Impact effects and regional tectonic insights: Backstripping the Chesapeake Bay impact structure

Travis Hayden*, Michelle Kominz — Department of Geosciences, Western Michigan University, 1187 Rood Hall, 1903 W. Michigan Avenue, Kalamazoo, Michigan 49008, USA
David S. Powars, Lucy E. Edwards — United States Geological Survey, National Center, Reston, Virginia 20192, USA
Kenneth G. Miller, James V. Browning, Andrew A. Kulpecz — Department of Geosciences, Rutgers, The State University of New Jersey, Piscataway, New Jersey 08854, USA

ABSTRACT

The Chesapeake Bay impact structure is a ca. 35.4 Ma crater located on the eastern seaboard of North America. Deposition returned to normal shortly after impact, resulting in a unique record of both impact-related and subsequent passive margin sedimentation. We use backstripping to show that the impact strongly affected sedimentation for 7 m.y. through impact-derived crustal-scale tectonics, dominated by the effects of sediment compaction and the introduction and subsequent removal of a negative thermal anomaly instead of the expected positive thermal anomaly. After this, the area was dominated by passive margin thermal subsidence overprinted by periods of regional-scale vertical tectonic events, on the order of tens of meters. Loading due to prograding sediment bodies may have generated these events.

Keywords: impact processes, passive margin, Chesapeake Bay, tectonics, Eocene, backstripping.

PREVIOUS WORK

The Chesapeake Bay impact structure (Fig. 1) was identified from core drilling and seismic reflection studies in Chesapeake Bay region (e.g., Powars et al., 1992, 1993; Poag et al., 1992, 1994). The Chesapeake Bay impact structure correlates with the North American tektite field, which has been dated as 34.3–35.5 Ma (Glass, 1989; Koeberl, 1989; Poag et al., 1994; Horton and Izett, 2006). At 85 km in diameter, this crater is the largest known crater in the United States, the seventh largest known on Earth, and one of the best preserved marine impact craters. The most recent studies include several continuously cored locations, both inside and outside the crater (Fig. 1; e.g., Poag 1997; Powars and Bruce, 1999; Powars, 2000; Powars et al., 2005; Edwards et al., 2005). These studies provided chronological, paleobathymetric, and lithologic constraints on the post-impact sediment packages.

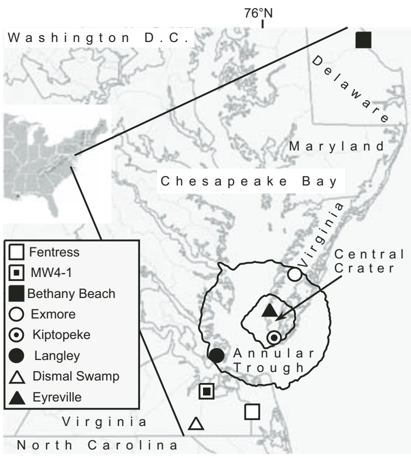

Figure 1. Location map showing Chesapeake Bay impact structure and coreholes used in this study.

Previous studies of impact processes at this and other impact locations have focused on immediate impact-related events, and were limited to hours and or years just after impact (e.g., Melosh and Ivanov, 1999; Melosh, 1989). In the Chesapeake Bay impact structure, the preservation of the impact-related sequences, as well as the immediate post-impact sequences, provides a record of the effects of the impact on time scales of millions of years. In addition, the impact structure provides a relatively complete record of the post-impact, normal, shallow-marine sedimentation.

METHODS

In this work we use backstripping to estimate the subsidence of the basement under water (Bond and Kominz, 1984). Backstripping removes the variable sedimentation by compensating for the subsidence caused by the sediment load. This yields both tectonic and eustatic basement subsidence.

Backstripping quantitatively estimates the depth of the hole or accommodation space that is filled with a combination of sediments and water. The thickness of the sediments deposited provides a limit on the minimum amount of subsidence that took place. However, the present-day sediment thickness must be decompacted to estimate its thickness at the time of deposition. This is done using porosity-depth curves generated from nearby cores. We used compaction curves based on New Jersey Coastal Plain cores (Van Sickel et al., 2004). Because these curves are based on sediment composition, detailed lithology is also required. Isostatic unloading of the sediment yields an estimate of how much of the hole was filled with sediment. In order to estimate the total subsidence, an estimate of paleowater depth must also be made. This water depth estimate is based on inferred depth preferences of selected benthic organisms. Organisms that live in shallower water depths tend to have a more limited range of suitable habitat, while organisms that live deeper tend to have wider habitable zones. This leads to increasing uncertainty in water depth ranges for increasingly deeper water facies translating directly into increasing error ranges for subsidence estimates.

The steps outlined above are codified in the following equation (modified from Steckler and Watts, 1978):

$$TS = S^* \frac{(\rho_m - \rho_s)}{(\rho_m - \rho_w)} - \Delta SL \frac{(\rho_m)}{(\rho_m - \rho_w)} + Wd. \quad (1)$$

This equation allows for the calculation of tectonic subsidence (TS) from decompacted sediment thicknesses (S*), densities (ρ_s), and water depths (Wd), given the change in sea level (ΔSL). The values of ρ_m and ρ_w represent the density of the mantle ($\rho_m = 3.33$ g/cm^3) and seawater ($\rho_w = 1.03$ g/cm^3). S*, ρ_s, and Wd are the unknowns that can be estimated from a sedimentary section.

Our analysis of backstripping results is time-dependent, requiring numerical age estimates of the sediment packages. In the case of this study, low age resolution makes the study of only slower (a million years or more) processes possible.

The goal of our backstripping is to determine the subsidence of the basement under water. This is the first reduction, or R1, of Bond et al. (1989):

*E-mail: t4hayden@wmich.edu.

$$
\begin{aligned}
R1 &= S * \frac{(\rho_m - \rho_S)}{(\rho_m - \rho_w)} + Wd \\
&= TS + \Delta SL \frac{(\rho_m)}{(\rho_m - \rho_w)}.
\end{aligned} \tag{2}
$$

It is important to note that this R1 curve includes both eustatic and tectonic signals. R1 data can be compared to a theoretical model of passive margin subsidence based on the work of McKenzie (1978) or Royden and Keen (1980). These are theoretical models of the expected amount and timing of subsidence due to thermal cooling of stretched continental lithosphere. As the Chesapeake Bay impact structure is underlain by the rifted passive margin of North America, these theoretical models are applicable. The general forms of the R1 backstripping results are similar to the McKenzie (1978) models (Fig. 2A). The difference between expected subsidence from thermal cooling and the actual R1 subsidence recorded by the sediments is a combination of eustatic and non-thermal tectonic effects. The interpretation of these results is the basis of this paper and we refer to more or less R1 subsidence in comparison to the predicted McKenzie (1978) model of thermal subsidence (MK means) (e.g., Figs. 2A, 2B).

DATA

The input data sets were collected and interpreted from several cores drilled inside and outside the impact crater (Fig. 1). This duality allows comparison between crater-exterior and crater-interior subsidence signals. Of the numerous cores located throughout the area, only seven cores provide sufficient age, lithologic, and paleodepth data to perform backstripping analysis. These cores included four cores located outside the crater, and three inside the crater (Fig. 1) that were drilled between 1986 and 2006 (Powars et al., 1992, 2005; Edwards et al., 2005; Powars, 2000; Powars and Bruce, 1999; Mixon, 1985). We reconfigured the existing data into a framework suitable for backstripping.

One of the major limiting factors for core selection was paleoenvironmental data. For the purposes of this study, previous work provided paleoenvironmental indicators. When possible, the authors of those works were contacted to better understand and interpret these into a framework suitable for backstripping.

The other major limiting factor for selection of cores was availability of age estimates. Relative dates based on strontium isotope techniques were used to backstrip the Oligocene–Miocene sections of the Bethany Beach core (Browning et al., 2006) and the Miocene in the Kiptopeke core (Powars and Bruce, 1999). In most cases, however, age estimates were based on biostratigraphy. These age estimates were made using the Gradstein et al. (2004) biostratigraphic time scale.

RESULTS

Of the four R1 curves from outside the impact crater, three show remarkable internal consistency in the magnitude and timing of subsidence (Fig. 2B). At the time of the impact none of these cores recorded deposition (the Bethany Beach core, however, did not penetrate sediments older than ca. 28 Ma.) An excess subsidence event is recorded in the Bethany Beach core (Browning et al., 2006) beginning at 22 Ma. This event began at 14 Ma in the MW4–1 core, and at 13 Ma in the Fentress core. Finally, starting ca. 2.5 Ma subsidence decreased in both the Fentress and MW4–1 cores, suggesting a small-scale uplift of the region.

From 20 Ma to the present, the subsidence history of the cores located within the crater is similar to that of the cores located outside the crater. These cores also record several events immediately post-impact (Fig. 3). A vertical line just before 35 Ma on the R1 plot (Fig. 3, event A) records the dramatic, instantaneous removal of the pre-impact sediment pile, followed by virtually instantaneous redeposition for each of the crater-interior cores. None of the cores analyzed were located within the centermost portion of the crater that also underwent removal of part of the upper crust. The scale of the diagram does not allow for a detailed view of this event, which occurred in hours or days (Poag et al., 1994). All three of the crater-interior cores record higher than expected subsidence rates after impact, that decrease into a hiatus ~3 m.y. after impact (Fig. 3, event B).

The rapid subsidence event was followed by a dramatic uplift event ranging between 50 m and 125 m in all three crater-interior cores (Fig. 3). The uplift is documented by an unconformity (Fig. 3, event C), and deposition (Fig. 3, event D) above the hiatus was shallower than below the hiatus. The uplift magnitudes range from 100 ± 5 m recorded at Kiptopeke to between 50 and 125 m at Langley. These uplifts occurred between 0.8 and 2.9 m.y. in Kiptopeke and Langley, respectively. Subsequent to the uplift event, Exmore and Kiptopeke both record unconformities between 28.6 and 18.8 Ma and between 31.7 and 16.5 Ma, respectively. The Langley core documents slow deposition throughout this time. Improved age control for this core may reveal hiatuses during this interval (Fig. 3, event D). Note that the Langley core shows evidence of water depth changes at higher frequency than we can analyze, due to limited age control during this time frame (Powars et al., 2005).

Sedimentation resumed ca. 18.8 Ma in the Exmore core, and began progressively later to the south (Fig. 3, event E). The subsidence recorded in event E is ~20–30 m between 19 and 5 Ma and is seen both inside and outside of the crater. However, both the Exmore and Kiptopeke cores record a period of increased subsidence rate before an unconformity. Thus, these cores suggest a larger and less continuous subsidence event in comparison to tailing off into the more general regional, crater-exterior trend. The Exmore core shows the largest subsidence event E, with a magnitude of ~75–100 m between 18.8 and 14.3 Ma. The Kiptopeke core records a similar amount of excess subsidence but it starts later, at 16.5 Ma, and occurs in two events, one from 16.5 to

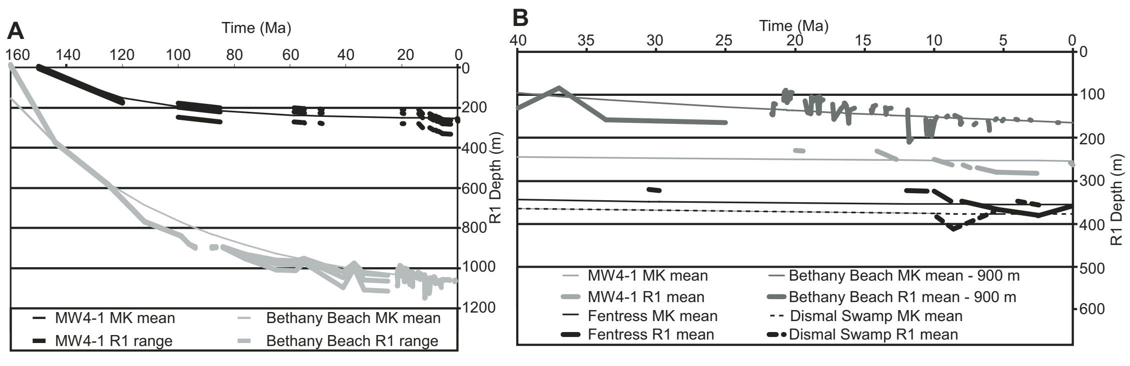

Figure 2. A: R1 plots for two representative crater-exterior cores. Heavier lines represent range of R1 results due to water depth uncertainty. MK means represent the McKenzie (1978) best-fit thermal models to the R1 curves assuming best-estimate water depths. Subsidence pattern for both cores is dominantly passive margin type. B: Expansion of the part of A from 40 Ma to present. Water depth ranges have been removed to aid readability, and all four crater-exterior cores are represented. Subsidence patterns of Bethany Beach, MW4–1, and Fentress are quite similar, while the trends at Dismal Swamp do not follow those of these wells.

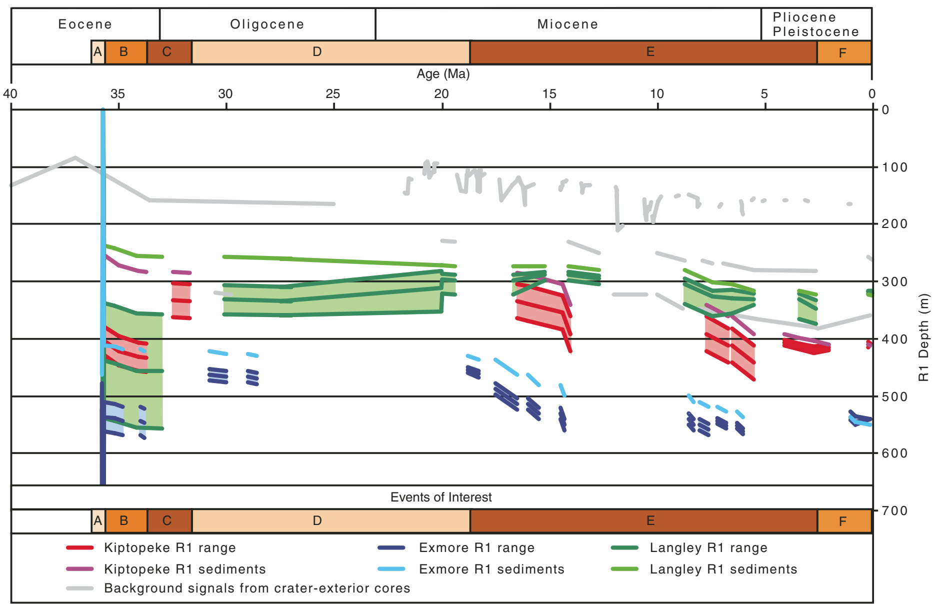

Figure 3. Results for the three crater-interior cores. Background crater-exterior results are shown in gray for comparison. R1 (first reduction) sediment lines illustrate only subsidence represented by the sediments without any water depth information. R1 ranges indicate the subsidence when water depth changes are taken into account. Event A represents the impact event. Events B and C are various stages of the post-impact response inside crater. Events E and F are regional.

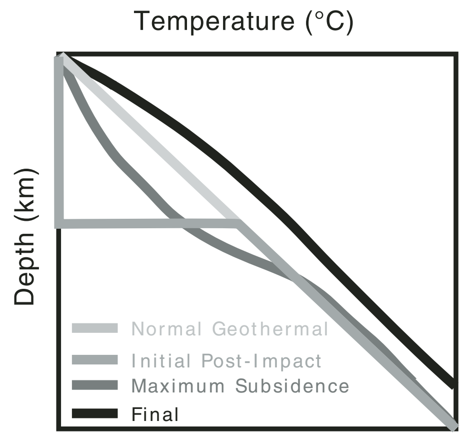

Figure 4. Theoretical temperature versus depth plots at key stages. The negative thermal anomaly is generated by rapid sedimentation and occurs immediately after impact. Initially, heating of the sediments results in cooling of the uppermost crust. This occurred ~3 m.y. after impact and represents maximum thermal subsidence. The final post-impact line has higher temperatures than the initial profile due to thermal blanketing of sediments that are still more loosely compacted than those originally present and is ~7 m.y. after impact.

14 Ma, accounting for 55 m of subsidence, and the other from 7.7 to 5.5 Ma, accounting for 50 m of additional subsidence.

Between 5 and 2.5 Ma, all three of the crater-interior cores record an unconformity leading to a thin veneer of recent sediments (Fig. 3, event F). This suggests a slight regional uplift, which is also seen in two crater-exterior cores, MW4–1 and Fentress.

INTERPRETATIONS

The subsidence histories of the crater-exterior cores are remarkably similar, with one exception. The Dismal Swamp core is located farthest south (Fig. 1), and subsided differently from the other coreholes (Fig. 2B). This core is located near the Norfolk Arch and may belong to a different tectonic regime than the rest of the cores (Powars et al., 1992).

The similarity in subsidence among the remaining crater-exterior cores suggests that these cores record the general subsidence trend of the Chesapeake Bay region (Fig. 2B). They provide a framework for comparison with the crater-interior results that allows us to separate impact-related and non-impact-related processes (Fig. 3).

Inside the crater, the subsidence was more complex than outside. A large impact would be expected to generate crustal-scale heating of the basement. This would result in initial uplift, followed by cooling, producing accommodation space in excess of the background, thermal subsidence. What actually occurred was initial subsidence, followed by uplift (Fig. 3). In detail, the immediate post-impact event (event B in Fig. 3) is characterized by rapid subsidence that tails off to nondeposition and/or erosion by ~3 m.y. after impact. The subsequent uplift event is indicated by deposition at a shallower level (Fig. 3, event C).

We attribute the immediate post-impact subsidence (event B) to the rapid compaction of the impact breccia. Having been deposited in a few tens of minutes to hours (Collins and Wuennemann, 2005), these sediments must have been poorly consolidated so that subsequent compaction was time-dependent (Schmoker and Gautier, 1989).

The subsequent uplift event (Fig. 3, event C) suggests crustal heating. We postulate that a crustal-scale negative thermal anomaly was caused by the deposition of cold sediments, resetting the geothermal gradient (Fig. 4). This first cooled the upper crust, adding to the accommodation space generated immediately after the impact (event B). Subsequently the upper crust and overlying sediment package reheated (Fig. 4), generating uplift. The exact interaction between subsidence due to sediment compaction and the effects of the crustal-scale thermal event is a topic for future modeling. Each of the three crater-interior records shows slightly different timing and magnitudes of uplift (Fig. 3, event C). We interpret this to be due to differential sediment properties and thicknesses, resulting in differing amounts of compaction and thermal blanketing.

The thermal anomaly due to the impact was removed by ca. 30 ± 2 Ma. From 30 to 20 Ma a hiatus dominated in the study region. Low average sedimentation rates are observed in the Langley core during this time interval, which records a slight uplift (Fig. 3, event D). However, the requisite age and paleowater depth data lack the precision required to quantify this interpretation through backstripping.

The first regional post-impact event occurred progressively from north to south. Excess subsidence started at Bethany Beach at 22.0 Ma (Browning et al., 2006), and proceeded southward to Exmore at 20.5 Ma, then to Kiptopeke at 16.5 Ma, to Langley at 14.5 Ma, to MW4–1 at 14.0 Ma, and to Fentress at 13.0 Ma (Fig. 2B and event E in Fig. 3). Between 22 and 5 Ma, this event generated ~20–30 m of subsidence, in excess of that expected in this passive margin setting. Event E is interpreted as the flexural response of the basement to the deposition of a large sedimentary load (Browning et al., 2006) that prograded from north to south along the coast. Marine seismic reflection surveys conducted in the Chesapeake Bay and offshore are consistent with this interpretation. However, the profiles are not high resolution and do not provide definitive details of the southward progradation (Powars and Bruce, 1999).

A small-scale uplift is recorded in all of the crater-interior cores and two of the crater-exterior cores (Fig. 3, event F). This uplift is of unknown origin, and appears to be ~2–20 m. It shows no clearly identifiable trends with respect to location or magnitude.

CONCLUSIONS AND FUTURE WORK

Backstripping of the Chesapeake Bay impact structure has increased our understanding of both long-term impact-related processes and normal tectonic and depositional processes that prevail along a passive margin. Our results show that impact-related tectonism is not a major contributor to depositional patterns after ~7 m.y. We postulate that accommodation space was generated by sediment compaction and the introduction and subsequent removal of a negative thermal anomaly. After the impact effects ceased, the crater interior operated as a part of the surrounding passive margin, which, in this case, was quite complex. These results show that a bolide impact can add unusual complexities to an already complex late-stage passive margin.

Future work in this area is in progress and includes numerical forward modeling of the negative thermal anomaly. This modeling work focuses on the 10 m.y. immediately following the impact, and focuses on the interaction between the negative thermal anomaly, the sediment compaction, and the observed sedimentation patterns (Fig. 4). In addition, detailed high-resolution analysis and the backstripping of the Eyreville A/B and C cores acquired in 2005–2006 will yield insight about the subsidence history within the central crater (Gohn et al., 2006).

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