceous/Tertiary mass extinction event in the northern Atlantic Coastal Plain: The Mosasaur, v. 5, p. 75–155.
- Gallagher, W.B., 2002, Faunal changes across the Cretaceous-Tertiary (K-T) boundary in the Atlantic Coastal Plain of New Jersey: Restructuring the marine community after the K-T mass-extinction event: In: Koeberl, C., and MacLeod, K.G., eds., Catastrophic Events and Mass Extinctions: Impacts and Beyond: Geological Society of America, Special Paper, v. 356, p. 291–301.

- Glass, B.P., 1969, Reworking of deep-sea sediments as indicated by vertical dispersion of Australasian and Ivory Coast microtektite horizons: Earth and Planetary Science Letters, v. 6, p. 409–415.
- Gradstein, F.M., Ogg, J.G., Schmitz, M., and Ogg, G., 2012, The Geologic Time Scale 2012 2-Volume Set: Elsevier BV, v. 1-2, p. 1176.
- Guinasso, N.L., and Schink, D.R., 1975, Quantitative estimates of biological mixing rates in abyssal sediments: Journal of Geophysical Research, v. 80, p. 3032–3043.
- Hildebrand, A., Penfield, G., Kring, D., Pilkington, M., and Camargo, Z., 1991, Chicxulub crater: a possible Cretaceous/Tertiary boundary impact crater on the Yucatan Peninsula, Mexico: Geology, v. 19, p. 867–871.
- Hull, P.M., Franks, P.J.S., and Norris, R.D., 2011, Earth and Planetary Science Letters: Earth and Planetary Science Letters, v. 301, p. 98–106, doi: 10.1016/j.epsl.2010.10.031.
- Jumars, P.A., Mayer, L.M., Deming, J.W., Baross, J.A., and Wheatcroft, R.A., 1990, Deep-sea deposit-feeding strategies suggested by environmental and feeding constraints: Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences, v. 331, p. 85–101.
- Keller, G., Adatte, T., Gardin, S., Bartolini, A., and Bajpai, S., 2008, Main Deccan volcanism phase ends near the K-T boundary: Evidence from the Krishna-Godavari Basin, SE India: Earth and Planetary Science Letters, v. 268, p. 293–311, doi: 10.1016/j.epsl.2008.01.015.
- Koeberl, C., 2002, Mineralogical and geochemical aspects of impact craters: Mineral Magazine, v. 66, p. 745–768, doi: 10.1180/0026461026650059.
- Kring, D., 2007, The Chicxulub impact event and its environmental consequences at the Cretaceous-Tertiary boundary: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 255, p. 4–21.
- Landman, N., Johnson, R., and Edwards, L., 2004, Cephalopods from the Cretaceous/Tertiary boundary interval on the Atlantic Coastal Plain, with a description of the highest ammonite zones in North America. Part 1. Maryland and North Carolina: American Museum Novitates, v. 3454, p. 1–64.
- Landman, N.H., Johnson, R.O., Garb, M.P., Edwards, L.E., and Kyte, F.T., 2007, Cephalopods from the Cretaceous/Tertiary boundary interval on the Atlantic Coastal Plain, with a description of the highest ammonite zones in North America. Part III. Manasquan River Basin, Monmouth County, New Jersey: Bulletin of the American Museum of Natural History, v. 303, p. 1–122.

- Meysman, F.J., Boudreau, B.P., and Middelburg, J.J., 2003, Relations between local, nonlocal, discrete and continuous models of bioturbation: Journal of Marine Research, v. 61, p. 391–410.
- Miller, K.G., Sugarman, P.J., Browning, J.V., Cramer, B.S., Olsson, R.K., de Romero, L., Aubry, M.-P., Pekar, S.F., Georgescu, M.D., Metzger, K.T., Monteverde, D.H., Skinner, E.S., Uptegrove, J., Mullikin, L.G., et al., 1999, Ancora Site (K. G. Miller, P. J. Sugarman, & J. V. Browning, Eds.): Proceedings of the Ocean Drilling Program, Initial Reports, v. 174AX (supplement), p. 1–65.
- Miller, K.G., Sherrell, R., Browning, J.V., Field, M., Gallagher, W., Olsson, R.K., Sugarman, P.J., Tuorto, S., and Wahyudi, H., 2010, Relationship between mass extinction and iridium across the Cretaceous-Paleogene boundary in New Jersey: Geology, v. 38, p. 867–870, doi: 10.1130/G31135.1.
- Norris, R.D., Firth, J., Blusztajn, J., and Ravizza, G., 2000, Mass failure of the North Atlantic margin triggered by the Cretaceous-Paleogene bolide impact: Geology, v. 28, p. 1119–1122.
- Officer, C., and Drake, C., 1985, Terminal Cretaceous environmental events: Science, v. 227, p. 1161–1167.
- Olsson, R.K., Miller, K.G., Browning, J.V., Habib, D., and Sugarman, P.J., 1997, Ejecta layer at the Cretaceous-Tertiary boundary, Bass River, New Jersey (Ocean Drilling Program Leg 174AX): Geology, v. 25, p. 759–762.
- Olsson, R.K., Miller, K.G., Browning, J.V., Wright, J.D., and Cramer, B.S., 2002, Sequence stratigraphy and sea-level change across the Cretaceous-Tertiary boundary on the New Jersey passive margin: Catastrophic events and Mass extinctions: impacts and beyond, v. 356, p. 97–108.
- Premovic, P., I., Ilic, B., S., and Djordjevic, M., G., 2012, Iridium anomaly in the Cretaceous-Paleogene boundary at Højerup (Stevns Klint, Denmark) and Woodside Creek (New Zealand): The question of an enormous proportion of extraterrestrial component: Journal of the Serbian Chemical Society, v. 77, p. 247–255, doi: 10.2298/JSC110404178P.
- Ravizza, G., and Pyle, D., 1997, PGE and Os isotopic analyses of single sample aliquots with NiS fire assay preconcentration: Chemical Geology, v. 141, p. 251–268.
- Robinson, N., Ravizza, G., Coccioni, R., Peucker-Ehrenbrink, B., and Norris, R.D., 2009, A high-resolution marine Os-187/Os-188 record for the late Maastrichtian: Distinguishing the chemical fingerprints of Deccan volcanism and the KP impact event: Earth and Planetary Science Letters, v. 281, p. 159–168, doi: 10.1016/j.epsl.2009.02.019.
- Ross, O.N., and Sharples, J., 2004, Recipe for 1-D Lagrangian particle tracking models in space-varying diffusivity: Limnol. Oceanogr. Methods, v. 2, p. 289–302.

- Ruddiman, W.F., and Glover, L.K., 1972, Vertical mixing of ice-rafted volcanic ash in North Atlantic sediments: Geological Society of America Bulletin, v. 83, p. 2817– 2836.
- Sawlowicz, Z., 1993, Iridium and other platinum-group elements as geochemical markers in sedimentary environments: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 104, p. 253–270.
- Schulte, P., Alegret, L., Arenillas, I., Arz, J.A., Barton, P., Bown, P., Bralower, T.J., Christeson, G., Claeys, P., and Cockell, C. et al., 2010, The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary: Science, v. 327, p. 1214–1218.
- Shukla, P., Bhandari, N., Das, A., Shukla, A., and Ray, J., 2001, High iridium concentration of alkaline rocks of Deccan and implications to K/T boundary: Proc. Indian Acad. Sci. (Earth Planet Sci.), v. 110, p. 103–110.
- Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Sciences, v. 27, p. 75–113.
- Smit, J., and Hertogen, J., 1980, An extraterrestrial event at the Cretaceous–Tertiary boundary: Nature, v. 285, p. 198–200.
- Smit, J., and Kyte, F.T., 1984, Siderophile-rich magnetic spheroids from the Cretaceous– Tertiary boundary in Umbria, Italy: Nature, v. 310, p. 403–405.
- Tanaka, Y., and Franks, P.J.S., 2008, Vertical Distributions of Japanese Sardine (Sardinops Melanostictus) Eggs: Comparison of Observations and a Wind-Forced Lagrangian Mixing Model: Fisheries Oceanography, v. 17, p. 89–100, doi: 10.1111/j.1365-2419.2008.00466.x.
- Toon, O.B., Pollack, J.B., Ackerman, T.P., Turco, R.P., McKay, C.P., and Liu, M.S., 1982, Evolution of an impact-generated dust cloud and its effects on the atmosphere: Special Papers-Geological Society of America, v. 190, p. 187–200.

5.10 Figures and Tables



Fig. 5.1. Geological map of the New Jersey Coastal Plain showing coreholes that sampled the K/Pg boundary. ODP Leg 174AX and ODP Leg 150X cores are shown as red circles. Shallow cores are shown in blue circles.



Fig. 5.2. The Lagrangian sediment mixing model (Hull et al., 2011). Three mixing curves highlight the effect of z_{scale} on depth-dependent diffusivity (K_v). All curves are parameterized with the same z_0 , indicated on the depth axis, and the same K_0 . Larger values for z_{scale} (black) increase the depth range over which K_v asymptotes to zero relative to a smaller z_{scale} values (gray) (Hull et al., 2011, their Fig. 1).



Fig. 5.3. Ir in parts per billion (ppb), fecal pellets in number/gram of sediment, core photographs, lithology, formational assignments for Tighe Park 1 and Bass River coreholes. Note the lowest occurrences of dinocyst taxa at Tighe Park 1 (after Landman et al., 2007) and Bass River (Olsson et al., 1997). *S.: Senoniasphaera*, *D.: Damassadinium*. Modified after (Miller et al., 2010).


Fig. 5.4. Ir in parts per billion (ppb), fecal pellets in number/gram of sediment, core photographs, lithology, formational assignments for Ancora, Double Trouble, and Buck Pit 1 coreholes. Planktonic foraminiferal biozones are indicated for Ancora and Double Trouble. Note the lowest occurrences of dinocyst taxa at Buck Pit 1 (E. Rudolph, 1994, personal commun.; J. Vellekoop, 2014, personal commun.). *S.: Senoniasphaera*, *D.: Damassadinium*.



Fig. 5.5. Ir in parts per billion (ppb), fecal pellets in number/gram of sediment, core photographs, lithology, formational assignments for Search Farm 1, Meirs Farm 1, Inversand, and Fort Monmouth 3 coreholes. Note the lowest occurrences of dinocyst taxa at Search Farm 1, Meirs Farm 1, and Fort Monmouth 3 (J. Vellekoop, 2014, personal commun.). *S.: Senoniasphaera*, *D.: Damassadinium*.



Fig. 5.6. Modeled Ir profiles for Ancora, Double Trouble, and Buck Pit 1 coreholes that provided best fits to the measured Ir profiles. Ir-modeled is plotted with blue crosses. Ir-measured is plotted with solid red circles; the open red circles show Ir samples that were not considered for the calculation of r^2 . The mixing parameters and r^2 values obtained are: Ancora (r^2 : 0.88, K_0 : 1 cm²/kyr, z_{scale} : 4 cm, z_0 : 150 cm); Double Trouble (r^2 : 0.85, K_0 : 10 cm²/kyr, z_{scale} : 9.5 cm, z_0 : 100 cm); Buck Pit 1 (r^2 : 0.80, K_0 : 0.75 cm²/kyr, z_{scale} : 9.5 cm, z_0 : 50 cm). Sedimentation rates (w_s) are 0.2 cm/kyr at Ancora and Buck Pit 1, and 0.3 cm/kyr at Double Trouble. Background values calculated are 0.04 ppb for Ancora and Double Trouble, and 0.11 ppb at Buck Pit 1. mbls: meters below land surface.



Fig. 5.7. Modeled Ir profiles for Search Farm 1, Meirs Farm 1, Tighe Park 1, and Bass River coreholes that provided best fits to the measured Ir profiles. Ir-modeled is plotted with blue crosses. Ir-measured is plotted with solid red circles; the open red circles show Ir samples that were not considered for the calculation of r^2 . The mixing parameters and r^2 values obtained are: Search Farm 1 (r^2 : 96, K_0 : 1000 cm²/kyr, z_{scale} : 1.5 cm, z_0 : 20 cm); Meirs Farm 1 (r^2 : 0.82, K_0 : 1000 cm²/kyr, z_{scale} : 1.5 cm, z_0 : 20 cm); Tighe Park 1 (r^2 : 0.95, K_0 : 10 cm²/kyr, z_{scale} : 4 cm, z_0 : 100 cm); Bass River (r^2 : 0.68, K_0 : 100 cm²/kyr, z_{scale} : 1.5 cm, z_0 : 10 cm). Sedimentation rates (w_s) are 0.2 cm/kyr at Search Farm 1 and Bass River, and 0.3 cm/kyr at Meirs Farm 1 and Tighe Park 1. Background values calculated are 0.07 ppb for Search Farm 1, 0.10 ppb for Meirs Farm 1, 0.11 ppb for Bass River, and 0.12 ppb for Tighe Park 1. mbls: meters below land surface. Note that depth of Bass River is shown in cm above/below the K/Pg boundary.

Table 5.1. Table showing the maximum and minimum concentration of Ir measured in each core, the total thickness of the Ir-enriched interval, background values, and the integrated Ir signal calculated.

Core	maximum concentration of Ir-measured (ppb)	minimum concentration of Ir-measured (ppb)	thickness of the Ir-enriched interval (cm)	background value (ppb)	integrated Ir signal (ppb * m)
Bass River	2.37	0.04	11.00	0.11	0.06
Tighe Park 1	0.46	0.06	30.24	0.12	0.06
Buck Pit 1	0.43	0.09	33.53	0.11	0.06
Double Trouble	0.16	0.04	118.87	0.04	0.06
Ancora	0.13	0.05	106.68	0.04	0.06
Meirs Farm 1	0.49	0.04	10.67	0.10	0.02
Search Farm 1	0.40	0.02	27.43	0.07	0.02
Inversand	0.27	0.03	28.96	0.04	0.02
Meirs Farm 1 Search Farm 1 Inversand	0.49 0.40 0.27	0.04 0.02 0.03	10.67 27.43 28.96	0.10 0.07 0.04	0.02 0.02 0.02





Ir-modeled	r ²	Ws	K ₀	Zscale	Z_0	bv
+	0.92	0.3	5	4	150	0.04
+	0.92	0.3	5	1.5	150	0.04
+	0.89	0.3	10	4	150	0.04
+	0.88	0.2	1	4	150	0.04
+	0.84	0.2	5	9.5	150	0.04

Ir-modeled	r ²	Ws	K ₀	Z _{scale}	\mathbf{Z}_0	bv
+	0.87	0.2	5	9.5	100	0.04
+	0.86	0.2	10	1.5	150	0.04
+	0.85	0.2	10	9.5	150	0.04
-	0.85	0.3	10	9.5	100	0.04
	0.84	0.3	5	4	100	0.04

Ir-modeled	r²	Ws	K ₀	Zscale	\mathbf{z}_0	bv
-	0.80	0.3	0.75	9.5	50	0.11
	0.78	0.3	1	1.5	50	0.11
-	0.77	0.3	1000	1.5	5	0.11
	0.77	0.3	1000	1.5	5	0.11
-	0.75	0.3	0.75	1.5	50	0.11

Fig. 5.S1. Modeled Ir profiles for Ancora, Double Trouble, and Buck Pit 1 coreholes that provided best 5 results in terms of fit to the measured Ir profiles. Ir-modeled is plotted with blue crosses. Ir-measured is plotted with solid red circles; the open red circles show Ir samples that were not considered for the calculation of r^2 . The mixing parameters, r^2 values, sedimentation rates, and background values are shown in tables below each graph. mbls: meters below land surface, by: background value.



11 11000100	1c	Ws	N0	Zscale	Z0	00
+	0.96	0.2	1000	1.5	20	0.07
-	0.93	0.3	1000	1.5	20	0.07
+	0.90	0.3	1000	4	20	0.07
-	0.90	0.2	1000	4	20	0.07
-	0.90	0.2	1000	4	20	0.07

Ir-modeled	r ²	ws	K ₀	Zscale	\mathbf{z}_0	bv
—	0.85	0.3	250	1.5	2	0.11
	0.85	0.2	1000	1.5	20	0.11
-	0.84	0.2	250	1.5	2	0.11
—	0.82	0.3	1000	1.5	20	0.11
—	0.81	0.3	1000	4	10	0.11

Ir-modeled	r²	ws	K ₀	Zscale	Z0	bv
—	0.95	0.3	10	4	100	0.12
—	0.95	0.2	1	4	100	0.12
-	0.94	0.2	10	9.5	100	0.12
	0.94	0.3	10	9.5	100	0.12
-	0.92	0.2	10	4	100	0.12

Ir-modeled	r²	ws	K ₀	Zscale	Z0	bv
-	0.71	0.3	10	1.5	10	0.11
	0.68	0.3	100	1.5	2	0.11
-	0.68	0.3	1	1.5	5	0.11
-	0.68	0.3	10	1.5	5	0.11
	0.68	0.2	100	1.5	10	0.11

Fig. 5.S2. Modeled Ir profiles for Search Farm 1, Meirs Farm 1, Tighe Park 1, and Bass River coreholes that provided best 5 results in terms of fit to the measured Ir profiles. Ir-modeled is plotted with blue crosses. Ir-measured is plotted with solid red circles; the open red circles show Ir samples that were not considered for the calculation of r^2 . The mixing parameters, r^2 values, sedimentation rates, and background values are shown in tables below each graph. mbls: meters below land surface, by: background value. Note that depth of Bass River is shown in cm above/below the K/Pg boundary.

Chapter 6

Conclusions and Future Directions

The unifying goal of the research presented in this dissertation is to provide new insights into different aspects of the K/Pg boundary. During the K/Pg boundary transition the Earth witnessed significant biotic, paleoceanographic, and geochemical changes. My research elucidates some aspects of these changes with a multidisciplinary approach. Examination of shallow marine sections with planktonic foraminiferal biostratigraphy, sedimentology, sequence stratigraphy, as well as geochemistry and their correlation with deep-sea sections yielded unique constraints on several issues. I summarize primary conclusions of my thesis below under three main headings and conclude with some thoughts on the future directions.

6.1 Stratigraphy and Sea-Level Changes across the K/Pg Boundary

The biostratigraphic record showed that the extinction of the planktonic foraminifera across the K/Pg boundary was abrupt and catastrophic in the Haymana Basin, Turkey, similar to other K/Pg boundary records globally (e.g., Olsson and Liu, 1993; Olsson et al., 1999). Planktonic foraminiferal Zone P0 is characterized by the evolution of early Danian planktonic foraminifera and associated with an increase in clay minerals and echinoid fecal pellets. An early Danian prominent increase in echinoid fecal pellets in a Tethyan section (Haymana Basin, Turkey) and numerous New Jersey Coastal Plain cores appears to be a correlatable level indicating an important environmental change after the K/Pg mass extinction. This stratigraphic horizon can be used as a global marker for the K/Pg boundary in addition to other criteria defining the boundary.

There is no consensus on the global sea-level change across the K/Pg boundary, possibly due to the fact that changes were subtle. Sea-level studies on sections close to the impact site interpreted the coarse-grained siliciclastic and limestone breccia deposits at the K/Pg boundary as lowstand deposits (Donovan et al., 1988; Keller and Stinnesbeck, 1996; Keller, 2007). However, this interpretation has been challenged by the fact that sedimentation in these regions was affected by the impact-generated tsunami (e.g., Smit, 1999) and they have no relation to eustasy (e.g., Claeys et al., 2002). In fact, distal sites free from stratigraphic complexities indicate different sea-level histories. The K/Pg boundary sections in the Basque Region (Apellaniz et al., 1997; Pujalte, 1998), Spain (Pardo et al., 1996; Alegret et al., 2003), Kazakhstan (Pardo et al., 1999), and Tunisia (Adatte et al., 2002) show that the K/Pg event occurred during a sea-level rise. Relative sea-level curves I present for the Haymana section, Turkey and the Campo Pit section, New Jersey based on detailed facies and sequence stratigraphic analysis show that the K/Pg boundary is recorded in the latest part of a transgressive systems tract, below a maximum flooding surface. Having similar trends in sea-level changes across the K/Pg boundary in geographically different regions suggests that the K/Pg impact occurred during a global sea-level rise.

6.2 Perturbations in the Marine Carbon Cycle across the K/Pg Boundary

Although previous studies explaining the perturbations in the marine carbon cycle associated with the K/Pg boundary have focused on deep-sea sections, I tested the three main hypotheses in a paleo-continental shelf, where $\delta^{13}C$ values and organic carbon accumulation rates can help distinguish among the hypotheses: The Strangelove Ocean, with near cessation of primary productivity; the Living Ocean, with greatly reduced export productivity; and the Resilient Ocean, with little or no change in productivity. Each hypothesis predicts distinctive δ^{13} C gradients from the photic zone to deep-water in a shelf setting. The New Jersey paleoshelf recorded a ~2.5‰ δ^{13} C decrease in bulk carbonate, a ~0.8‰ $\delta^{13}C$ decrease in organic carbon, a collapse of the vertical $\delta^{13}C$ gradient between planktonic and benthic foraminifera, and a significant decrease in organic carbon mass accumulation rates. Similar to deep-sea isotope records, the trophic structure of the planktonic foraminifera in the water column was lost, and no planktonic for a ninifera occupied the surface layer in the early Danian. A ~1.0‰ water column δ^{13} C gradient in planktonic foraminifera and a ~0.75% cross-shelf δ^{13} C gradient in benthic foraminifera in the early Danian indicate the presence of active primary productivity, strongly supporting the Living Ocean hypothesis.

The interbasinal deep-sea benthic foraminiferal δ^{13} C gradients between the Pacific (ODP Site 1210) and the Atlantic (ODP Site 1262) Oceans are also most consistent with the Living Ocean hypothesis. Findings of my research substantiate that the interbasinal benthic foraminiferal δ^{13} C gradient strongly depends on the age models used to correlate the sites. The age model of Westerhold et al. (2008), which was also used by Alegret et

al. (2012), and the age model of Hilgen et al. (2010) show that the benthic foraminiferal δ^{13} C gradient was maintained after the K/Pg boundary between the South Atlantic and North Pacific, supporting the Resilient Ocean hypothesis. However, new age models I constructed for the sites based on magnetostratigraphy and biostratigraphy, also supported independently by the Fe-count correlation, shows that benthic foraminiferal δ^{13} C gradient was reduced.

Integration of the New Jersey paleoshelf and deep-sea δ^{13} C records are most consistent with the Living Ocean hypothesis, where marine productivity was relatively high, but the fraction of total productivity that sank to the deep-sea was reduced. However, I also acknowledge that there might have been no or little change in export productivity in highly productive regions like the equatorial Pacific (Alegret and Thomas, 2009; Hull and Norris, 2011; Alegret et al., 2012). This heterogeneity in response should not be interpreted as a weakness in the argument of this research, but points to the complexity of reconstructing productivity variations globally.

6.3 Iridium Profiles and Delivery at the K/Pg Boundary

The shape of the Ir profiles across the K/Pg boundary and the maximum Ir concentrations measured are highly variable in shallow shelf cores on the NJCP, although they were deposited closely spaced geographically. The shape of the Ir profiles and the maximum concentrations measured appear to be determined primarily by sedimentary and geochemical processes, especially bioturbation. Evidence supporting sediment mixing includes burrows visible in the cores and good fits between a mixing model and

measured Ir profiles. Using the Lagrangian advection-diffusion model (Hull et al., 2011), I was able to show the extent of mixing effects on the Ir records quantitatively across the K/Pg boundary. Modeling and fitting the model to the measured iridium profiles showed that Ir anomaly shapes could generally be explained by physical sediment mixing. Hull et al. (2011) showed that the Lagrangian advection-diffusion model successfully explains Ir profiles in a distal deep-sea site (Site 577B) with a less complicated mixing history. Our modeling effort in shallow water cores in an intermediate distance from the impact site showed that the model is also applicable in the sites with much more complicated mixing histories. However, the mixing model should be applied carefully taking the background values into account, where there is low signal-to-noise ratio with low to moderate Ir concentrations. Furthermore, for complex profiles the model occasionally fails to fit secondary peaks that are associated with the sediment mixing (Ruddiman and Glover, 1972).

Another important result I present in my dissertation is that the integrated Ir signal in the majority of cores is the same, strongly suggesting that the total Ir delivery to this region was similar at the time of the K/Pg event. Although having received comparable amounts of Ir, the sites show a different maximum concentration of Ir, which is inversely proportional to the total thickness of the Ir-enriched interval providing another evidence of diffusion of Ir after the original time of deposition. It has been often thought that low Ir concentrations recorded in many geological boundaries do not show necessarily an extraterrestrial impact, but indicate remobilization and accumulation of Ir at redox boundaries delivered by cosmic influx when sedimentation rates were low. However, this study shows that low Ir concentrations ($\sim 0.1 - 0.5$ ppb) can be a product of the same Ir delivery as nearby high peaks (~ 2.4 ppb).

6.4 Future Directions

While this study provided valuable insight regarding the K/Pg boundary, the results obtained can be the foundation for the next generation of research. In order to enhance our interpretations on the δ^{13} C changes across the K/Pg boundary, the survivor planktonic foraminiferal taxa *Guembelitria cretacea* can be analyzed to unequivocally test whether the collapse in δ^{13} C gradient is an artifact of taxonomic changes and vital effects across the boundary. *Guembelitria cretacea*, a microperforate planktonic foraminiferal taxa often considered as "disaster" species, is hypothesized to have given rise to four genera (*Globoconusa, Parvularugoglobigerina, Woodringina*, and *Chiloguembelina*) (Olsson et al., 1999). Its range and relative dominance expanded following the end-Cretaceous extinction (Keller and Pardo, 2004). The New Jersey margin cores seem to have sufficient specimens of survivor taxa to test the hypothesis that the collapse of the vertical δ^{13} C gradient was not an artifact due to taxonomic changes after the extinction as proposed by Alegret et al. (2012).

It will be very interesting to study the recovery period of δ^{13} C gradients between planktonic and benthic foraminifera and the redevelopment of planktonic foraminiferal trophic structure (shallow, intermediate, and deep dwellers). This appears to have occurred in stages after the extinction in a paleoshelf setting. Preliminary data in New Jersey shallow cores show that vertical δ^{13} C gradient started to recover in planktonic foraminiferal Zone P1c, but the full recovery in vertical gradient and the redevelopment of trophic structure of planktonic foraminifera took place more than 3 Myr after the K/Pg boundary. This seems to be similar to deep-sea records (D'Hondt et al., 1998; Coxall et al., 2006), but detailed work in shelf and its correlation to deep-sea records may reveal new aspects of the recovery period.

Measurements of other platinum group elements such as Os, Pt, Ru, Rh, and Pd in addition to Ir may provide more insights into the distribution of Ir due to the processes after the time of original deposition, source of the enrichments, and the positions of the enrichments relative to the biostratigraphic K/Pg boundary. Likewise, analyses of redox sensitive elements such as Mn, Cd, Mo, V, Re and U with varying degrees of mobility would be very useful to further test the remobilization hypotheses, especially across two sections (Tighe Park 1 and Bass River), where the Ir has been inferred to have moved by geochemical remobilization.

- Adatte, T., Keller, G., and Stinnesbeck, W., 2002, Late Cretaceous to early Paleocene climate and sea-level fluctuations: the Tunisian record: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 178, p. 165–196.
- Alegret, L., and Thomas, E., 2009, Food supply to the seafloor in the Pacific Ocean after the Cretaceous/Paleogene boundary event: Marine Micropaleontology, v. 73, p. 105– 116, doi: 10.1016/j.marmicro.2009.07.005.
- Alegret, L., Molina, E., and Thomas, E., 2003, Benthic foraminiferal turnover across the Cretaceous/Paleogene boundary at Agost (southeastern Spain): paleoenvironmental inferences: Marine Micropaleontology, v. 48, p. 251–279.
- Alegret, L., Thomas, E., and Lohmann, K.C., 2012, End-Cretaceous marine mass extinction not caused by productivity collapse: Proceedings of the National Academy of Sciences of the United States of America, v. 109, p. 728–732, doi: 10.1073/pnas.1110601109/-/DCSupplemental/ST01.doc.
- Apellaniz, E., Baceta, J.I., Bernaola-Bilbao, G., Nunez-Betelu, K., Orue-Etxebarria, X., Payros, A., Pujalte, V., Robin, E., and Rocchia, R., 1997, Analysis of uppermost Cretaceous-lowermost Tertiary hemipelagic successions in the Basque Country (western Pyrenees); evidence for a sudden extinction of more than half planktic foraminifer species at the K/T boundary: Bulletin de la Societe Geologique de France, v. 168, p. 783–793.
- Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of Chicxulub ejecta at the Cretaceous-Tertiary boundary (C. Koeberl & K. G. MacLeod, Eds.): Catastrophic Eevents and Mass extinctions: Impacts and Beyond: Geological Society of America Special Paper, v. 356, p. 55–68.
- Coxall, H.K., D'Hondt, S., and Zachos, J.C., 2006, Pelagic evolution and environmental recovery after the Cretaceous-Paleogene mass extinction: Geology, v. 34, p. 297– 300, doi: 10.1130/G21702.1.
- D'Hondt, S., Donaghay, P., Zachos, J.C., Luttenberg, D., and Lindinger, M., 1998, Organic carbon fluxes and ecological recovery from the Cretaceous-Tertiary mass extinction: Science, v. 282, p. 276–279, doi: 10.1126/science.282.5387.276.
- Donovan, A.D., Baum, G.R., Blechschmidt, G.L., Loutit, T.S., Pflum, C.E., and Vail, P.R., 1988, Sequence stratigraphic setting of the Cretaceous-Tertiary boundary in Central Alabama. In Wilgus, C.K., Kendall, C.G.S.C., Posamentier, H.W., and Ross, C.A., eds., Sea-level changes: An integrated approach: Society of Economic Paleontologists and Mineralogists Special Publication, v. 42, p. 299–307.

- Hilgen, F.J., Kuiper, K.F., and Lourens, L.J., 2010, Evaluation of the astronomical time scale for the Paleocene and earliest Eocene: Earth and Planetary Science Letters, v. 300, p. 139–151, doi: 10.1016/j.epsl.2010.09.044.
- Hull, P.M., and Norris, R.D., 2011, Diverse patterns of ocean export productivity change across the Cretaceous-Paleogene boundary: New insights from biogenic barium: Paleoceanography, v. 26, p. PA3205, doi: 10.1029/2010PA002082.
- Hull, P.M., Franks, P.J.S., and Norris, R.D., 2011, Earth and Planetary Science Letters: Earth and Planetary Science Letters, v. 301, p. 98–106, doi: 10.1016/j.epsl.2010.10.031.
- Keller, G., 2007, The Chicxulub impact and K-T mass extinction in Texas: Bulletin of South Texas Geological Society, v. XLVII, p. 15–44.
- Keller, G., and Pardo, A., 2004, Disaster opportunists Guembelitrinidae: index for environmental catastrophes: Marine Micropaleontology, v. 53, p. 83–116.
- Keller, G., and Stinnesbeck, W., 1996, Sea-level changes, clastic deposits, and megatsunamis across the Cretaceous–Tertiary boundary: In: MacLeod, N., Keller, G. (Eds.), Cretaceous–Tertiary Mass Extinctions: Biotic and Environmental Change. Norton, London, p. 415–449.
- Olsson, R.K., and Liu, C., 1993, Controversies on the placement of Cretaceous-Paleogene boundary and the K/P mass extinction of planktonic foraminifera: Palaios, v. 8, p. 127–139.
- Olsson, R.K., Hemleben, C., Berggren, W.A., Huber, B.T., Editors and Members of the Paleogene Planktonic Foraminifera Working Group, 1999, Atlas of Paleocene Planktonic Foraminifera: Smithsonian Contributions to Paleobiology, v. 85, p. 252.
- Pardo, A., Adatte, T., Keller, G., and Oberhansli, H., 1999, Paleoenvironmental changes across the Cretaceous-Tertiary boundary at Koshak, Kazakhstan, based on planktic foraminifera and clay mineralogy: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 154, p. 247–273.
- Pardo, A., Ortiz, N., and Keller, G., 1996, Latest Maastrichtian and Cretaceous-Tertiary Boundary Foraminiferal Turnover and Environmental Changes at Agost, Spain (N. McLeod & G. Keller, Eds.): Cretaceous-Tertiary Mass Extinctions: Biotic and Environmental Changes Norton, New York/London, p.139-171.
- Pujalte, V., 1998, Paleocene Strata of the Basque Country, Western Pyrenees, Northern Spain: Facies and Sequence Development in a Deep-Water Starved Basin: Mesozoic and Cenozoic Sequence Stratigraphy of European Basin, SEPM Special Publications, v. 60, p. 311–325.

- Ruddiman, W.F., and Glover, L.K., 1972, Vertical mixing of ice-rafted volcanic ash in North Atlantic sediments: Geological Society of America Bulletin, v. 83, p. 2817– 2836.
- Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Sciences, v. 27, p. 75–113.
- Westerhold, T., Röhl, U., Raffi, I., Fornaciari, E., Monechi, S., Reale, V., Bowles, J., and Evans, H., 2008, Astronomical calibration of the Paleocene time: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 257, p. 377–403.