

Structure of the Chesapeake Bay Impact Crater from Wide-Angle Seismic Waveform Tomography

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ABSTRACT

The Chesapeake Bay impact structure is one of the largest and most well preserved impact structures on Earth. It has a unique morphology composed of an inner crater penetrating crystalline basement surrounded by a wider crater in the overlying sediments. In 2004, the U.S. Geological Survey conducted a seismic survey with the goals of constraining crater structure and in support of the drilling of a borehole into the deepest part of the crater. Travel-time and waveform inversion were applied to the data to produce a high-resolution velocity model of the crater. Low-fold reflection processing was also applied. Northeast of the crystalline crater, undeformed, eastward-sloping crystalline basement is ~1.5 km deep. The edge of the inner crater is at ~ 15 km radius and slopes gradually down to a depth of 1.5 – 1.8 km. A central peak of 4-5 km radius rises to a depth of ~0.8 km. Basement velocity in the crystalline crater is much lower than undeformed basement, which suggests ~10% fracturing of the crater floor, and up to 20% fracturing of the central uplift. A basement uplift and lateral change of velocity, interpreted as the edge of the transient crater, occurs at a radius of ~ 11 km. Assuming a 22 km diameter transient crater, scaling laws predict a ~30 km diameter crater and central peak diameter of 8-10 km. This indicates that post-impact collapse processes that created the ~ 30 km diameter crystalline crater were unaffected by the much weaker rheology of the overlying sediments.

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Introduction

The 35 Ma old Chesapeake Bay impact structure (CBIS) (Figure 1) is one of the largest impact structures on Earth. Its 85km diameter makes it the 7th largest recognized crater on Earth (Grieve and Shoemaker, 1994), and its young age and location on a passive margin make it one of the best preserved. Despite its size and pristine nature, it was recognized as an impact structure just 15 years ago, when shocked minerals were identified in core from an offshore borehole (Powars et al., 1993; Poag et al., 1994). Since then, geophysical and borehole data have been analyzed to understand the structure the crater (Powars et al. 1993; Poag 1997; Poag 1999; Shah et al., 2005). The impact occurred in a target of strongly layered rheology, which created a unique “inverted sombrero” morphology. The CBIS is comprised of a deeper inner crater occurring in crystalline basement within a much wider, shallower outer crater occurring in clastic sediments (Figures 1 and 2).

Crater morphology has long been observed to be a function of crater size. Scaling laws exist which relate the size and shape of the final crater to the transient crater and impact energy (Melosh, 1989). Additionally, target strength and crater size are key parameters in modeling the crater collapse process (Melosh and Ivanov, 1999). Recently, Collins and Wuenemann (2005) successfully modeled the first-order features of the Chesapeake Bay impact by incorporating strongly layered target rheology and acoustic fluidization, a mechanism for temporarily and severely weakening target materials during the impact process. However, their model predicts a much smaller inner crater (~28 km diameter) than the ~32-40 km that has been observed by Poag and co-workers (Poag et al., 2003). Other than recent gravity data (Shah et al., 2005), little data have existed previous to this study to actually constrain the structure of the inner crater. Detailed knowledge of crater structure and distribution of related lithologies are required for more accurate modeling of the impact process, and the role of target rheology.

In 2004, the Virginia Department of Environmental Quality (VDEQ) and the U.S. Geological Survey (USGS) drilled a ~ 0.8 km borehole into the center of the CBIS near Cape Charles, VA. In 2005, the USGS and International Continental Drilling Program (ICDP) have drilled a 1.8 km deep borehole at Eyreville, VA in the deepest part of the CBIS. The goals were to further understand regional groundwater problems caused by the impact structure, the crater’s effect on evolution of the eastern coastal plain, and the mechanics of the impact cratering process.

During fall 2004, a 30-km long wide-angle seismic survey (Figures 1a and 1b) was conducted as part of on-going site characterization in support of the drilling program, as well as to provide constraints on the structure of the deep inner crystalline crater. In this study, the data have been analyzed using travel-time and waveform inversion to produce a high-resolution velocity model of the inner crystalline crater. Travel-time tomography is a robust inversion technique that produces low-resolution velocity models by inverting only the arrival time of seismic energy. Utilizing arriving waveforms can provide more information about seismic velocity than using just travel-times. By inverting for phase and amplitude of seismic energy, full waveform inversion is capable of producing velocity models with approximately wavelength resolution. Low-fold reflection processing has also been used to image the crater floor. The resulting images will provide new, high-resolution constraints on crystalline crater structure. The structure of the crystalline crater may provide insights into the actual size of the Chesapeake Bay impact, as well as the effect of target rheology in the impact process.

Geologic Background

Impact Process

Impact craters form by modification of the target in 3 stages (Melosh, 1989; Melosh and Ivanov, 1999). During the first, the initial energy of the impactor is transferred into the target upon impact. The target is compressed, creating a shock wave that travels radially outward from the impact. The tremendous pressure of the impact may melt or vaporize both the projectile and target. The shock wave and following rarefaction explosively excavate a bowl-shaped transient crater in the second stage. The third stage involves modification of the transient crater. Gravitational instability causes the transient crater to subsequently collapse at its margins, while elastic rebound forms a central peak. The final crater is much wider and shallower than the transient crater. It only takes on the order of several minutes for a crater of the size of the CBIS to form. (Melosh and Ivanov, 1999).

The final crater is classified either as a complex or simple crater based on its size and morphology. Simple craters are circular, bowl-shaped depressions that are the result of small impacts. Larger impacts create complex craters characterized by a central uplift or ring of mountains and/or outer concentric zones of normal faulting. Thus complex craters can further be classified as central peak, peak-ring, and multi-ring craters (Melosh, 1989; Melosh and Ivanov, 1999). On Earth, the transition between simple and central peak crater occurs at 3-5 km diameter, depending on the strength of the target material. There is not yet a well-defined transition from central peak to peak-ring craters on Earth (Melosh, 1989).

Chesapeake Bay Impact

The Chesapeake Bay impact structure was formed ~35 million years ago (Poag and Aubry, 1995), when a bolide struck the continental shelf of the eastern margin of North America (Figure 1). The impact target consisted of ~300m of water (Poag and Norris, 2005), and a 0.5 – 1km thick wedge of unconsolidated Cretaceous – Eocene clastic passive margin sediments (Poag and Ward, 1993) overlying late Proterozoic to Paleozoic Piedmont granitic and metasedimentary basement (Thomas et al. 1989). The water and weak sediments over crystalline rock presented an impact target with strong rheological layering. The effects of rheological layering on the impact process are not well understood.

The impact generated a massive tsunami, which deposited a thick breccia (Exmore breccia) in the ensuing backwash. The Exmore breccia fills the crater and displaces stratigraphic units of the eastern coastal plain. Deposition continued immediately, resulting in a 200-500m thick succession of Eocene-Quaternary clastic marine sediments. Due to the rapid burial and location on a passive margin, the CBIS is free from erosion and tectonic deformation, resulting in a level of preservation uncommon with terrestrial impact structures.

The most obvious effect of the impact is the deposition of impact-related units, such as the Exmore breccia. A tektite strewn field associated with the crater (Poag et al., 2003) has been found in the North Atlantic, Caribbean, and Gulf of Mexico. The impact disrupts regional stratigraphy, including a ~6,400 km² aquifer system in eastern Virginia (Poag, 1997a). A brine has been trapped within the Exmore breccia that replaces the fresh water supply of a large (~2 million) urban population along the southwestern rim of the crater. Despite the apparent size and excellent paleo-stratigraphic dating, it has not been associated with an extinction event.

CBIS Morphology

Seismic data (Figures 1a and 1b) that have been collected by the USGS and oil companies delineate CBIS crater structure (Powars et al., 1993; Poag, 1996; Poag et al., 1999). These data constrain crater structure within the sediments, but due to depth of penetration and sparse reflectivity, provide little information about the inner crystalline crater.

Based on these seismic data, as well as gravity data, the CBIS has been classified as a complex crater. The CBIS has a unique “inverted sombrero” morphology (Figure 2) that is a combination of a broad, shallow outer crater surrounding a deeper inner crater. The 85-km diameter outer crater only occurs in the weaker Cretaceous-Eocene sedimentary units, while the ~32-40 km diameter inner crater is cut into granitic and metasedimentary basement rock. The presence of weak, unconsolidated shelf sediments may exaggerate the horizontal dimensions of the final impact structure. This effect has been observed in other sedimentary marine impact craters (Tsikalas et al., 1998; Stewart and Allen, 2002). The “inverted sombrero” morphology is a combination of a crystalline crater and a broader weak-rheology crater.

The outer rim of the CBIS has been defined by a steep fault scarp in pre-impact sediments. Deformation within the outer crater primarily includes slumping and varying degrees of rotation of large blocks of pre-impact sediment. The distance between the margins of inner and outer crater varies, but averages ~ 21 km. The floor of the sedimentary crater is defined as the interface between undisturbed pre-impact basement and deformed sediments. Crystalline basement has been distinguished regionally by a strong seismic reflection that can be mapped to the edge of the crystalline crater (Hansen, 1978; Dysart et al., 1983). Undisturbed basement on the eastern flank of the crater has an eastward slope of ~0.5 - 1.5° (Powars, 1992).

The boundary between outer sedimentary crater and inner crystalline crater has been defined by the termination of the basement reflector, which implies excavation or deformation of basement. Termination of the basement reflector on different seismic profiles indicates the diameter of the crystalline crater is in the range of 32 – 40 km, and that the crystalline crater is somewhat irregularly shaped (Figure 1). Basement is uplifted at the edge of the crystalline crater.

Although (Poag et al., 1999) tentatively interpreted a ~1km deep reflector as a central peak within the inner crystalline crater, there has been, until recently, little data constraining the structure of the inner crater. A 10-12 km long ovoid central peak has been interpreted from high-resolution gravity data (Shah et al., 2005). Gravity data also indicate walls of the inner crater gently slope towards the center of the crater, but this is in contrast to the model suggested by (Poag, 1999). Magnetic data suggests the presence of 0.4 - 7 km³ of impact melt (Shah, 2005). No other data constrain the depth, the detailed structure of the crystalline crater, or the distribution of lithologies within the crater.

Estimates of impact size and energy from scaling laws, as well as subsequent effects of the impact, depend strongly upon the size of the transient crater. Without detailed knowledge of the structure of the CBIS inner crater, the transient crater size and impact energy are unknown. It is unclear if the boundary between inner and outer crater approximates the dimensions of the transient crater, or if the transient crater was substantially smaller. Three possible scenarios have been proposed (Hole et al., 2003; Kenkmenn, 2003): 1) the edge of the crystalline crater is a peak ring formed by the collapse of a central peak, and the diameter of the transient crater was larger than that of the crystalline crater; 2) the edge of the crystalline crater marks the approximate diameter of the transient crater; and 3) the crystalline crater is the edge of the impact crater, and the diameter of the transient crater was smaller than the crystalline crater.

The first model requires very large impact energies and implies a transient crater diameter of over 40 km diameter. The outer crater rim corresponds to the most distal limit of deformation due to crater collapse. This should excavate a larger crystalline crater than is observed, and is considered an unlikely scenario. The second model supposes a transient crater of 32-40 km diameter, but implies significantly less energy and collapse than the first model. The deformation of the outer sedimentary crater is due to gravitational collapse of the transient crater, and may be exaggerated because of the weak rheology of the sediments. The third scenario suggests the edge of the crystalline crater is really the extent of crater collapse as defined by traditional impact processes, and the outer crater is a sedimentary slump/tsunami structure. This scenario implies a transient crater of ~ 25 km diameter. It predicts an impact with significantly less energy than the first two, and is favored by some numerical models (Collins and Melosh, 2003; Collins and Wunnemann, 2005).

Data Acquisition

The seismic survey was conducted by the USGS (led by Rufus Catchings) during the Fall of 2004 to illuminate the structure of the inner crystalline crater. Seismic data were acquired along a 30 km SW-NE trending transect along the Delmarva Peninsula in Virginia (Figures 1a and 1b). The survey was oriented along a radius of the crystalline crater, and crossed the center and the outer edge of the crystalline crater, and part of the sedimentary crater. The survey was a straight line, and consisted of 29 shots nominally spaced 1 km apart, and 602 receivers nominally spaced 50 m apart deployed in a fixed spread. Gaps existed due to landowner permission at 15-16 km along the line, and receiver failure at the northeast end of the line. Sources were explosive, ranging in size from 5-45 kg.

Figure 3 shows shot gather from the survey. First arrival times were picked for each shot gather (Figure 3 and 4). The slopes of first arrivals indicate a slower layer (~ 2 km/s) over much faster material ($\sim 5-6$ km/s), interpreted as sediment over crystalline basement. Crossover points (defined as the point where refracted waves from a deeper layer overtakes direct waves from a shallow layer) occur earlier in time and at nearer offsets for shots at the southwestern and northeastern end of the line, and occur later and at further offsets for shots in the middle of the line (Figure 4). Basement arrivals at the northeastern end of the line correspond to the outer crater. The later arriving crossover points in the middle of the line are due to deeper basement within the crystalline crater. The earlier crossover points in the southwest are due to elevated basement, most likely a central peak.

Travel-time Tomography

Methodology

Travel-time tomography was used to produce a seismic velocity model of the CBIS. This velocity model was interpreted to constrain the crystalline crater structure, as well as lithology. The model was then employed as the starting model for waveform tomography. The travel-time algorithm of Hole (1992) was used to invert first arrival times of direct or refracted P-waves for 2-dimensional velocity structure. The nonlinear relationship between travel-time and seismic velocity is linearized by only considering small perturbations in the velocity field, and iteratively converging on a solution. An initial grid of velocities is specified. Travel-times are then propagated from a source location to each grid node using a finite-difference approximation to the eikonal equation (Vidale, 1990; Hole and Zelt, 1995). Rays are traced by following the computed travel-time gradient back from receiver locations to the source. The linearization assumes ray paths are independent of velocity for small changes of velocity or slowness, based on Fermat's principle. The travel-time misfit is uniformly distributed over ray paths (backprojection tomography) using:

$$\text{Eq.1} \quad \frac{\delta t}{L} = \delta u$$

where δt is the travel-time misfit, δu is the slowness update applied to grid cells along the ray, and L is the total ray length (Hole 1992). Slowness updates for all rays are averaged at each grid node, weighting by ray length in a volume surrounding a grid node.

A moving average filter is passed over the resultant grid of slowness perturbations, which are then converted to velocity and added to the original velocity model. The resultant velocity model is used as the starting model for the next iteration. The size of the averaging volume and moving average filter control the smoothness of the model. Initial iterations use large smoothing, to allow only large-scale changes in velocity structure. As the large-scale structure is fit, the smoothing size is decreased during subsequent iterations to allow finer structural detail to emerge. The least smooth model without any spurious structure or ray streaking is chosen as the final smoothing.

Travel-Time Inversion Results

First arrivals were picked from shot gathers, and input into the inversion as the observed times. The 2D model volume is 603 x 201 grid nodes, with a 50 m node spacing - for a total model size of 30.1 x 10 km. The inversion was performed with a variety of starting models (Figure 5, left panel) and smoothing schemes to test non-uniqueness and find the most effective trade-off between spatial resolution and artifacts such as ray streaking.

The size of the initial smoothing (grid averaging and moving average filter) was set at a fixed aspect ratio between x direction and z direction, and then reduced by 2/3 every 3 iterations. For example, for a 1:1 aspect ratio, the initial smoothing parameters were 400 nodes (20 km) x 400 nodes (20 km). After 3 iterations, it was reduced to 270 nodes (13.5 km) x 270 nodes (13.5 km), and so forth until a final smoothing of 0.6 km x 0.6 km. This scheme first models large scale structures to fit the travel times, resulting in a velocity model with minimal structure.

Aspect ratios from 12:1 to 1:1 were tested, but the preferred choice of smoothing aspect ratio is 4:1. Vertically oscillatory velocity structure begins to appear above the 5:1 smoothing aspect ratio for most starting models, while the 3:1 ratio oversmooths the strong variation in velocity with depth. The initial smoothing parameters for the 4:1 smoothing ratio were 400 nodes (20 km) x 100 nodes (5 km), and the final smoothing ratio was 12 nodes (600m) x 3 nodes (150m). Final spatial resolution is limited by the coarse 1 km shot spacing.

Several starting models (Figure 5) were tested to observe possible dependence on the initial model. Some models had linear increases in velocity with depth, but 2 starting models contained changes in gradient with depth, to simulate the expected velocity model. The inversion of each starting model converged to the same solution to a depth of ~5 km (the limit of ray coverage). The preferred final model (Figure 6a) used a starting model with a constant, steep linear velocity gradient and 27 iterations, a smoothing ratio of 4:1, and a final smoothing of 0.6 km x 0.15 km.

The final velocity model (Figure 6a) exhibits very slow (~ 2 km/s) surface velocities that correspond to sediment, a strong increase in velocity (> 4 km/s) to crystalline basement, then a low velocity gradient with depth. Ray coverage in the model (Figure 7) indicates that the model is well constrained down to a depth of ~3km, beyond the depth of basement. There are many turning rays at the shallowest and deepest depths of ray coverage, but few at intermediate depths in the deeper sedimentary section. Rays at intermediate depths are mostly vertical, and so vertical velocity structure is smoothed.

Interpretation

Figure 6a shows the preferred velocity model from travel-time inversion. Slower velocities ($< 2\text{--}3$ km/s) are interpreted as the pre-impact and post-impact unconsolidated sediments. Basement has velocities of over 4 km/s, but exhibits strong lateral variations. The interface between basement and overlying sediments is interpreted as the maximum vertical velocity gradient (Figure 8a). On the northeastern end of the line, outside the crystalline crater, there is a sharp change in velocity at ~ 1.5 km in depth that is interpreted as the sediment/basement interface. Basement velocity of $\sim 5 - 5.5$ km/s is consistent with earlier estimates of basement interval velocity from seismic surveys in southeastern Virginia and Maryland (Klitgord and Schneider, 1994).

At 19 km distance along the line the interface slopes up to 1.25 km depth, and is interpreted to be the raised margin of the crystalline crater. The basement then gradually slopes down to ~ 1.8 km depth, between 7 km and 19 km interpreted as the base of the crystalline crater. Suggestions of finer structure along the crater floor are evident, but are beyond the resolution of the model.

The basement gradually slopes up to about 1km depth near the southwest end of the line - interpreted as the central peak of the crater. The basement contact here is much more gradational than basement beneath the sedimentary crater at the northeast end of the line. Basement

velocities of the inner crystalline crater are consistently lower than beneath the outer sedimentary crater. One possible reason is fracturing of basement within the inner crater.

Waveform Tomography

Methodology

To obtain better constraints on crater structure, particularly depth to crater floor, waveform tomography was used to create a higher-resolution velocity model of the CBIS. The waveform inversion method used is a finite-difference, frequency domain, local-descent method described by Pratt (1999). The wave equation is formulated in the frequency domain, and inversion is performed for small sets of discrete frequencies.

The waveform inversion problem is highly non-linear, mainly due to cycle-skip issues, where the velocity model is changed to shift peaks and troughs of synthetic seismic data to match the wrong peaks and troughs of the observed data. Because of this, the method requires a starting model that produces synthetic data matching the observed data to within half a wavelength. This is necessary to encourage convergence on a global rather than a local minimum (Pratt, 1999). Fortunately, travel-time tomography produces a sufficiently accurate starting model. In this usage, waveform inversion is an extension of travel-time inversion to improve the resolution of a velocity model.

One major advantage of a frequency-domain approach is the ability to first invert for lowest data frequencies (i.e. longest wavelengths), since the smooth travel-time tomography velocity model may not match the observed higher frequencies to within half a wavelength. This is especially useful in mitigating any starting model dependence and cycle-skip issues. Therefore, it is necessary to have strong low frequency signal for waveform inversion. Once the longest wavelength structure has been satisfactorily modeled, the inversion can proceed to higher and higher frequencies to produce a more detailed model (Pratt, 1999).

The ability to invert for small groups of discrete frequencies simultaneously is computationally advantageous since the model must only be sampled densely enough for the shortest wavelengths used in the inversion. There are also other computational advantages to a frequency-domain approach. Much of the computational expense is attributed to the forward modeling calculation, which in the frequency domain, is alleviated by choice of a finite difference operator that decreases the required number of grid nodes in a model for a given wavelength, ordering scheme for the elements of the finite difference matrix, and choice of matrix solver (Pratt, 1999).

The algorithm uses a nine-point finite difference operator described by Jo et al. (1996), which is a linear combination of the more traditional five-point operator and a rotated version of the five-point operator. Using this finite-difference operator, the number of grid nodes required per wavelength for 95% or better accuracy reduces from 10 grid nodes using the five-point operator down to 4 (Jo et al, 1996; Stekl, 1998). The elements of the finite-difference matrix are ordered using the nested dissection scheme (Liu and George, 1981), a technique for efficiently arranging the elements of a sparse matrix for LU decomposition. The system of wave equations is then solved using LU decomposition (Pratt, 1999). LU decomposition is efficient when forward modeling is required for multiple sources. This is especially critical, as waveform

inversion requires many forward models – one per forward and backpropagated wavefield per source per frequency per iteration, and one per update vector for each iteration (Pratt, 1999).

The forward modeling is accomplished using a visco-acoustic wave equation. The forward modeling algorithm solves for the Fourier-transformed pressure wavefield, $u(\mathbf{x}, \omega)$, at each grid node in the model using the frequency-domain wave equation (Pratt, 1999):

$$\text{Eq. 2a} \quad \nabla^2 u(\mathbf{x}, \omega) + \frac{\omega^2}{c^2(\mathbf{x})} u(\mathbf{x}, \omega) = -f(\mathbf{x}, \omega)$$

Here $u(\mathbf{x}, \omega)$ is the total wavefield, ω is frequency, $c(\mathbf{x})$ is velocity, ∇ is the 2D Laplacian, $f(\mathbf{x}, \omega)$ is a source term, and $\mathbf{x}$ is the 2D position vector. In discretized matrix form it is:

$$\text{Eq. 2b} \quad S(\omega)u(\omega) = f(\omega)$$

where $f(\omega)$ is a source term, and $S(\omega)$ is the finite difference matrix, incorporating the nine-point finite difference operator and the velocity term (following the notation of Pratt 1999). The calculated Fourier-transformed wavefield is compared with the observed wavefield to form the complex data residual vector δd . The problem is then to minimize the L2-norm misfit function:

$$\text{Eq. 3} \quad E(m) = \frac{1}{2} \delta d^T \delta d^*$$

where T denotes the transpose, $*$ denotes the complex conjugate, and m is the vector of model parameters (slownesses in the case of this survey). Model parameters are then updated according to a vector defined by the gradient of the misfit function (with respect to model parameters), and a step length η :

$$\text{Eq.4} \quad m^{(k+1)} = m^{(k)} - \eta \nabla_m E^{(k)}$$

where $\nabla_m E^{(k)}$ is the gradient of the misfit function with respect to each model parameter, and k is the iteration number. The gradient defines a search direction that is scaled by a steplength. It has been shown that the gradient in parameter space can be found without direct computation of the Frechet derivative matrix, which would require a single forward modeling step per model parameter (Pratt, 1999). Instead, the gradient can be computed as the product of the forward propagated and back propagated wavefields, similar to the numerical solution to the undiscretized wave equation (Slaney, 1984):

$$\text{Eq.5} \quad \nabla_m E = \text{Re}\{J^T \delta d^*\} = \text{Re}\{F^T v\}$$

where $\text{Re}\{\}$ indicates the real part, J is the Frechet derivative matrix, F is a column vector of “virtual sources” and v is the back propagated wavefield. Each virtual source is calculated as $f = \frac{\partial S}{\partial m} u$. These simulate the actual diffraction that occurs as a forward propagated wavefield interacts with a scatterer in the model. Data residuals are back propagated into the model

according to $v = S^{-1} \delta d$. Equation 5 resembles the Born approximation of the wave equation, through which a “wave path” is defined (Woodward, 1992). The wave path, in a more practical sense, defines an elliptical region in which velocity updates are distributed. The similarity between this imaging technique and prestack depth migration has been noted and discussed in other papers (Woodward, 1992; Tarantola, 1984),

Convergence of the gradient method is slow, but can be drastically improved. By adopting the conjugate gradient method for the inversion, each search direction calculated is orthogonal to the previous search directions. This drastically reduces the number of iterations, as the gradient method may search in the same direction repeatedly, rather than just computing one update per direction with a larger step length.

Pre-Processing

Before waveform inversion can proceed, special preparation of the data is required to eliminate energy and amplitude affects the algorithm does not model (Pratt, 1999). The computation uses a 2D visco-acoustic wave equation, and so real 3D elastic waves and mode conversions are not accounted for. It is also desirable to initially force the inversion to predict the first arrivals, since forward scattering contains the low and intermediate wavenumber (long-wavelength) information about the velocity model. After modeling the first arrivals and longer wavelength structure, later arriving diffracted or back-scattered energy may be included to add higher wavenumber information to the model.

The data were first muted at the near offsets, to remove high-amplitude surface waves and time windowed around the first arrivals (Figures 9a and 9b). Windowing is necessary to exclude undesirable energy such as surface, shear or multiply scattered waves from the inversion. The time window begins 50 ms before picked first arrivals, and is 750 ms long. It is computed to contain at least two cycles of first arrival energy in the data at 3 Hz, the lowest frequency with adequate signal to noise ratio. The windowed data also contain wide-angle reflections from the crystalline basement.

Synthetic data were calculated in the travel-time tomography velocity model using the algorithm of Pratt (1999) (Figures 9c and 9d). RMS amplitudes were calculated for all shots using offset bins of 1 km from the synthetic data and the muted, windowed, observed data. These amplitudes were compared (Figure 10) to compute an empirical correction to scale the observed data to the synthetics. The empirical correction (C) is calculated as a ratio of synthetic RMS amplitudes, $A(d)$, and observed RMS amplitudes $A(d)'$ as a function of source-receiver offset:

$$\text{Eq. 6} \quad C = \frac{e^{A(d)}}{e^{A(d)'}}$$

The scale of the amplitude corrections depends on the amplitudes of sources used in forward modeling. Global scaling is not crucial, as it will be incorporated into source inversions; only the amplitude vs. offset trend is considered. The short-wavelength amplitude corrections need only be approximate, as the results are, at least in the case of this survey, only moderately sensitive to them.

Source Estimate

The first step of the inversion process is to estimate seismic sources. It is necessary to account for the source wavelet, since the Born approximation relies on solving for the Green's

function of the medium. The wavelet is estimated using the method detailed in (Pratt, 1999). An arbitrary source is modified by a complex scalar, s , which can explicitly be solved for:

$$\text{Eq. 7} \quad S(\omega)u(\omega) = s f(\omega)$$

The source problem is vastly overdetermined due to the large number of receivers, and converges in a single iteration. The 29 sources were inverted for separately (Figure 11). Phase is similar for all 29 sources, with small differences possibly due to source depth. Significant amplitude variation is observed because the sources were of variable size.

Inversion Strategy

A range of frequencies for inversion was selected upon spectral analysis and inspection of bandpass filtered time-domain data. 3Hz is the lowest frequency in the data with adequate signal to noise ratio. The data contain signal up to 35Hz, but the highest frequency used was limited by computational considerations.

Sirgue and Pratt (2003) describe a strategy for frequency selection for waveform inversion, using different source-receiver offsets (i.e. different incident angles) to extend the wavenumber coverage of a single frequency in the data. This is derived using reflection geometries, and no such strategy has been demonstrated for refracted and forward scattered waves. However, the strategy has been successfully implemented in waveform inversions of first arrivals in densely sampled reflection surveys (Sirgue and Pratt, 2003; Hole et al., 2005). In the case of this survey it did not yield very stable results. Forward scattered waves for this survey may not sample the same point in the model at as large a range of incident angles, and so there may not be as broad a range of wavenumber covered by a single frequency. Instead, groups of frequencies with smaller spectral spacing were used. The data were windowed in a 0.75 s window around the first arrivals, so the required spectral spacing to avoid aliasing is $\frac{1}{0.75s} = 1.33\text{Hz}$. Smaller frequency spacing is used for some data redundancy in the inversion.

The range of frequencies to be inverted was selected based on the highest frequency in synthetic data from the current velocity model that matched the observed data to within half a wavelength. Ranges for inversion began with the lowest frequencies and proceeded to higher frequencies (Table 1). Five iterations of waveform inversion were performed for each group. Each range of frequencies overlapped the previous range to ensure model updates at higher frequencies do not violate those at lower frequencies. Similar results were obtained in tests using different inversion bandwidths. For example, similar velocity models were produced by inverting up to 10Hz using groups of 3-6Hz, 5-8Hz, and 7-10Hz as for inverting up to 10Hz using just the groups 3-6Hz and 5-10Hz. Sources were estimated before each group, and updated during each iteration.

Inversion results are very sensitive to residual offset weighting and wavenumber filtering of the gradient. Only source-receiver offsets from 1 -10 km were used. Offsets beyond ~10 km introduced artifacts and spurious structure to the velocity models. This is presumably due to decreasing signal to noise ratio, even though the raw data indicate good signal at all offsets. The wavenumber filter limits spatial wavelengths of model perturbations because the gradient may oscillate at the wavelength scale or introduce perturbations of smaller wavelength than is resolvable. The longest wavelength at 16 Hz in the final velocity model is ~0.4 km in the

basement; wavelengths in the sediment are up to one-third smaller. The final wavelength filter was 0.460 km in the horizontal direction by 0.230 km in the vertical direction.

The final spatial resolution of the velocity model may be limited by the large shot spacing. The denser receiver spacing partially compensates for undersampling in the shot domain, but more shots are needed to cancel out wavelength oscillations in the velocity model. These oscillations are subtly present in the final velocity model, and look similar to migration “smile” artifacts and/or ray streaking. The streaking artifacts are especially evident as edge effects at the bottom corners of the model.

Starting Model

The result of travel-time inversion (Figure 6a) was used as the starting model for full waveform inversion. Absorbing boundary conditions of the waveform algorithm required an additional 10 grid node buffer on each side of the model. Starting velocities range from ~1.5 km/s to ~6.3 km/s, and were extrapolated to 1 km/s in the buffer zone at the top of the model. A low velocity zone was built in the bottom 0.5 km of the model to reduce a wide-angle reflection from the base of the model.

For frequencies less than 7 Hz, velocity models only require 50 m grid spacing (621 x 121 nodes). 25 m spacing is required to accurately forward model frequencies up to 12 Hz (1242 x 242 nodes), and 20 m spacing is required to accurately model frequencies up to 16 Hz (1550 x 1300 nodes) (Table 1). Higher frequencies were not inverted due to current memory limitations on our computer. However, it should be possible to invert higher frequencies in these data to obtain higher spatial resolution.

Results

Figure 6b is the result of inverting for waveforms from 3 to 16 Hz. Despite some minor artifacts, the velocity model produced using waveform inversion shows a strong improvement in resolution and detail compared to the travel-time velocity model. Figure 12 shows the difference between waveform and travel-time inversion results. Figure 12 shows only short and medium-wavelength changes, but no long-wavelength changes, as the travel-time tomography accurately models the long wavelengths, but cannot resolve the shorter wavelengths. The Figure 13 shows 1D velocity profiles at different locations along the line. The travel-time tomography velocity profiles are very smooth with only long-wavelength velocity structure. The waveform tomography velocity profiles show similar long-wavelength structure, but add smaller-scale structure.

Near-surface velocities exhibit little deviation from the travel-time results. Velocities in the deeper sediment section were decreased by ~0.5 km/s in the inner crater, and by ~0.2 km/s in the outer crater (Figure 12). These velocity changes correct for the lower sediment velocity at intermediate depths that were previously unconstrained by the smooth travel-time tomography. Velocity at the top of basement was increased by ~1 km/s at places along the inner crater, and by ~0.5 km/s in the outer crater (Figure 12). The increase in basement velocities and decrease in overlying sediment velocities sharpens the sediment – basement boundary from that of the smooth travel time tomography velocity model. Velocities below ~2.5 km depth show little change from the travel-time tomography result, and show small-amplitude artifacts. The sharpening of the basement interface and the small anomalies in the poorly constrained portion of the model at depth both suggest that the technique is not producing substantial artifacts.

Synthetic data calculated in the final velocity model (Figure 9e-f) more closely resemble the observed data (Figure 9a-b). The shape of the arriving wavelet of the waveform synthetic data is similar to that of the observed arrivals. The synthetic waveform data also show the reflection from basement at near offsets that is evident in the observed data.

Interpretation

Comparison to lithologies observed in the borehole drilled at 9 km along the line illustrate the improvement in resolution as compared to travel-time tomography, as well as the overall success of the waveform inversion technique (Figure 13). The travel-time tomography result shows an increase in velocity with depth that is very smooth across the drilled interfaces between sediment and granite and between deeper suevite and lithic breccia and schist basement. The waveform tomography result not only resolves the top of the granitic megablock encountered during drilling, it also sharpens the contact between breccia and underlying basement.

Comparison of lithologies and the waveform inversion velocity model show post-impact sediments have velocities below 2 km/s, and Exmore breccia has velocities between 2 and 3 km/s. The 2 km/s contour coincides with a subtle vertical gradient in the waveform velocity model at shallow depths (Figures 12). The base of post-impact sedimentary strata is interpreted to follow this contour, which occurs at 0.4 – 0.5 km in depth over most of the image but is uplifted to 0.2 km depth over the central peak. This is consistent with previous seismic and borehole data (Poag, 1997). The Exmore impact breccia underlies the post-impact sediments, and ranges in thickness from 500 m over the central peak to ~1300 m thick at the deepest parts of the inner crater.

The contact between schist basement and breccia infill may be interpreted as occurring at the largest vertical velocity gradient, although the Eyreville well indicates that the largest vertical velocity gradient locally can occur at the top of crystalline megablocks. The vertical gradient of the waveform inversion result (Figure 8b) clearly shows sharper features than the gradient of the travel-time result (Figure 8a). The maximum vertical gradient at the northeast end of the line corresponds to the 5.5 km/s contour at ~1.5 km depth. The steepest vertical gradient beyond 20 km along the line is interpreted as undeformed basement of the outer crater. It slopes down to the east, in agreement with regional basement trends, but is slightly deeper than previous estimates of ~1.3 km extrapolated from distant regional boreholes (Powars, et al., 1992).

A basement uplift at ~ 19 km along the line corresponds to the edge of the inner crater. A similar sized ridge of basement surrounding the inner crater has also been interpreted in previous reflection data for the western side of the crater (Poag 1997; Poag et al., 1999). Basement uplifts interpreted in previous data average ~9 km in width, with ~0.12 km of relief. The uplift interpreted in this survey is ~ 2 km wide, with ~ 0.2 km of relief. There is a block of high velocities adjacent to the uplift that may be a large crystalline megablock detached from the edge of the crystalline crater. In the lower resolution travel-time tomography velocity model, this megablock is indistinguishable from the uplift at the edge of the crystalline crater, and so appears as a single, broader uplift. The broad uplift observed in previous reflection data is typically faulted and may indicate similar megablocks.

The wall of the crystalline crater gently slopes down to the crater floor, in agreement with interpretations of a gradual slope from gravity data (Shah et al., 2005). The maximum vertical velocity gradient in the crystalline crater roughly corresponds to the ~4.5 km/s contour, and so the basement interface is interpreted along the 4.5 km/s contour. Basement occurs at depths of

1.5 – 1.8 km, and there is ~ 0.4 km rise in basement velocity at 13 – 15 km along the line. The crater floor is up to 0.3 km deeper than the undisturbed basement of the outer crater, but the true relief is in the range of 0.4 – 0.7 km when the regional slope of the basement (Powars, 1992) is accounted for. The interpreted depth to basement is in good agreement with the deep borehole. Basement velocity as low as 4.5 km/s is very slow for granite and schist, likely indicating a high degree of fracturing.

At the central peak at the southwest end of the line the maximum vertical velocity gradient lies between the 3.5 km/s and 4 km/s contours at ~ 0.8 km depth. Crystalline breccia was observed at the same ~0.8 km depth in the nearby Cape Charles borehole. The central peak interpreted here is a broad irregular uplift ~10 km in diameter with gently sloping sides, consistent with gravity results (Shah et al, 2005). There is a small local rise in basement velocity on the northeastern flank of the uplift, which may be slumped material. Basement velocity less than 4 km/s is also very slow for granite and schist. There is a gradational increase with velocity in depth at the central uplift. These observations suggest a highly fractured and brecciated basement, consistent with observation in the Cape Charles borehole.

Reflection Imaging

Low-fold seismic reflection processing using a simple brute-stack approach (Figure 14) was used to image reflectors (Figure 15). The binned CDP spacing is 50 m, giving a maximum fold of 30. The best stacking velocities were 1.1 km/s down to 0.5 s and ~2.3 km/s for later arriving energy.

A continuous, sub-horizontal reflection is observed arriving at ~0.5 s. This reflector has been mapped in other seismic studies of the shallow coastal plane (Poag et al., 1993). Based on earlier seismic and borehole data, it was interpreted as a thin pelagic layer overlying the crater fill breccia. It is relatively flat lying, with a slight sag over the inner crater.

A more prominent reflection occurs at ~1.3 seconds at the northeastern end of the line. Regional seismic and well data indicate that the reflector corresponds with the top of pre-impact crystalline basement outside the CBIS and beneath the outer sedimentary crater (Poag et al., 1993; Hansen, 1978; Dysart et al., 1983). The reflector has a small dip to the northeast, consistent with the regional trend of basement. The reflector rises in depth at about 19 km along the line, and is interpreted as the basement uplift marking the boundary of the inner crater, similar to Poag (1999). This corresponds in location to the basement uplift interpreted from the waveform results.

Consistent with previous reflection data, the basement reflection in the crystalline crater is much weaker and deeper. At 8-10 km along the line, there is a reflector at ~ 1.5 s. This reflector corresponds to the top of some crystalline megablocks interpreted in the velocity model and observed in the Eyreville borehole. These megablocks overly suevite and lithic breccia, and do not represent the true floor of the crater.

The reflector rises sharply to ~ 1 s at the southwestern end of the line. The steep rise at the southwest end of the line is interpreted as the central uplift. The uplift as interpreted from reflection data has a diameter of ~8-10 km, which is in excellent agreement with the final velocity model.

Implications for the CBIS

The seismic reflection section and velocity model provide new, higher-resolution constraints on the structure of the crystalline crater of the CBIS. An updated model of the CBIS is proposed in Figure 16. The model contains many of the features described in previous work (Poag, 1997; Shah et al., 2005). There is flat, east-dipping basement in the outer sedimentary crater, a basement uplift defining the boundary between inner crystalline and outer sedimentary crater, a deeper inner crater entirely within crystalline basement, and uplift at the center of the crystalline crater.

The velocity of undisturbed basement beneath the sedimentary crater is ~ 5.5 km/s. Basement velocity is much slower in the inner crater, as slow as 4.5 km/s. The crater uplift has basement velocities as low as 3.5 km/s. Tectonostratigraphic mapping of basement shows the CBIS occurred in the Chesapeake Bay block (Thomas et al., 1989), which is a large accreted terrain of metamorphosed sedimentary rocks. Large granitic plutons are interspersed throughout the Chesapeake block, but it is predominately schist. Schist and granite are both observed in the Eyreville and Cape Charles boreholes. Some of the difference in velocities between undisturbed basement and basement in the inner crater and central uplift may be due to variations between granite and schist. However, the observed basement velocities in the inner crystalline crater are slow for both rock types. The most plausible explanation for the velocity contrast between basement of the inner crater and that of the outer crater is impact-induced fracturing. Fracturing of the inner crater is inferred to occur as deeply as the model is constrained.

The fracture porosity required to account for differences in seismic velocity may be estimated using Wyllie's time average equation (Wyllie et al., 1956):

$$\text{Eq. 8} \quad \frac{1}{V} = \frac{\phi}{V_{pfl}} + \frac{1-\phi}{V_{po}}$$

where ϕ denotes porosity, V is the velocity of fluid saturated rock, V_{pfl} is the velocity of saturating fluid, and V_{po} is the velocity of the unfractured rock. Using $V = 5$ km/s, $V_{pfl} = 1.5$ km/s (for water), $V_{po} = 5.5$ km/s, fracture porosity is estimated to be $\sim 5\%$. For $V = 4.5$ km/s, $V_{pfl} = 1.5$ km/s (for water), $V_{po} = 5.5$ km/s, fracture porosity is $\sim 10\%$. Using $V = 3.5$ km/s as an estimate of basement velocity of the central uplift, fracture porosity of the central uplift is estimated to be as high as $\sim 20\%$. Such high fracture porosity indicates that basement at the central uplift is highly brecciated, which is consistent with the Cape Charles borehole.

The inner crater reaches a maximum depth of ~ 1.8 km, and contains slump blocks on its floor. The Exmore breccia filling the inner crater contains megablocks up to 2 kilometers wide and 100s of meters deep, one of the largest of which (0.3 km in depth) was unexpectedly drilled through in the Eyreville borehole. The furthest radius of the crystalline crater is defined by the basement uplift at a depth of ~ 1.2 km at ~ 19 km along the line. The travel-time tomography and reflection data show a broader uplift, but the waveform tomography apparently resolves a separate block at ~ 18 km that is higher than the more moderate uplift at ~ 19 km. The diameter of the crystalline crater is then ~ 30 - 32 km.

A local uplift in basement is observed on the crater floor in Figure 6b at ~13-14 km along the line (~11 km from the center of the central uplift). Basement velocity inside this radius is ~ 0.5 km/s slower than outside (Figure 6b). The lateral change in velocity may indicate a change in basement fracture volume that represents the boundary of the transient crater. The uplift at the edge of the crystalline crater then is at the edge of the impact crater.

Empirical scaling laws exist based on observations of lunar and terrestrial craters, small-scale experiments, and numerical modeling that relate various crater features. These scaling relationships are often used in the literature to predict features of craters that may have been heavily eroded or tectonically deformed. Therriault et al. (1997) describe one such scaling relationship that relates diameter of central peak D_{cp} to final crater diameter D :

$$\text{Eq. 9} \quad D_{cp} = 0.31D^{1.02}$$

The observed 8-10 km diameter central peak and 30 km wide crystalline crater fit this scaling relationship.

The diameter of the transient crater D_{tc} may be estimated from the diameter of the final crater D using another scaling law (Croft, 1985):

$$\text{Eq. 10} \quad D_{tc} = 1.23D^{0.85}$$

If the extent of the crystalline crater represents the size of the transient crater, then $D_{tc} = 30$, which predicts $D = 42$ km. This would require the weak, saturated sediments to exaggerate the crater's final dimensions to twice the predicted dimensions. Alternatively, interpreting the features at ~ 13 km along the line as indicative of the transient crater gives $D_{tc} = 22$ km, which predicts a final crater diameter of $D = 30$ km. This final crater diameter matches the observed diameter of the crystalline crater. This interpretation is preferred due to the internal consistency of diameters of the observed central peak, interpreted transient crater, and observed crystalline crater.

Grieve and Pesonen (1992) suggested different scaling laws between crater depth d and diameter D for impacts into crystalline targets and weaker sedimentary targets:

$$\text{Eq. 11a} \quad d = 0.12D^{0.30} \quad \text{for a sedimentary target}$$

and

$$\text{Eq. 11b} \quad d = 0.15D^{0.43} \quad \text{for a crystalline target}$$

Using $D = 85$ km for the sedimentary crater, the scaling laws predict a depth of ~0.45 km, which is a low estimate of the observed relief from the ~0.5 km deep top of pre-impact sediments to the 1.5-1.8 km deep top of basement in the inner crystalline crater. Using the diameter of the crystalline crater as the outer limit of the crater, the predicted depth is ~ 0.65 km. This is consistent with a crystalline crater depths corrected for a regional basement slope of 1.15°. This is again consistent with the crystalline crater coinciding with the full extent of the collapsed impact crater.

In this model, deformation of sediments to create the outer crater is a very weak rheology collapse associated with the existence of the crystalline crater, rather than a product of transient crater growth and collapse. The behavior of weak, unconsolidated sediments in response to the

void created by excavation of target material is similar to turbidites. The slope of pre-impact basement, though slight, may be near or beyond the angle of repose for such weak sediments. An impact-generated tsunami or water flowing into the void created by excavation may also be at least partially responsible for deformation of weak sediments.

Impact energy may be estimated from the transient crater dimensions using a variety of scaling laws. Assuming a vertical impact, impactor density of $\sim 3000 \text{ kg/m}^3$, velocity of 17 km/sec, the Pi-scaling laws described by Melosh (1989) predict an impactor $\sim 2.3 \text{ km}$ in diameter, and an impact energy of $\sim 6.7 \times 10^5$ Megatons. This estimate is somewhat smaller, but not inconsistent with numerical models (Collins and Wunnemann, 2005) that predict an impact a $\sim 3.2 \text{ km}$ diameter impactor and impact energy of $\sim 1.75 \times 10^6$ Megatons.

It is possible that the water and sediments absorbed some of the energy of the impact, resulting in a smaller transient crater than would have formed if the bolide struck only crystalline rock. However, once the excavation crater was created, collapse of the transient crater to create the final crystalline crater proceeded as if the sediments were absent. If the sediments had played a role in the crystalline crater collapse, the crystalline crater diameter would not match scaling laws. The rheology of unconsolidated, water-saturated, clastic sediments is relatively well known. Crystalline crater collapse occurs by viscous fluid-like deformation with a rheology much weaker than normal rocks, weakened by impact shock processes such as acoustic fluidization or thermal softening (Melosh, 1999; O'Keefe and Ahrens, 1993). The observations at CBIS indicate that the shock-weakened crystalline crater collapse rheology may be much weaker than normal rock, but is still much stronger than unconsolidated, water-saturated clastic sediments. This provides a bound on mechanism models for crater collapse.

Conclusions

The Chesapeake Bay impact structure represents a unique opportunity to examine the effects of target rheology on crater impact processes. Waveform tomography from the 2004 seismic survey successfully produced a high-resolution velocity model of the structure of the inner crater that shows significantly improved resolution from previous knowledge or from using travel-time tomography. The final velocity model approaches wavelength resolution in basement at the maximum frequency modeled.

The resultant velocity model and low-fold reflection image are interpreted in order to construct a model of the inner crater (Figure 16), the missing puzzle piece in understanding the CBIS. The features interpreted are consistent with both recent gravity observations (Shah et al., 2005) and numerical modeling (Collins and Wunnemann, 2005; Collins and Melosh, 2003). The crystalline crater has a diameter of $\sim 30 \text{ km}$. The outer margins of the crystalline crater are uplifted in one or more separate fault blocks. The floor of the inner crater ranges from 1.5 – 1.8 km in depth. The crater floor is only 0.1 – 0.4 km deeper than adjacent pre-impact basement, but correction for the regional slope of the basement provides a greater depth consistent with crater scaling laws. The crystalline crater has a $\sim 10 \text{ km}$ diameter central uplift with nearly 1 km relief above the deepest part of the crystalline crater.

The transient crater is interpreted to have a diameter of $\sim 22 \text{ km}$, corresponding with an observed basement uplift and a lateral change in basement velocity within the crystalline crater. The crystalline crater, central peak and transient crater dimensions observed are self-consistent with published scaling laws. This suggests that the impact collapse process was controlled by the

crystalline basement rock and that the weaker overlying sediments collapsed to fill the excavated hole. The crystalline basement rock would have had to be substantially stronger than unconsolidated, water-saturated sediments, even though it was fluidized by some mechanism (i.e. acoustic fluidization or thermal softening).

An impact energy of $\sim 6.75 \times 10^5$ Megatons is predicted by scaling laws for a crystalline crater of the observed dimensions. This smaller impact energy may explain the lack of a major extinction event associated with the impact.

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Tables

Frequency Range	Frequency Spacing	Number of Frequencies	Model Grid Spacing (km)	Wavelength Filter (km)	Residual Reduction	Computation Time (relative to first group)
3-6Hz	0.3333 Hz	10	0.5	1.07 x 0.535	29.50%	1
5-8Hz	0.5 Hz	7	0.25	0.791 x 0.395	21.23%	1.25
7-10Hz	0.5 Hz	7	0.25	0.700 x 0.350	17.80%	1.25
10-16Hz	1.0 Hz	7	0.2	0.460 x 0.230	31.41%	4

Table 1. Waveform inversion sequential strategy.

Figures

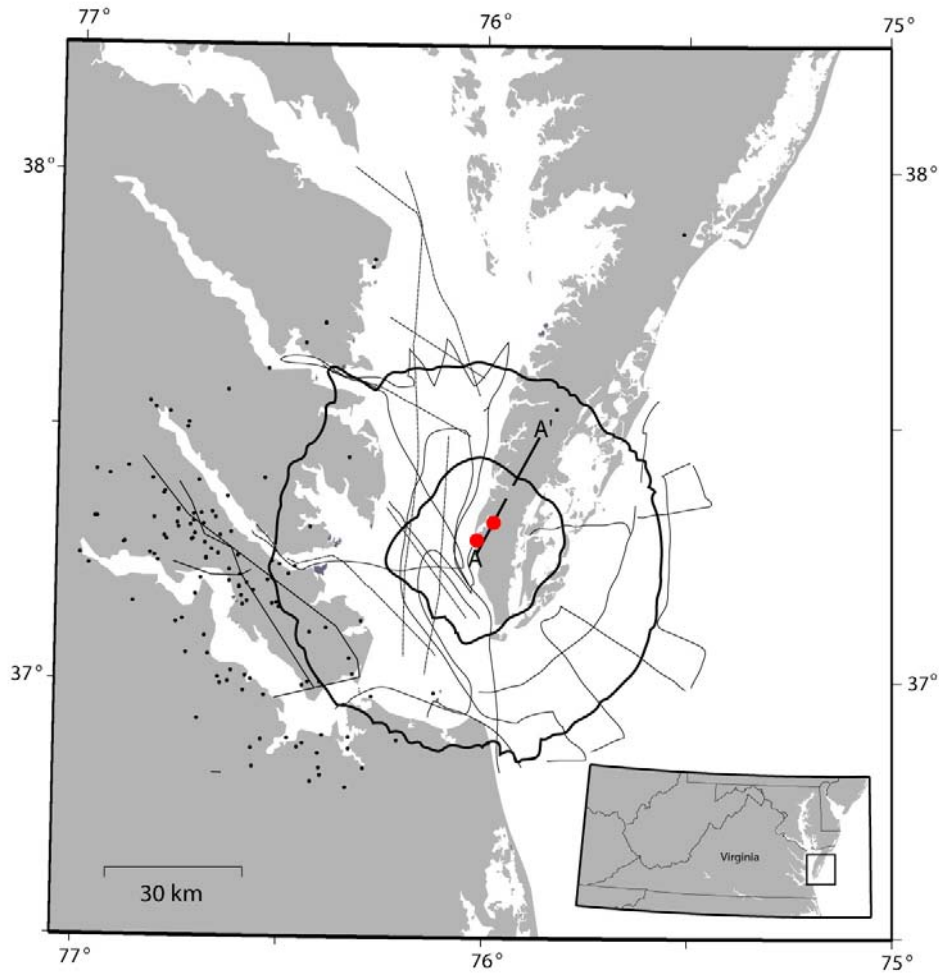


Figure 1a. Map of Chesapeake Bay impact structure on the eastern margin of Virginia. The large circle represents the outer crater as interpreted from previous seismic reflection data, and the smaller circle represents the inner crater. Solid lines are previous seismic surveys, and dots represent borehole locations. The thick SW – NE line is the 2004 seismic survey, the red dots are the Cape Charles borehole and Eyreville deep boreholes. The Eyreville borehole is the only borehole to penetrate basement in the inner crater.

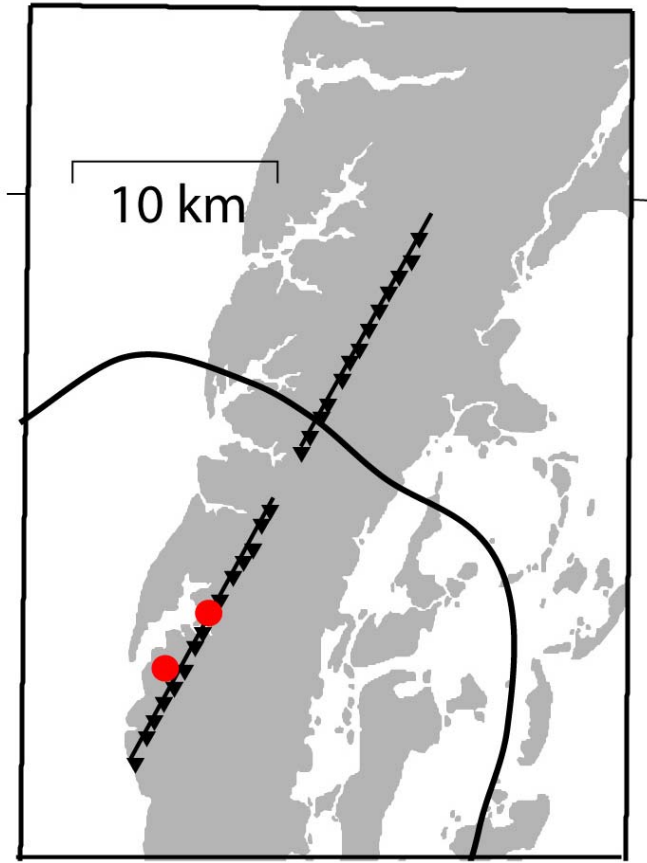


Figure 1b. Survey geometry. Black triangles represent shot points, the red circles are the Eyreville and Cape Charles boreholes. The gap in the line at 15 – 17 km is due to lack of landowner permission. The solid line is the extrapolated extent of the crystalline crater from previous seismic reflection data.

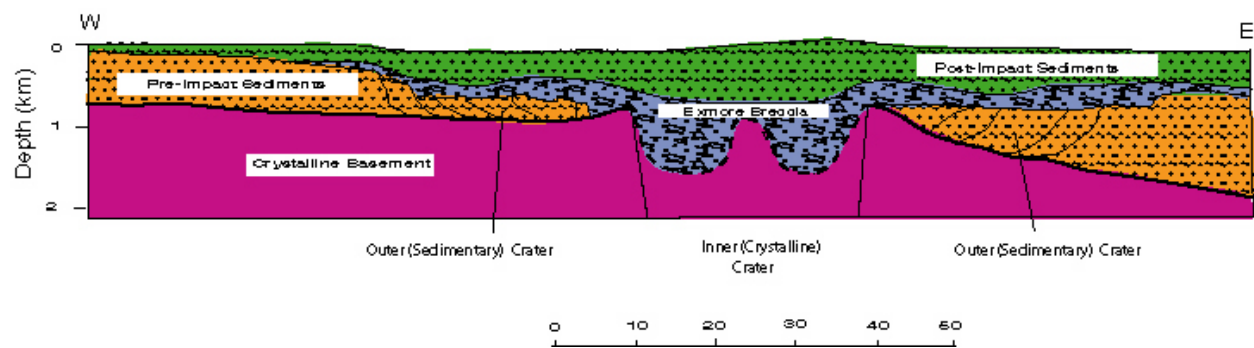


Figure 2. Schematic west-east cross-section of CBIS (after Poag et al., 1999). Outer crater has been interpreted from previous seismic reflection data (Poag, 1997). Inner crater (dashed line) has little data constraining it. The survey line is SW – NE, and so the observed basement slope will not be as pronounced as in this west-east cross-section. Vertical exaggeration is 10:1.

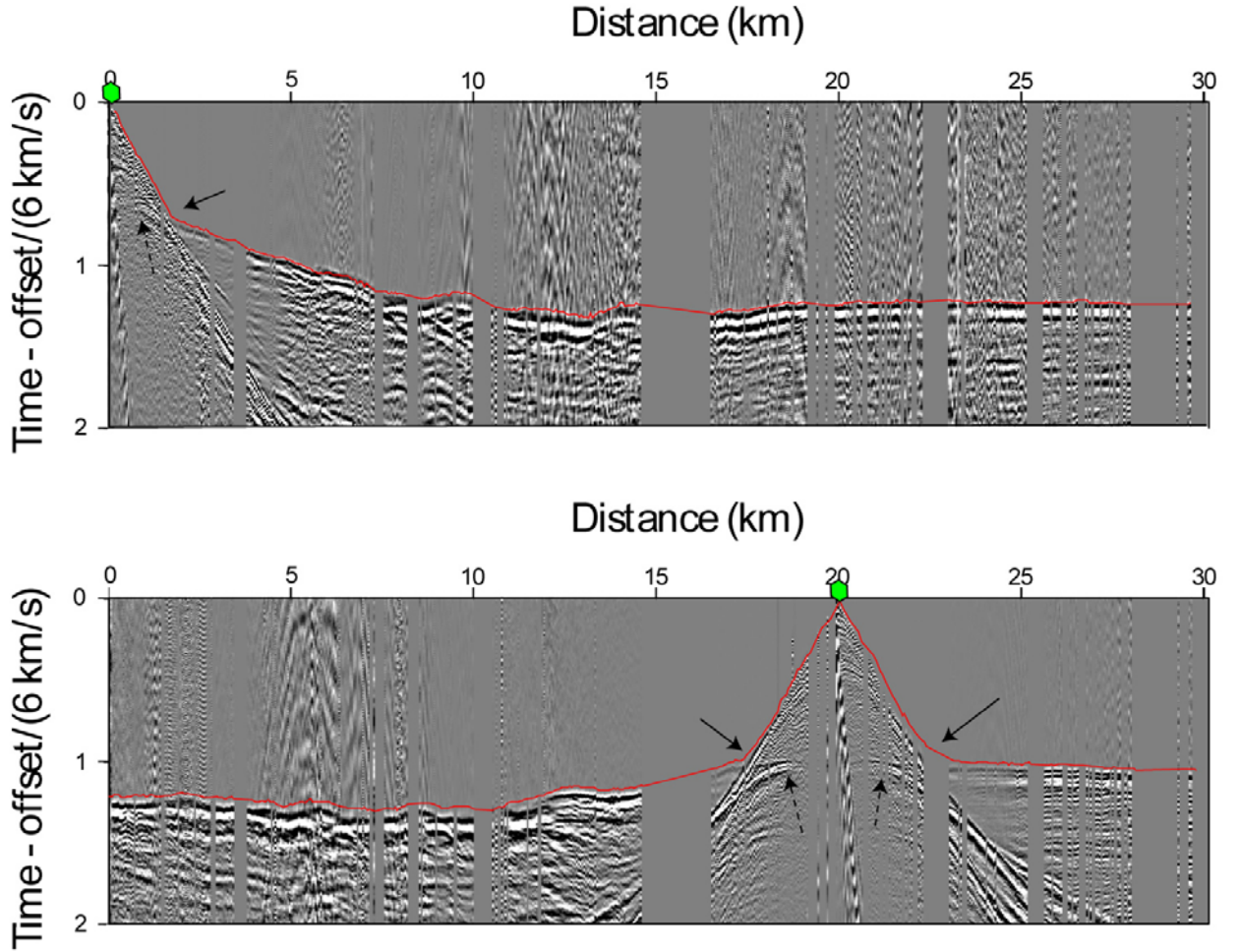


Figure 3. Shot gathers for shot 1 (top panel) and shot 19 near the edge of the crystalline crater (bottom panel). The time scaled is reduced by 6 km/s and the amplitudes are trace-normalized. Shot locations are labeled in Figure 1b. First arrivals are shown as the red line, shot location as the green hexagon. Shot 19 shows strong asymmetry between arrivals from inside (left) and outside (right) the crystalline crater. The cross-over point (solid arrows) is the source-receiver offset at which refracted waves from basement arrives earlier than direct waves through the sediment. Reflections from basement are indicated by dashed arrows. Significant structure is evident in basement first arrivals.

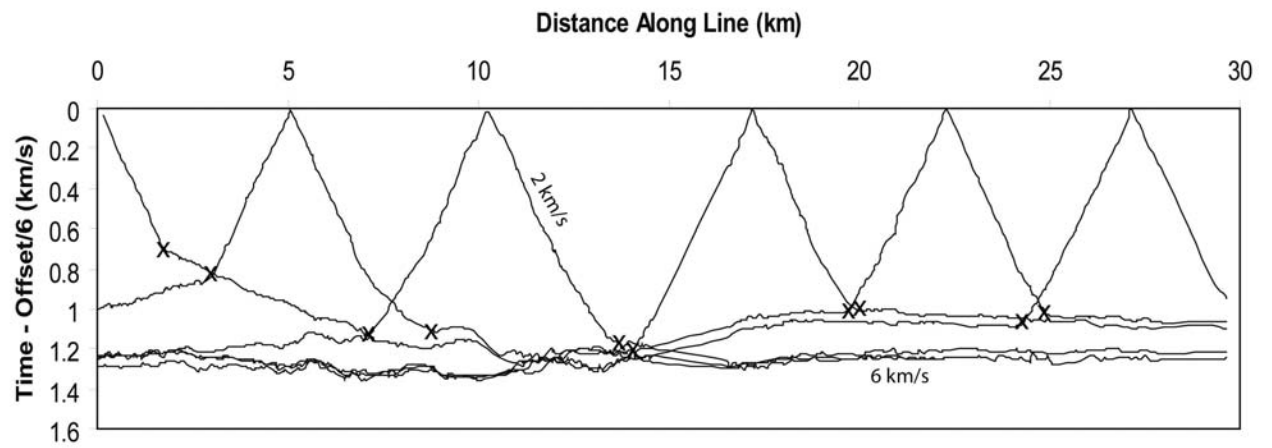


Figure 4. First arrival times, reduced by 6 km/s plotted for shots every 5 km. Bold “X”s are cross-over points, and approximate apparent velocities are labeled.

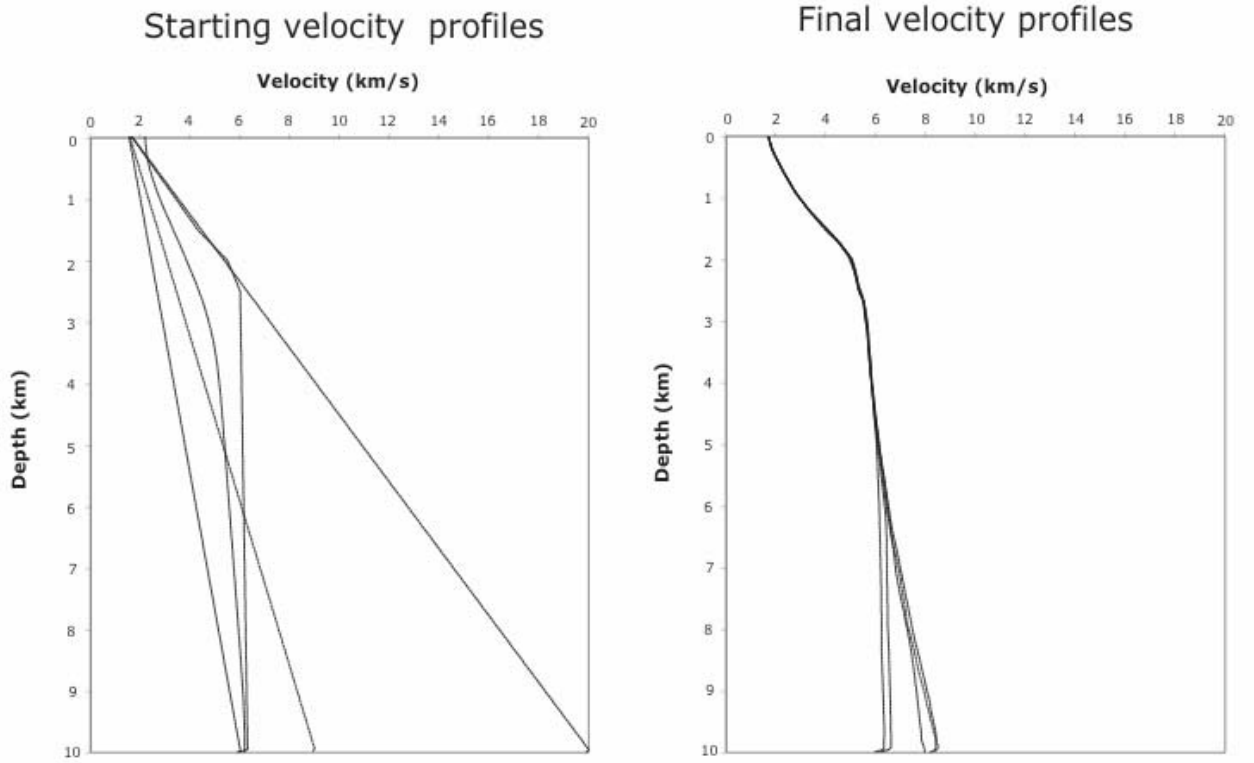


Figure 5. Comparison of different starting velocity models (left panel) and corresponding final models (right panel) for travel-time tomography. The final models are 1D profiles taken at ~ 10 km along the line. The final models all converge to the same solution above a depth of ~ 5 km.

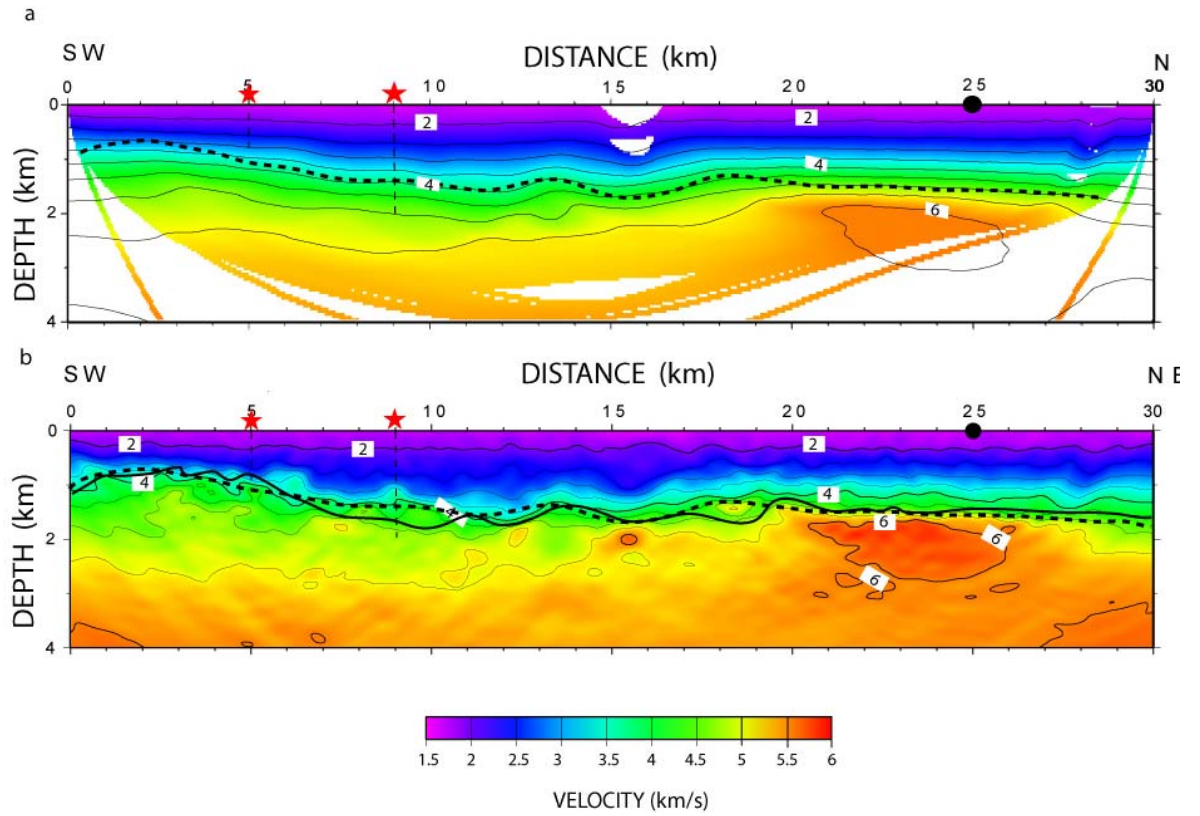


Figure 6. a) velocity model produced by travel-time tomography. White areas indicate no ray coverage. Dashed line is interpreted crater from maximum vertical velocity gradient (Figure 12). B) 3-16 Hz waveform inversion velocity model showing improved structural detail. The dashed line is from (a), and solid line is the interpretation of crystalline basement from Figure 15, and the dotted line is the interpretation based on travel-time inversion (Figure 9). The red stars and underlying dashed lines are the Eyreville borehole and the projection of the Cape Charles borehole. The black dot is the location of another 1D profile shown in Figure 13.

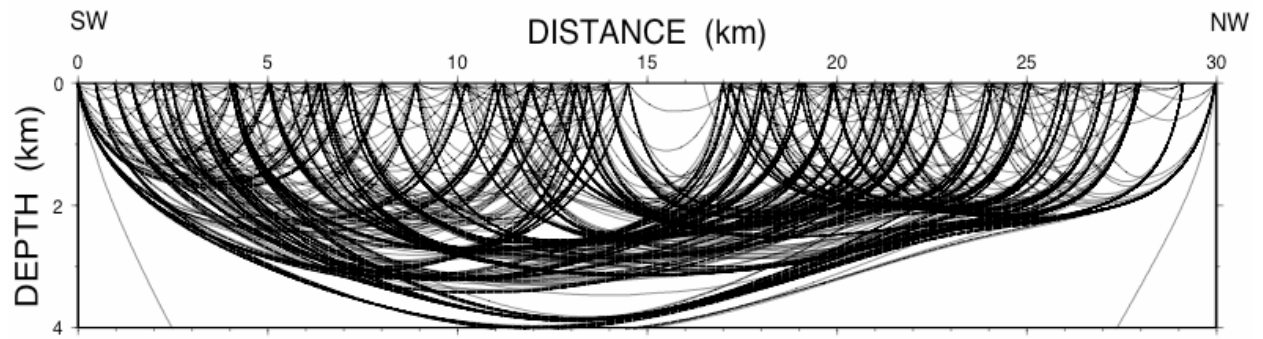


Figure 7. Ray coverage (every 10th receiver only) traced in velocity model of Figure 6a.

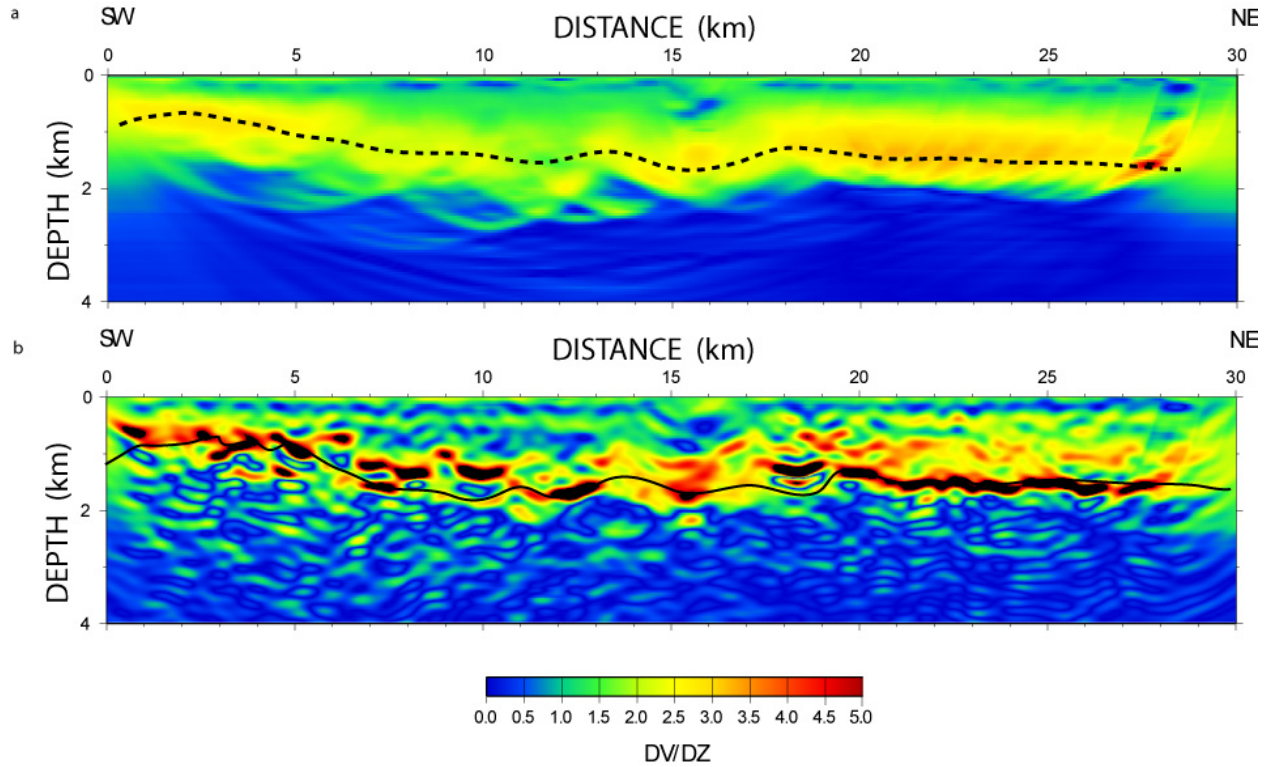


Figure 8. Vertical derivative of (a) travel-time tomography velocity model (Figure 6a) and (b) waveform tomography velocity model (Figure 6b). The maximum gradient of (a) is interpreted as the contact between crystalline basement and sediment (dashed line); the maximum gradient of (b) is interpreted as the contact between crystalline basement and sediment in many locations, and the contact between sediment and granitic megablocks in other locations (solid line). The contact between post-impact deposits and impact breccia follows a subtle gradient at shallow depths (~ 0.4 – 0.5 km) in (b).

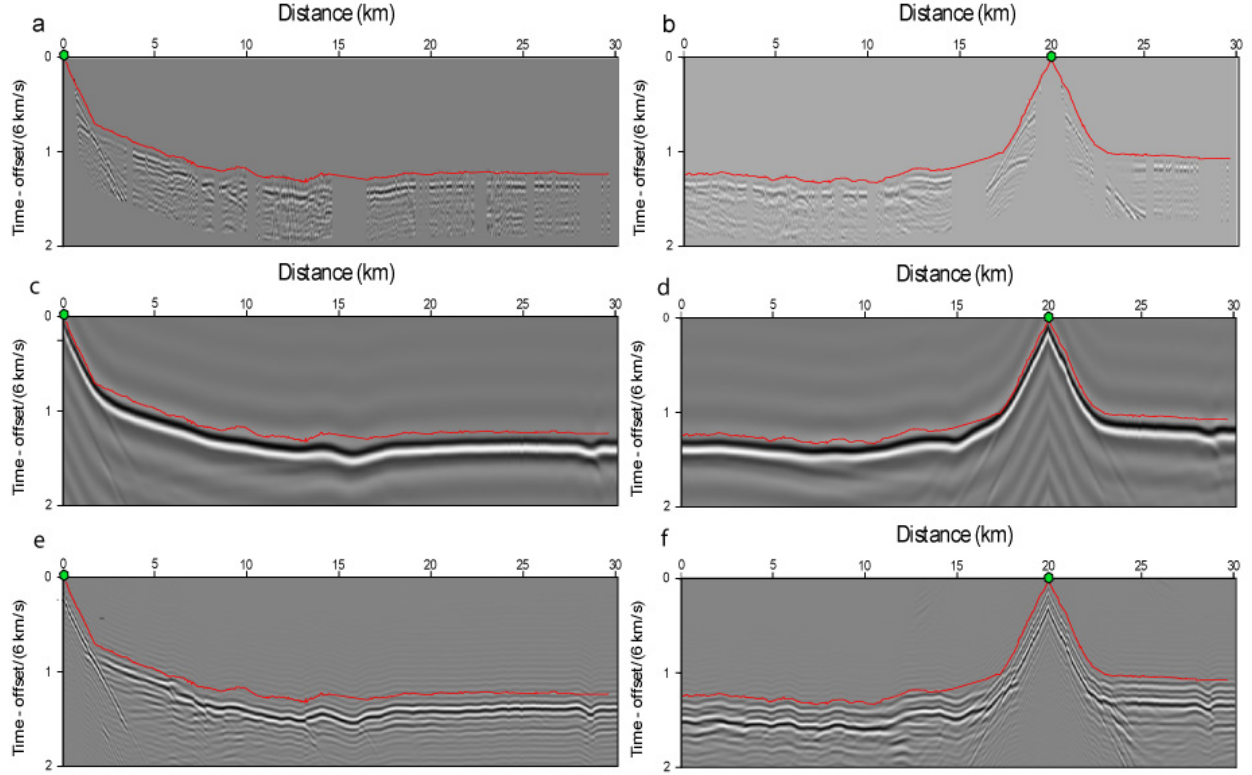


Figure 9. (a) Shot 1 and (b) shot 19 gathers pre-processed for waveform inversion. The data have been amplitude corrected, time windowed, muted at near offsets, and bandpass filtered (3-18 Hz). Only the first 10 km offsets were used in the inversion. Only the first 10 km source-receiver offset was used in the inversion. Travel-time picks are in red. (c) and (d) Corresponding synthetic shot gathers 1 computed in the travel-time tomography velocity model (Figure 9a) using frequency domain forward modeling technique of Pratt (1999). Source is a Keuper wavelet with central frequency of 14 Hz. Predicted arrivals match travel-time picks (red line) to less than half wave period at 6 Hz, indicating a sufficiently accurate starting model for waveform tomography. Bottom panel: Shot 1 synthetic gather (a) and shot 19 synthetic gather (19) calculated in the waveform tomography velocity model of Figure 9 with the source signature of Figure 13.

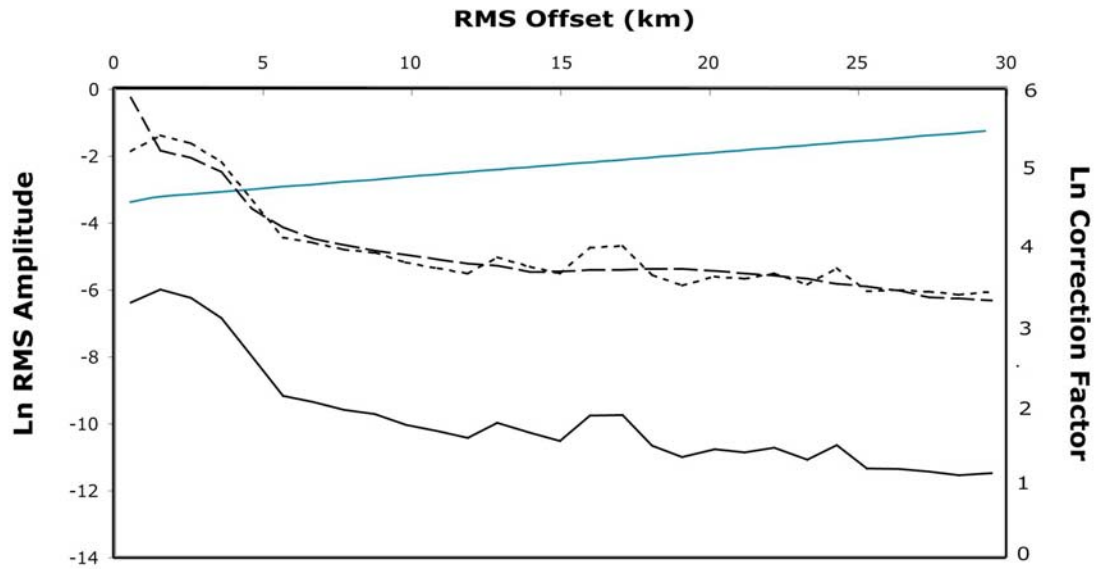


Figure 10. RMS amplitudes computed in 1 km offset bins using all shots gathers. Solid line is observed data, dashed line is synthetic data calculated in travel-time tomography model (Figure 9a), blue line is the amplitude correction (straight line in this log-linear plot), and dotted line is corrected observed data.

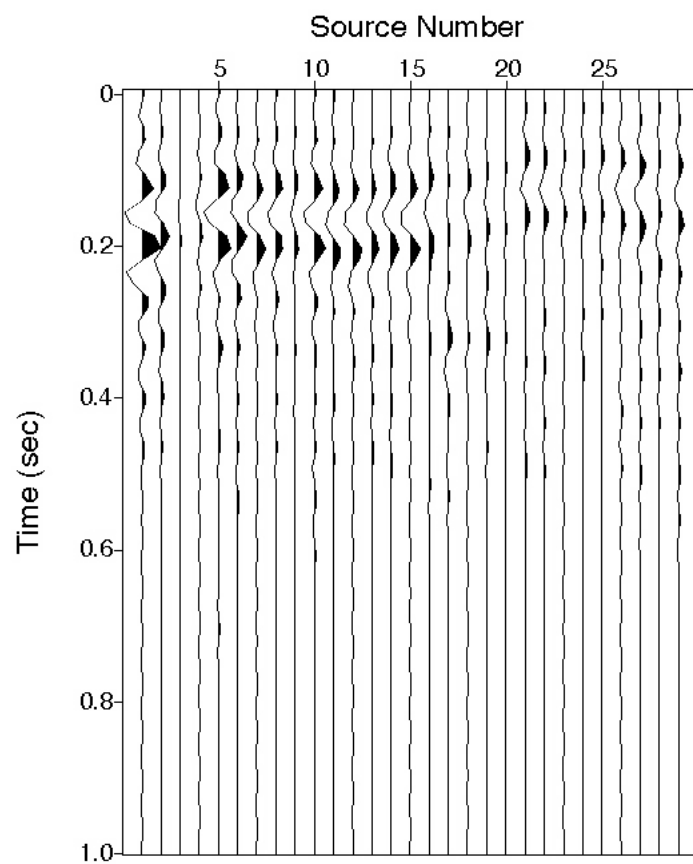


Figure 11. Final source wavelets after inversion for 3-16Hz.

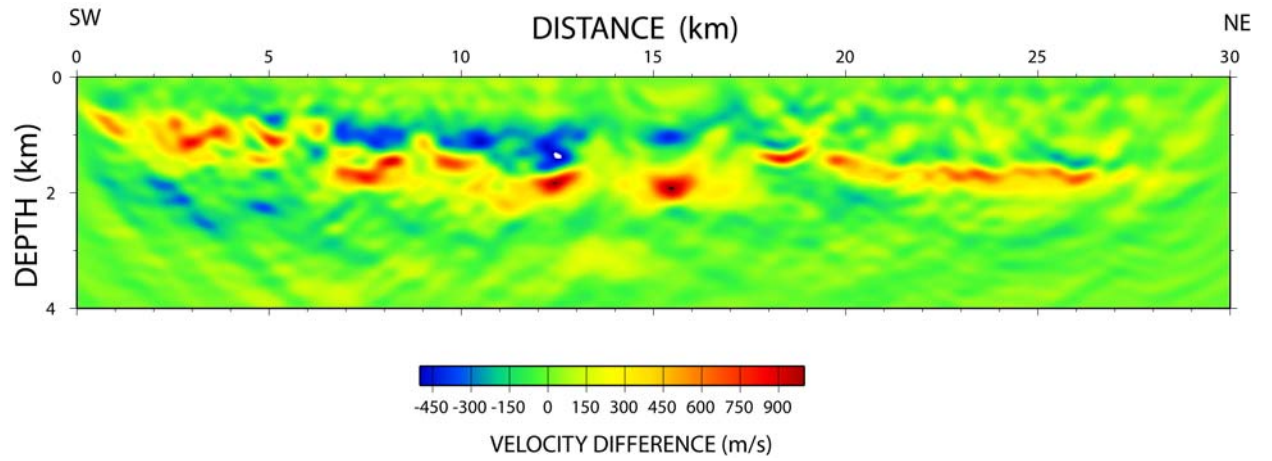


Figure 12. Velocity changes from travel-time (Figure 6a) to waveform inversion (Figure 6b). Red is waveform velocity model faster than starting model.

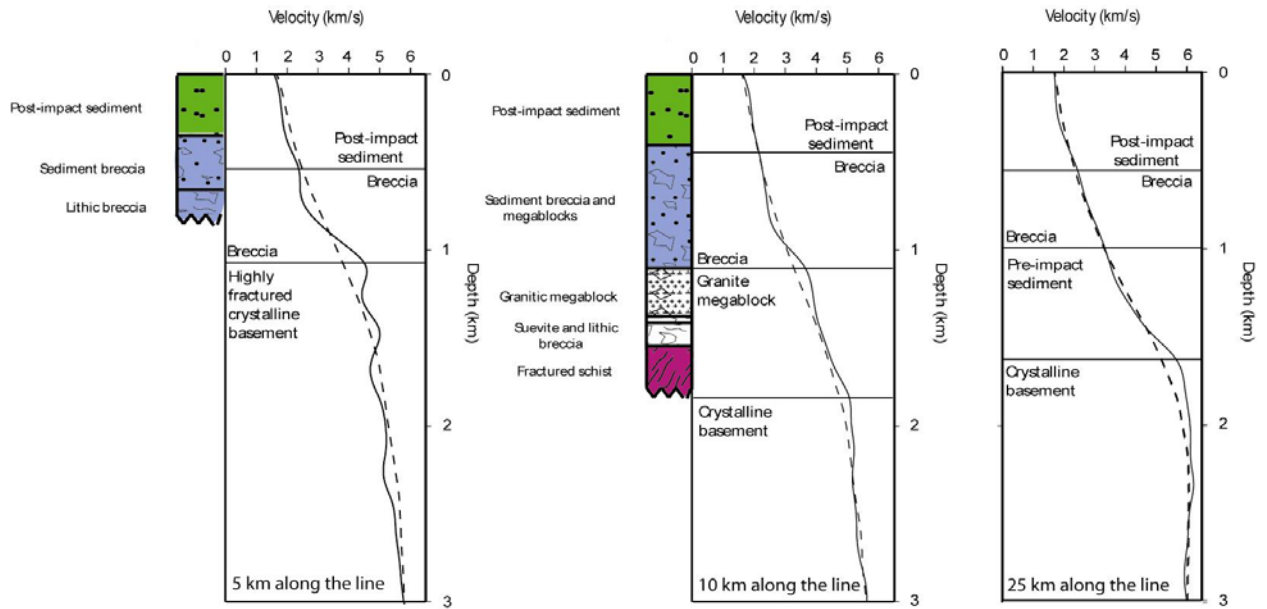


Figure 13. Velocity profile at model distances 5, 9 and 25 km from travel-time tomography (dashed lines; from Figure 9, top panel) and waveform tomography (solid lines; from Figure 9, bottom panel). Interpreted contacts are indicated: A – pre-impact sediment/basement contact; B – impact breccia/crystalline basement contact; C-post-impact sediment/impact breccia contact; D – breccia/granite megablock contact. Borehole lithologies observed in Cape Charles (projected from off-line) and Eyreville wells are indicated for appropriate profiles.

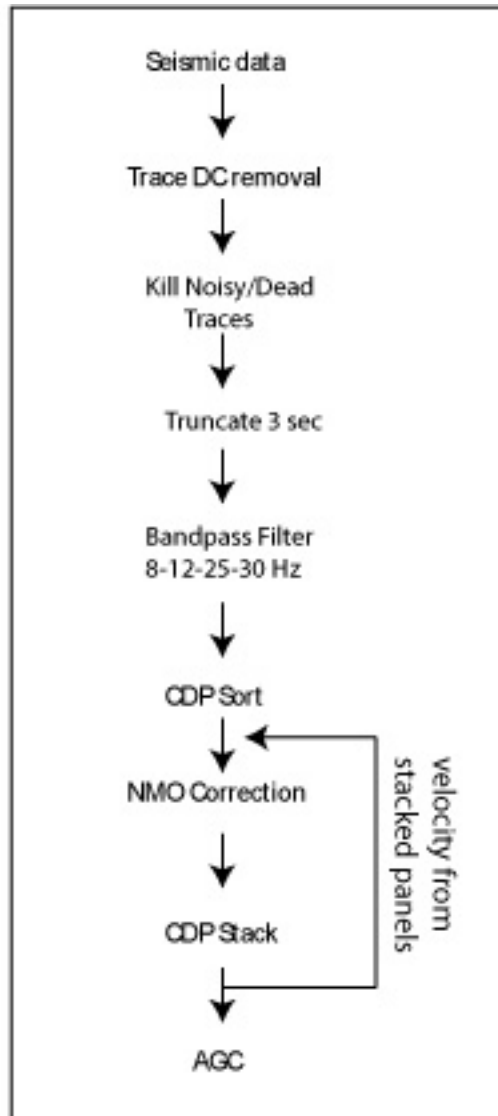


Figure 14. Reflection processing flow. The data has had DC bias and noisy/dead traces removed, and been truncated at 3 seconds, bandpass filtered (trapezoidal filter with corners at 8-12-25-30 Hz), sorted, NMO corrected and stacked according to best stacking velocities determined from constant velocity panels. An AGC window of 350 ms was applied for display.

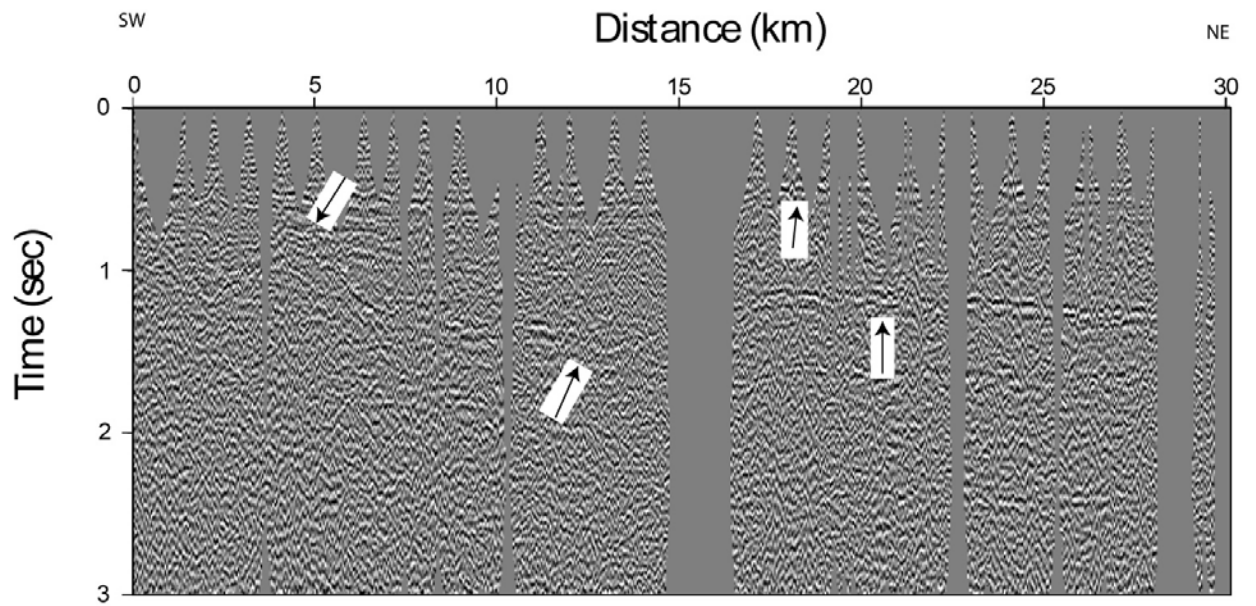


Figure 15. Reflection image of inner crater of the CBIS, resulting from processing flow outlined in Figure 14. A strong reflector is evident at ~ 1.3 s from 17 km to the northeast end of the line. The reflector continues across the landowner gap, but arrives later and with less amplitude. A strong sub-horizontal reflector is also evident at 0.5 s across the profile.

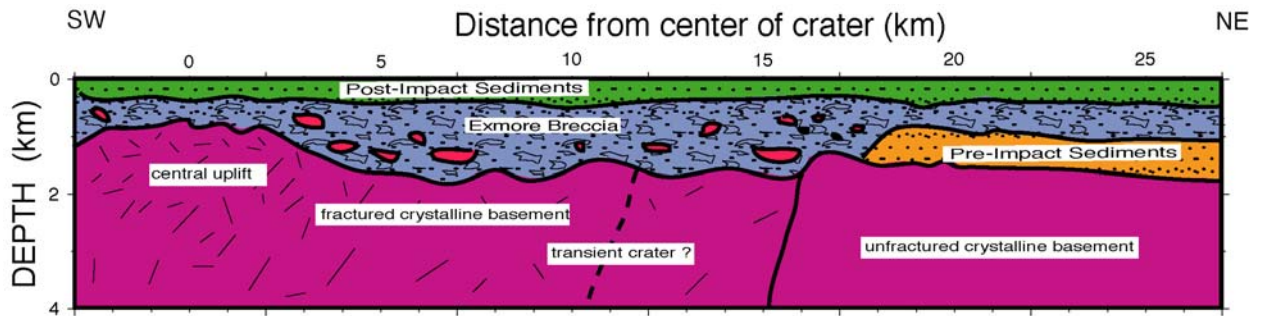


Figure 16. Model of the inner crystalline crater of the CBIS as interpreted from the seismic models. Pre-impact basement is at 1.5-1.6 km in depth. The crater margin is ~15 km from the center of the crater, and has a complex faulted and uplifted basement block structure. The inner crater is at a depth of 1.5-1.8 km and contains structural complexity and crystalline megablocks. Crystalline basement under the crystalline crater and central peak is highly fractured, with fracture volume increasing towards the crater center. The central peak is 8-10 km in diameter and is uplifted to a depth of ~ 0.8 km. The transient crater is represented by a local basement uplift and a lateral change in fracture volume.

Vita

Education

2004-2006	M.S. in geophysics, Virginia Tech
1999-2004	B.S. in Geoscience, minors in physics and astronomy Trinity University, San Antonio, TX
2004	SAGE geophysics field camp
2004-2006	M.S. in geophysics, Virginia Tech, in progress (expected date of completion: Spring 2006)

Honors/Affiliations

2006	Torgeson award for teaching excellence
2006-present	AGU membership
2005-present	AAAS membership
2005	ConocoPhillips Graduate Fellowship
2004-2005	ChevronTexaco Graduate Fellowship
2004-2005	Graduate Student Liaison Committee
2004	W. A. Tarr Award (awarded by Sigma Gamma Epsilon)
2002-2004	Sigma Gamma Epsilon, the geology honor society
2003-2004	Treasurer, Sigma Gamma Epsilon
2003-present	SEG and AAPG membership
1999-2003	Trinity University Presidential Scholarship

Research Experience

2004-2006	<u>Master's Thesis Project</u> , Virginia Tech Advisor: John Hole Acquire data (field work in Oct. 2004), use waveform and travel-time tomography to create model of Chesapeake Bay meteor impact crater, interpret structure of crater in support of drilling and in terms of geologic history of Chesapeake Bay area
2004	<u>SAGE Geophysics Field Camp</u> (In New Mexico) Acquire, process and interpret data for a variety of geophysical techniques (seismic and potential field methods) for three projects: Characterizing leakage under a dam, which was damaging Native American land; Shallow geophysics in support of an archaeological survey,

	Seismics to study a fault bounding a basin in the Rio Grande Rift
2003-2004	<u>Independent Research Project</u> , Trinity University, Advisor: Glenn Kroeger Processed shallow seismic data sets, including shallow seismic data to image a buried landfill
2002; 2004	<u>Research Participation</u> , Trinity University, Advisor: David Hough Analyzed radio telescope data from quasars to determine quasar structure, variability and timescale
2004	<u>Research Participation</u> , Trinity University, Advisor: Gordon MacAlpine Analyzed long-slit spectra from the Crab Nebula to determine distribution of heavy elements

Papers, Abstracts and Presentations

2006	<u>Structure of the Chesapeake Bay Impact Crater from Wide-Angle Seismic Waveform Tomography</u> , Lester, W, Ryan; Hole, John A.; Catchings, Rufus; Bleibinhaus, Florian; AGU abstract and presentation
2006	<u>Structure of the Chesapeake Bay Impact Crater from Wide-Angle Seismic Waveform Tomography</u> , Lester, W, Ryan; Virginia Tech GSRS abstract and presentation
2006	<u>Applying Full Waveform Inversion to Wide-Angle Seismic Data</u> , Bleibinhaus, Florian; Lester, Ryan; Hole, John A.; AGU abstract and presentation
2006	<u>Applying Full Waveform Inversion to Wide-Angle Seismic Data</u> , Bleibinhaus, Florian; Lester, Ryan; Hole, John A.; Deep Seismix abstract and presentation
2006	<u>Waveform Tomography of the Chesapeake Bay Impact Crater</u> , Lester, Ryan; Hole, John A.; Eastern AAPG abstract and poster
2005	<u>Waveform Tomography of the Chesapeake Bay Impact Crater</u> , Lester, Ryan; Hole, John A.; SEG Student Expo abstract and poster

2005 Waveform Tomography of the Chesapeake Bay Impact Crater,
Lester, W, Ryan; Virginia Tech GSRS abstract and presentation

Teaching Experience

Spring 2006 TA, Elements of Geophysics (GEOS-3104), Virginia Tech
Introductory geophysics for geoscience majors and engineers; organize and run computer and lab components

Fall 2005 TA, Potential Fields (GEOS-4124), Virginia Tech
Lab section for upper division potential fields course; organize, and run field, computer and lab components; help plan lab exercises

Professional Experience

Fall 2006 – present ConocoPhillips
Geophysicist, interpret seismic data for hydrocarbon potential

Fall 2005 Virginia Tech Teaching assistant
Teaching assistant for upper division geophysics lab (Potential Fields), plan and conduct lab experiments using potential field methods

Aug. 2003 Landex-(independent consulting geophysicist)
Seismic and well-log interpretation for petroleum exploration

Jun-July 2003 Baker Hughes
Build and operate gravel pack simulation

June 2003 ExxonMobil
Compiled and edited well logs and well files

June – Aug 2002 Trinity University Student Worker
Digitize archive of geosciences slides