In plain sight: The Chesapeake Bay crater ejecta blanket

David L. Griscom
impactGlass research international, 3938 E Grant Rd #131, Tucson, Arizona 85712-2559, USA

ABSTRACT

On the basis of a literature review and impact-geologic theoretical considerations, it is argued that the loams and gravels that blanket ~10,000 km^2 of the U.S. Middle-Atlantic Coastal Plain in an arc about the buried Late-Eocene, 90-km-diameter Chesapeake Bay structure display morphologic, lithologic, and stratigraphic features that are consistent with their being ejecta from that crater and are absolutely incompatible with the currently held belief that they are the work of rivers in Late-Miocene or more recent times. Also supporting this conclusion is a wide variety of unusual clasts found within the upland deposits that both individually and as a group have no other interpretation than as being ejecta from a crater whose target area included (1) minor pelagic limestone overlying (2) a deep accumulation of siliciclastic sediments ranging from the finest silts to well rounded pebbles, cobbles, and boulders largely comprised of Paleozoic quartzite and (3) a granite component within the crystalline basement.

INTRODUCTION

From the moment of the buried structure's discovery, the conventional wisdom has been that the ejecta blanket of the 35.5-Ma, 90-km-diameter Chesapeake Bay crater (Poag et al., 1994; Koeberl et al., 1996; Poag 1997; Poag et al., 2004; Gohn et al., 2008) had long since been removed by erosion from the surface of the U.S. Middle-Atlantic Coastal Plain and the abutting Piedmont province. All seminal papers, books, and authoritative reviews of the subject tacitly imply that not even a scrap of ejecta has ever been found on the present-day surface.

Notwithstanding, there did arise one "voice crying in the wilderness." For 9 years and counting I have been gathering evidence that ~5,000 km^2 of "upland deposits" mapped in Southern Maryland, Virginia, and the District of Columbia are far better interpreted as crater ejecta than as fluvial deposits (Griscom, 1999, 2001, 2002, 2003a, 2007). Most significantly, Griscom et al. (2003a) employed a wide range of solid-state-physical methods to demonstrate that a type of iron ore endemic to the "upland deposits" is a peculiar nanocrystalline form of "ferric oxyhydroxide" characterized by atomic arrangements similar to Goethite but hardness closer to that of quartz. In retrospect, this material might turn out to be better described as a unique form of ferrihydrite (Michel et al., 2007). Thin-section photomicroscopy has revealed these hard ferric oxyhydroxides (ferrihydrites?) to comprise the matrix material of a type of matrix-supported breccia ubiquitous to the upland deposits, leading Griscom et al. (2003a) to argue that (1) they must be *melt-matrix* breccias because iron oxide precipitates too slowly from solution to solidify around free falling clasts and therefore that (2) they must be impactoclastic in origin because there are no known igneous rocks composed of 95% pure iron oxide.

I begin the present paper with a review of the lithology of the "upland deposits" and critically discuss their geomorphogeny in relation to the uniformitarian explanation historically assigned to them. I then compare their materials properties and geomorphography to quantitative predictions of Chesapeake-Bay-crater ejecta-blanket thicknesses and properties expected on the basis of well-established experimental, observational, and theoretical principles of impact geology (Melosh, 1989). In addition, I report field studies of a diamict siliciclastic stratum discovered topographically below the upland deposits in Northern Virginia and several unusual clasts found therein, which I argue can only be interpreted as crater ejecta. And finally I describe a unique granite boulder found within the upland deposits, which I argue is a meteorite from the Earth (Gladman et al., 1996) – almost certainly a fragment of basement rock ejected from the interference zone of the Chesapeake Bay impactor.

THE UPLAND DEPOSITS: NOMENCLATURE AND HISTORY

The "upland deposits," sometimes termed "upland gravels" or "upland terrace gravel," are not consistently designated by a single formation name, despite their having been mapped over an area as large as 5,000 km^2. Therefore, I will employ the nomenclature "upland deposits" without the quotation marks to denote the particular lithofacies to be described below, wherever they may have been mapped (Fig. 1), irrespective of any other names that may have been historically assigned to them. The upland deposits of the Patuxent-Potomac peninsula of Southern Maryland are denoted the Brandywine formation (formerly the Lafayette), whereas lithologically similar upland deposits blanketing lower terraces have traditionally been assigned different formation names according to their relative elevations, the most common example being the Sunderland at 52 m (e.g., Schlee, 1957). Ironically, the scarp-free Brandywine formation of Southern Maryland slopes gently from its highest elevation of ~82 m east of the District of Columbia down to just ~30 m about 100 km to the southeast.

One good reason for not using the word "gravel" alone to characterize this particular upland unit is the fact that the upland deposits comprise a *bilayer consisting of a lower gravel member and an upper loam member* (Hack, 1955). The overall thickness of the upland deposits normally ranges from ~3 to ~10 m (Hack, 1955), although exceptional thicknesses up to 15 m have been noted. According to Schlee (1957), the *gravel member* is fairly well stratified and sedimentation units range from ~2 m down to ~5 cm in thickness, while the contact with the overlying loam is gradational over ~0.5 to 1 m.

The origin of these unusual deposits had been controversial for well over a century when Schlee (1957) interpreted his important new data in terms of Hack's (1955) version of the fluvial-deposition model. In a handy overview of the geology of Virginia, Frye (1986) describes the upland deposits in terms of the Hack (1955) model, including in particular the conclusion that these deposits were laid down by rivers in Miocene, Pliocene, or Pleistocene times, given that they overlie clay terraces long-ago dated as Miocene (see also, Hack, 1955; Schlee, 1957). A recent check of web sites treating the U.S. Mid-Atlantic Coastal Plain suggests that the Hack-Schlee model has not changed in the slightest since the discovery of the Chesapeake Bay crater. Notably, however, *the upland deposits themselves are generally devoid of fossils and therefore have been dated solely on the basis of the inferred ages of the fossils in the shallow-water clays that immediately underlie them* ...without modern radiometric confirmation.

The loam member

According to Schlee (1957), the "massive" upper loam member consists chiefly of compacted silt with some sand, clay, and scattered pebbles, and ranges in color from pale-yellowish-brown to grayish-orange to moderate-yellowish-brown. Hack (1955) reported this loam to be about 90% quartz silt, with 2-5% heavy minerals. My personal experience from living in a home perched on a steep embankment composed of this upland loam is that it is so coherent that grass roots can scarcely penetrate it, yet even a steeply-sloped surface partially denuded of grass showed no noticeable sign of pluvial erosion in 30 years.

The gravel member

Schlee (1957) dedicated his field work to elucidating the *lithology of the gravel member* of the upland deposits, mainly in Southern Maryland (88 sites). However, he also sampled the District of Columbia (7 sites) and Northern Virginia (3 sites ~10 km west of Washington, D.C.). The samples analytically studied by Griscom et al. (2003a), as well as several that I will illustrate for the first time below, were collected ~10 km south of Washington in Springfield, Va., and near Alexandria, Va. (roughly 15 km southeast of Schlee's 3 Northern Virginia sites).

According to Schlee (1957), the gravel member is pale-to-dark-yellowish-orange to mottled pale-yellowish-brown poorly sorted gravel and sandy gravel, as well as very pale-orange gravelly clay. "The gravel member is fairly well stratified, and sedimentation units [ranging from 2 m to 5 cm thick] are easily recognized." "Where bedding is poor, sedimentation units up to [5 m] thick are present. Laterally the beds range from [30 to 60 m] or more down to [3 m] or less in the case of sand lenses."

Schlee (1957) continues "The gravel has a *bimodal size distribution* and consists of a gravel framework, averaging 61% of the whole, and a matrix of sand, silt, and clay (emphasis added)." This fact led him to conclude that the bed load of heavy gravel and the suspended load of sand, clay and silt were likely to have been deposited simultaneously (very unusual, if not impossible for rivers). And he was perfectly right. His carefully determined average framework of 61% is surely statistically valid, and this is significantly less than the theoretical maximum for random close packing of spheres, i.e., 64% (Jaeger and Nagel, 1992).

Correlation of the upland deposits of southern Maryland with those of eastern Virginia

If, as I believe, all of the upland deposits mapped in Figure 1 came out of the Chesapeake Bay crater, then any given mapped locality should be lithologically similar to all others. Strong support for this notion comes from the fact that Wentworth's (1930) extensive field work led him to apply the name Brandywine to the upland deposits of eastern Virginia. Moreover, note the similarity of the following descriptions of two widely separated upland-deposit locations, the former being an average over all of Southern Maryland and the latter pertaining to a small cluster of sites west of Richmond, Va., which I will described in greater detail later. These sampled locals are separated laterally by an average distance of ~140 km, vertically by ~40 m, and angularly by ~60° with respect to the crater center:

Schlee (1957), regarding the upland deposits of Southern Maryland:

(a) Describing *the gravel member only*: "The predominance of mature siliceous rocks is one of the unusual features of the upland gravels..." "In the 64- to 128-mm size grade, quartzite usually far exceeds 'vein' quartz and chert." "Other rock types constitute a few percent or less of the total." "In all but two analyses [out of 72], the size distribution is bimodal..." In addition to showing 12 of these 72 distributions in a figure as bar graphs spanning a size range of <1/16 mm to 128 mm, Schlee (1957) presented a histogram of the modal sizes of those 72 distributions, which is indeed bimodal, showing 81% of the modes to fall in the gravel group (4 to 128 mm) and 19% in the sand group (1/4 to 1 mm).

(b) Footnote: "The 'only fossil' seen by the author was *Cassius madagascariensis spinella* Clench, found at the gravel-loam interface near Silver Hill, MD... It is unlikely that a tropical marine form such as this... was indigenous to the upland deposits; it may have been buried by Indians." (N.B. The Earth was a "hothouse" for ~ 500 m.y. prior to the Late Eocene.)

(c) "Pseudo-bedding in the form of wavy bands, which parallel and transect the bedding at a low angle, occurs in many exposures. The bands are chocolate to ruddy brown, well indurated, and a fraction of an inch thick." "...secondarily introduced iron oxide ... locally cements the sand and gravel along definite zones and in large irregular masses up to [1 m] across." N.B. Griscom et al. (2003) have argued that these hard ferric oxyhydroxide (ferrihydrite?) bands can be interpreted in no other way than as impactites.

(d) Speaking of the upland deposits of Southern Maryland as a single unit: “This sheetlike deposit, which successively overlaps older formations to the northwest, has a fairly constant thickness of [6 to 9 m] and dips southeastward at approximately [1 m/km]” (in the direction of the crater).

Goodwin and Johnson (1970), RE the Midlothian gravels west of Richmond:

(a) “The most striking feature of the Midlothian gravels is the thick gravel member. The gravel member, which ranges in thickness from zero to nearly [12 m], is composed of abundant pebbles and cobbles in a sandy matrix containing variable amounts of clay. Although cobbles up to [230 mm] in maximum diameter were observed, the average cobble measures [50 to 75 mm] in diameter.” “The clasts are variable in composition but most are of some variety of quartz. Vein quartz, massive quartz, quartzite and quartz sandstone are the dominant lithologies comprising the clasts. A few clasts of other metamorphic rocks, igneous rocks, and chert constitute a minor portion of the gravels.”

(b) “No fossils have been found within the gravels to aid in determining their age or their origin.”

(c) “The basal cobble zone is one to four cobbles thick and is commonly solidly cemented by iron oxide to form a ferricrete zone.”

(d) “The [~60 km^2] upland gravels around Midlothian [~10 km west of Richmond, Va.] are isolated and have been truncated by erosion on all sides.” “Although locally the contact between the gravels and underlying rocks exhibits more than [6 m] of relief due to channeling, the surface of unconformity dips gently to the east or southeast at approximately [1.7 m/km]” (in the direction of the crater).

UPLAND DEPOSITS: A GRAVEL-SIZE GRADIENT RADIAL TO THE CRATER

Perhaps the most illuminating of Schlee’s (1957) many innovative contributions to our knowledge of the upland deposits was his experimental demonstration and mathematical analysis of the dramatic cobble-size gradient decreasing from 32-mm modal size in Washington, D.C., down to 4 mm midway down the Patuxent-Potomac peninsula (Fig. 2). Figure 2A, taken from Schlee (1957), shows the isoliths of modal gravel size that he determined for Washington and the Patuxent-Potomac peninsula. In contemplating these isoliths, Schlee (1957) observed that “The orderly changes in modal size in the northern portion of the sheet suggest that the size change may be a mathematical function similar to those found for the size diminution in modern rivers.” He therefore selected “four profiles parallel to the current flow direction and as far as possible normal to the size contours,” plotted them on the semilog graph (Fig. 2B), fit them to a straight line, and then compared the slope of this line to those reported for the modern Rhine and Mur Rivers.

Given that uniformitarian geology was "de rigueur" in those days, it is understandable that Schlee (1957) incorrectly described the slopes of those three curves of Figure 2B as "of the same general order of magnitude." In fact, the cobble-size-reduction-rate slope that he was forced to attribute to the ancestral Potomac River is well over one order of magnitude greater than those of the modern Rhine and Muir Rivers, each of which has cut channels into crystalline rock. The contradiction grows worse when one considers that the ancestral Potomac River envisioned by Hack (1955) was now proven to have ground down the upland cobbles at the absolutely fantastic rate implied by Figure 2B, *while at the same time failing to cut a channel down to bed rock in the "soft easily eroded Coastal Plain sediments* (Schlee, 1957)*."*

Still, Schlee (1957) deserves credit for realizing that, whenever data follow an exponential law as well as they do in Figure 2B, it should be possible to safely extrapolate them forward or backward by at least one or two factors of two. So he did indeed extrapolate the upland gravels of the Patuxent-Potomac peninsula and the District of Columbia to a modal cobble size of 128 mm in an area ~18 to 30 km northwest of the U.S. Capitol Building. However, he found this putative "source area" (roughly within the dashed circle in Fig. 2A) to comprise the "granites, diorites, schists, gneisses, and quartzites of the Piedmont province," in sharp contrast to the coarser fraction of upland deposits, which "contain little or no gneiss, schist, diorite, or granite." Although Schlee (1957) does not explicitly state that he found no 128-mm sized quartzite cobbles in that area, one may safely assume that, if he *had* found them, his story line would have been very different.

So, where *was* the source of the quartzite component of the gravel member of the upland deposits? Well, Schlee (1957) remarked that "Though generally unfossiliferous, a few of the pebbles display fossil brachiopod impressions which indicate a Devonian age: the source rock was probably the Oriskany sandstone." He noted, however, that the Oriskany formation is ~130 km distant from the upland gravels, whereas the nearest outcrop of Devonian quartzite is the Weaverton formation at ~58 km.

However, given the total absence of a trail of ever-larger quartzite boulders leading back to either of these putative sources, Hack's (1955) fluvial model is seen to completely fail the central test of *uniformitarian* geology: We do *not* see today's rivers transporting cobbles dozens of kilometers before depositing the first one. Although such things could happen in the case of superfloods (Baker, 2002), this scenario would surely have resulted in the stripping of the soft coastal plain sediments *before the deposition of the first cobble*. Thus, it is truly difficult to imagine any single river behaving in the manner that Hack (1955) imputes to the ancestral Potomac River ...*much less the ancestral Potomac, Rappahannock, York, and James Rivers all performing this same "ballet fantastique," as though choreographed by Nijinsky!*

ADDITIONAL UPLAND-DEPOSITS FACIES BEST EXPLAINED BY IMPACT

With reference to properties of the upland deposits already mentioned

Regarding Schlee's wavy bands locally cementing the gravels, I note that those ~1-cm-thick "wavy bands" are better described as "wavy *sheets*" and that, in all cases I have inspected, *such sheets have been terminated by fresh conchoidal fractures* – implying that they were fractured during and/or just before emplacement. Many examples of such fracture surfaces are illustrated in Griscom et al. (2003a).

Here I propose an explanation for why, after investigating fully 98 locations, Schlee (1957) reported the *largest* "irregular mass" cemented by "fraction of an inch thick" bands "which parallel and transect the bedding at a low angle" to be only ~1 meter across (as opposed to many meters across as might be expected under the canonical notion that they were precipitated from water solution). If, based on the observations of Griscom et al. (2003a), it is assumed that all such "irregular masses" were not much thicker than the average pebble or cobble webbed together by those hard ferric-oxyhydroxide(ferrihydrite?)-matrix sheets, then Schlee's largest object of this nature would have had an aspect ratio of ~10 to 1. It turns out that such large aspect ratios are expected of interference-zone impact ejecta. Melosh (1989) has pointed out that "At low speeds (up to a few hundred meters per second) the near surface ejecta consist of spall plates. *These plates are several to ten times broader than they are thick* (emphasis added)."

Melosh (1989) further instructs that "[Spall plates] are the largest and least shocked fragments thrown out at any given velocity" but that they "...contain so much elastic energy from the interfering stress waves that they themselves break up into smaller, Grady-Kipp fragments immediately upon ejection." Thus, the maximum size of an interference-zone ferric-oxyhydroxide-welded-quartzite-pebble spall plate that will commonly survive both launch and re-impact may well turn out to be of the order of just 1 meter. N.B. An impact-based mechanism for *creating* such "peanut-brittle-form" objects has been sketched by Griscom et al. (2003a).

Finally, not mentioned above but certainly pertinent, is the fact that in some exposures the upland deposits include "extremely large boulders ranging from [~0.02 m^3] to [3.7 m^3] (Schlee, 1957)." I have observed a great many such boulders in a broad area south of Alexandria, Va. The conventional explanation of these boulders, *ice rafting,* certainly must be called into question, given that the ice ages of the Quaternary are associated with extreme sea-level *lowstands* (Lambeck and Chappell, 2001). How then could boulders have been ice rafted to *upland* locations?

Some new lithological data from Northern Virginia

I present here some new lithological data bearing on the question of the origin of the upland deposits. As a quasi-random sample of the coarser upland gravels in our former neighborhood of Hollin Hills, Fairfax County, Va., I investigated 214 quartzite pebbles and cobbles already raked up by a neighbor (Fig. 3A). I assume that he likely suppressed smaller pebbles and possibly set aside larger cobbles, but at least this collection has no investigator-imposed biases. In particular, I decided to count the number fractured rocks in each size class. Then, since many of the samples were evidently fractured along two or more intersecting planes, I separately noted the number of rock fragments with more than one *non-parallel* fracture. For this purpose, I counted as "second fractures" any planar fractures running part way through any fragment of what was obviously once a well-rounded rock prior to the first fracture (the one which separated it from the rest of the original rock). However, I did not count as second fractures any cracks that were parallel to the surface of the first complete fracture, even though these occurred very commonly. In all, I noted only one rock fragment exhibiting the slightest evidence of post-fracture re-rounding (and I counted that one as not fractured). My results are shown in Figure 3B. Overall, 54.5% of the 32-64-mm fraction, 65% of the 64-128-mm fraction, and 57% of the 128-256-mm fraction had at least a single fracture. The advanced surface weathering of these rocks rules out the fractures being of very recent origin. In any event, it is clear that the present condition and location of these rocks (high prevalence of fresh fractures and lying atop a 60 m clay terrace) are inconsistent with fluvial deposition.

Figure 3C exhibits a quartzite cobble that I noticed elsewhere in the same residential yard (not in the pile of Fig. 3A). The fracture facies of this rock are reminiscent of the way glassy materials fracture under tensile stress. In fact, defect-free fused-silica optical fibers can survive up to 5 Giga Pascal tensile stress before fracturing (e.g., Bradt and Tressler, 1994); and when fracture finally occurs at such high tensile stress levels it leaves a characteristic "mirror-mist-hackle" fracture surface, as cartooned on Figure 3D. Even though optical fibers are typically ~50 to 500 μ in diameter (producing mirror-mist-hackle patterns analyzable only under a microscope), it seems reasonable to interpret the macroscopic fracture surfaces of this metaquartzite rock (whose fracture strength is also determined by the strength of silicon-oxygen bonds) as possible indications of fracturing under extreme tensile stress. While there does not seem to be a mist zone on the rock of Figure 3C, the mirror and hackle are unmistakable. In fact, the surface enclosed by the smaller dashed ellipse is much more mirror-like than one would expect for metaquartzite, which is normally subject to "uneven, splintery to conchoidal fracture (Chesterman and Lowe, 1978)." Unlike any other part of this particular rock, the "mirror" area exhibits vitreous luster and is so flat that a straight edge laid across it is within a fraction of a millimeter of the surface over the entire 7-cm width. Moreover, there is an internal planar fracture sub-perpendicular to the "mirror" that runs about two thirds of the way through the entire rock fragment; its intersection with the mirror surface (indicated by arrows) parallels the semi-major axis of the ellipse defining the mirror. I believe it will one day be confirmed that these unusual features resulted from two nearly orthogonal tensile waves having passed through this cobble, very likely at the same moment.

N.B. Tensile waves result from reflection of pressure (shock) waves at free surfaces (e.g., Melosh, 1989). Thus, for example, any pressure wave reflected from the surface of bed of water-saturated sediments will propagate backward into the bed as a tensile wave. And tensile waves of sufficient intensity to rupture a metaquartzite cobble surely occur in nature only as results of large impacts.

CALCULATION OF THE EJECTA BLANKET PROFILE

To aid in the search for proximal ejecta deposits of degraded or buried terrestrial craters, it is helpful to have a quantitative idea of what the ejecta blanket might have looked like when it was first emplaced. An empirical rule has emerged from target-strength-governed explosion experiments (McGetchen et al., 1973), which by very good fortune turns out to be applicable as well to gravity-dominated impact craters ranging from 1.3 km to 436 km (Melosh, 1989). The rule is that the heights $h(r)$ of ejecta blankets outside of the crater rim are typically proportional to the negative third power of the radius r from the crater center:

$$h(r) = f(R)r^{-3} \quad (r > R), \qquad (1)$$

where R is the transient-crater radius and the coefficient $f(R)$ depends in principle on specific details of the crater in question (Melosh, 1989). The total volume of ejecta V_{ej} present in such an ejecta blanket can be obtained by taking the area under a surface of revolution with the radial profile defined by equation 1 by integrating from $r = R$ and to $r = \infty$:

$$V_{ej} = \int_R^\infty 2\pi r \cdot h(r)\, dr = \int_R^\infty 2\pi f(R) r^{-2}\, dr = \pi f(R)/R \qquad (2)$$

Whence, $f(R) = RV_{ej}/\pi$. So, in essence, if one knows the crater radius R and the total volume of the ejecta blanket, then $f(R)$ is also known. However, the value of V_{ej} is uncertain to the degree that the amount substrate incorporation is not easily determined (Oberbeck,, 1975, Melosh, 1989). Thus, by using the amount of material excavated from the crater V_{exc} to approximate V_{ej}, as I will do below, I obtain a *lower limit* to the actual value of $f(R)$.

Following the general procedures adopted by Warren et al. (1996), I arrived at a value of V_{exc} = 4,508 km^3 for the Chesapeake Bay crater. However, I defer here to a similar number given by Poag (1997): 4,300 km^3. I used the final crater radius R = 45 km (Koeberl et al., 1996;

Poag, 1997) to arrive at the number I needed to calculate the (minimum) thickness of the Chesapeake Bay crater ejecta blanket:

$$f_{\min}(45) = 45V_{\text{exc}}/\pi = 61{,}600 \text{ km}^4. \qquad (3)$$

Next, to lend "realism" to my ejecta-blanket plots based on equation 1 and the scaling coefficient of equation 3, I decided to include a profile representing the crater itself. For this purpose I approximated the crater as a cylindrical hole in the ground of radius R having a depth adjusted to make its capacity exactly equal to the excavated volume V_{exc}. The depth calculated in this way turns out to be exactly equal to the height of the rim above the target surface. This cylindrical profile represents neither that of the transient crater nor of the final crater. By comparison with figures in (Koeberl et al., 1996) and (Poag, 1997), the result of Figure 4 might be viewed as a fictitious (never achieved) intermediate state before the collapse of an idealized (never realized) Chesapeake Bay crater rim.

In Figure 4, I have cartooned the depth of the crystalline basement to match the depth at which the USGS-NASA Langley Corehole at Hampton, Va., intercepted basement granite: 626.3 m (Horton et al., 2005). And to add further realism, I decided the give the target surface a gentle seaward slope of 0.5 m/km – just about the same as the present day slope of *the base* of the upland deposits seen in the Southern Maryland section (Fig. 5).

Except for the range r = 45±20 km, where details the final crater are badly represented, the calculated profile Figure 4 should paint a fairly accurate picture of the *lower-limit average* Chesapeake Bay crater ejecta blanket (neglecting dunes, hummocks, ridges, and rays) when it was only tens of minutes old and not yet assaulted by the resurge of the Atlantic Ocean – which, when the return wave finally appeared, surely carried the tallest part of the seaward rim back into the crater (Poag, 1997). The landward side, however, should have fared far better. The volume of water to the landward at the time of the impact would have been locally comparable to the volume of the landward solid ejecta, and this water would have been imparted radial momentum causing most of the westward-moving wave to deflect northeastward or southward, depending on the angle at which it struck the Blue Ridge. So the westward ejecta blanket should have experienced little or no return wave from the west and would have been shielded from the primary Atlantic resurge by the eastward facing blanket. Therefore, I argue that the landward crater rim should have remained in at least partially intact for a geologically substantial period of time.

In fact, I will propose that remnants of this ejecta blanket are still there. From my juxtaposition of the calculated Chesapeake Bay crater ejecta blanket profile on the U.S. Geological Survey sections in Figure 5, it can be seen that *the upland deposits of Southern Maryland could in principle comprise part of these remnants*. That they actually exceed the calculated ejecta depth at the longest distances is not a contradiction, since my calculation neglects the inevitable incorporation of (soft or friable) substrate material into the ejecta blanket (Oberbeck, 1975).

ASSESSING THE LANDWARD EJECTA BLANKET

If the upland deposits of Figure 5 are truly Chesapeake Bay crater ejecta, then three questions immediately come to mind: (1) Did the impact itself sculpt the gently sloping base of the Southern Maryland deposits, (2) do the bases of the upland deposits in Virginia also slope radially in the direction of the crater, and (3) do the contiguous terrains (Bacons Castle formation) between the upland deposits and the crater rim also harbor vestiges of ejecta? To help answer these questions, I went to Google Earth and picked off the elevation profiles corresponding to the crater-centric radials and sub-radials mapped in Figure 6. (The U.S. Geological Survey sections of Southern Maryland relevant to Figure 5 are shown at the top of Fig. 6).

In Figure 7A and B I have again superposed my calculated Chesapeake Bay crater ejecta blanket profile, again employing gently sloping bases to represent the uppermost coastal sediments on which the ejecta likely re-impacted. The values I selected for the slopes (~0.4 m/km) and the vertical offsets (~ 10 m below present sea level at r = 0) of the dashed lines were based on my operating assumption that the Bacons Castle formation may be ~10 to 20 m deep and that, like the upland deposits, it too may rest on gently a sloping base of older sediments.

In contrast with the gentle slope of the base of the upland deposits in Southern Maryland seen in Figure 5 and the extreme flatness of the upland deposits near Springfield and Alexandria, Va. (vide infra), we see in Figure 7 that in Virginia the upland gravels blanket V-shaped valleys and depressions with embankments in the direction of the crater having slopes ranging from 4 m/km on the York-Rappahannock peninsula (Fig. 7B) to 6.3 m/km just east of Richmond (Fig. 7C). This observation strongly suggests that in these cases the upland deposits were laid down directly on top of pre-existing topography having these particular slopes. This could easily have happened in the case of ballistic emplacement, whereas the prospects for fluvial deposition on these slopes become even more remote than they are in Southern Maryland (vide supra). And given the obvious armoring effects of a thick layer of coarse gravel, there is no *sedimentological* reason to assume that the upland deposits are geologically young.

Although Figure 7 may raise more questions than it answers, I hope it will be found useful to the reader wishing to test the various models. Below I speak of some of the things that I believe I see in this figure, together with some speculations as to they may mean.

Did the impact sculpt the base of the upland deposits? I believe it entirely possible and even probable that the jetting phase of an impact into shallow coastal waters would erase any pre-existing scarps imprinted on soft coastal-plain sediments as former shorelines. That is, I suggest that the answer to question (1) above is *probably yes*, but as yet I have no proof.

Do the Virginia upland deposits mimic those of Southern Maryland in a crater-centric way? The following passage from Poag (1997) suggests that the answer to question number (2) is *definitely yes*: "Above the [presumed] lower Miocene unit, coarser siliciclastic units of middle Miocene to Quaternary age [i.e., the upland deposits and Bacons Castle fm.] are widespread throughout southeastern Virginia. Outside the crater, *these units thicken gradually as they dip gently to the southeast. But where they cross the crater rim, the units abruptly thicken (moderately to slightly) and sag into the annular trough* (emphasis and bracketed words added)." The passage that I've italicized would certainly describe the Chesapeake Bay crater ejecta blanket *per se* in its last stages of existence ...and I suggest that Homo sapiens in fact appeared on the scene just in the nick of time to find it in exactly this condition!

Elsewhere Poag (1997) remarks: "The modern topography of the Chesapeake Bay region also appears to reflect the buried crater's influence (Peebles, 1984; Mixon, 1985; Mixon et al., 1989). For example, *the middle Pleistocene-upper Pleistocene contact approximates the position of the Suffolk scarp, a feature of 11-22 m relief, which parallels the western rim of the crater* (emphasis added)." This undisputed fact raises an interesting question: *Just how does a deeply buried 35.5-Ma crater exercise structural control over the geomorphography of coastal-plain sediments supposedly emplaced by rivers 28 m.y. later and subsequently subjected to the sea level transgressions and regressions of the Quaternary?*

Is the Bacons Castle formation Chesapeake Bay crater ejecta? Figure 7 gives no clue, except possibly for the seaward rising scarp seen in Figure 7D, which is unexpected for a coastal-plain terrace but is consistent with its being a vestige of an originally (vastly) higher crater rim. But that is pure speculation. On the other hand, the following information gleaned from Goodwin and Johnson (1970) is solid fact:

"*One strong contrast noted by Wentworth (1928) between the Brandywine gravels* [i.e., the Eastern Virginia upland deposits] *and lower, younger gravels* [i.e., the Bacons Castle formation] *was the absence of striated boulders and cobbles in the higher level gravels and the presence of such striated boulders and cobbles in the lower gravels."*

Apropos of that, I note that relatively rare striated rocks have been found in outcrops of Chicxulub-crater ejecta slightly inside of 5 crater radii (Ocampo et al., 1996; Pope et al., 1999; King and Petruny, 2003), although nothing is known of their radial distribution since no outcrops closer to the Chicxulub crater have been found. King et al. (1997) have argued that such striations result from hypervelocity interactions among clasts during excavation and ejection – consistent with the fact that ejected fragments "seldom interact with one another" in the ejecta curtain (Melosh, 1989). Therefore, since the excavation-flow crater ejecta is volumetrically so much greater than, and kinetically so much slower than the interference-zone ejecta (which I will argue are the origin of the coarse upland gravels), striated rocks should be an increasingly common occurrence in ejecta blankets as the crater rim is approached from great distance along a radial. Indeed, *the Bacons Castle formation, which is known to commonly contain striated cobbles and boulders* (Wentworth, 1928), *occupies a ~120° segment of the 1-to-2.5-crater-radii annulus that can be drawn about the center of the Chesapeake Bay structure* (Fig. 1).

In Early Oligocene times, the rivers would have found courses circumferential to the ejecta blanket. Presently, we see that the North Anna and Mattaponi Rivers are flowing perpendicular to the most southerly radial section reproduced in Figure 7B (small black squares). Likewise, for the James and Chickahominy Rivers in Figure 7C. And the same thing again for the Nottoway and Blackwater Rivers in Figure 7D. In fact, all six of these rivers are, at least at their respective intersections with one of these radials, flowing circumferentially about the Chesapeake Bay crater. Of course, most of these rivers do not flow very far circumferentially before turning back in the direction of the crater. The exception is the Nottoway River, which follows a 115-km radius for fully 60 km (Fig. 1)!

Note also that a V-shaped valley coincident with the mouth of the Chickahominy River extends all the way across the James-York peninsula (Fig. 7C) where it directly faces the mouth of the Mattaponi River on the York-Rappahannock peninsula (Fig. 7B). I propose that this groove could represent a paleochannel of the York River (defined as the confluence of the Pamunkey and Mattaponi Rivers) dating to the period when it had not yet breached the landward portion of the Chesapeake Bay crater rim (dashed arrow in Fig. 8A). If I am correct in this, then the western crater rim would have survived for at least 2 m.y. until the Early-Oligocene lowstand finally permitted the rivers to begin searching anew for their old channels.

Since all six of the rivers flowing circumferentially to the crater in Figure 7 happen to be moving in the counterclockwise sense, one might guess that this may be the result of the Corriolis force acting on southward flowing rivers. However, there may be an alternative explanation: an impenetrable radial barrier on the York-Rappahannock peninsula. Remembering that my calculated ejecta-blanket profile neglects dunes, hummocks, ridges, and rays, one must admit to the possibility of an extra-deep ray running up one of the peninsulas radial to the crater center. But is there any evidence for this? Well, yes, it appears that there is:

The Dragon Run watershed (Fig. 8A) is a raised-rim depression, which evidently has never during its existence allowed a major river to enter it from the north or west – a true oddity, given the canonical view that the U.S. Middle-Atlantic Coastal Plain gravels were all deposited by rivers within the past 10 m.y. (If rivers should have a way to self-construct barriers around depressions, then New Orleans could be secured without the Army Corps of Engineers!) It is seen in Figure 8A (Dragon Run Steering Committee, 2003) that the Dragon Run watershed is almost perfectly radial to the center of the Chesapeake Bay crater and its outline looks as though it could be decomposed into a sequence overlapping of ellipses – very similar to the string of secondary craters of the lunar crater Copernicus reproduced in Figure 8B.

Melosh (1989, p. 95) states: "Nearly every large crater or basin seems to have one or more especially prominent radial troughs extending from the rim out to nearly one crater diameter." "Originally thought to be tectonic features because of their often impressively straight walls, it is now believed that they are created by lines of coalescing secondary craters..." While I believe that this is precisely the correct explanation of the Dragon Run watershed, here is some truth in advertising: (1) by "large crater" Melosh (1989) is referring to those in the 100-to-200 km range on the Moon, Mars, and Mercury, and (2) the secondary crater chain associated with the 93-km-diameter crater Copernicus shown in Figure 8B begins at ~4 crater radii, not one. That said, even our knowledge of Martian craters excavated into sediments likely containing some liquid water at depth are insufficient guidance for what to expect for equivalently large impacts *into ~600 m of water-saturated sediments lying beneath ~150 m of sea water on the Earth*. The only way to know anything for sure about such a crater on the Earth is to discover one with a sufficiently well preserved ejecta blanket *...and study it!*

IN PLAIN SIGHT: SECONDARY CRATERS POCK THE MIDLOTHIAN UPLANDS!

Goodwin and Johnson (1970): "The flat, undissected upland surface of the Midlothian gravels is marked by numerous elliptical to subcircular depressions or basins. These basins were first reported by Johnson and Goodwin (1967). Over twenty such basins have been recognized in this area... The basins have formed on the Midlothian gravels and the basin sediments are immediately underlain by coarse gravels. The elevation of the basin floors is between [105 and 110 m] above sea level and a low ridge or rim surrounds the basins... The basins range in size from [~50 m] to more than [1 km] and where elliptical, their major axis trends from N 60° W to 80° W." N.B. The direction to the Chesapeake Bay crater center from these sites is N 78° W.

Goodwin and Johnson (1970) further mentioned that the 1.5-to-5-m-high rims of the Midlothian upland basins are "...most commonly best developed on the south and east sides of the basins," i.e., the crater side. N.B., Melosh (1989) states that "Secondary craters are typically asymmetric, having steeper slopes on the side closest to the primary crater."

Goodwin and Johnson (1970) led a field trip to one of these “Carolina Bays” where a drainage ditch had not long before been dug from the interior through the rim. Quoting from their guide book:

“In general the section shows that the rim is underlain dominantly by sand containing some pebbles but only very minor amounts of clay. Rarely cobbles or pebbles occur within the sand. A pronounced decrease in sand occurs from the rim toward the bay's interior and massive, brownish-gray, silty clay with a few scattered quartz pebbles becomes the dominant sediment. *This clay is in direct contact with the underlying gravels* (emphasis added). The lateral transition from sand to clay is gradational but occurs within a distance of less than [60 m] from the rim.”

Elsewhere Goodwin and Johnson (1970) remark that “A reconnaissance survey of the surrounding region revealed that these basins are restricted to the upland surfaces of the Midlothian gravels. No evidence was found of similar basins in Piedmont areas without a gravel veneer or in areas which have been severely dissected by erosion. Apparently the development of the bays is in no way related to the Piedmont bedrock which underlies the Midlothian gravels. Some of the bays have formed on gravels overlying granite, and some have developed on gravels overlying Triassic shale, sandstone, and coal measures.”

Stop #4 on Goodwin and Johnson’s (1970) field trip was a road cut exposing ~9 m of upland gravels “which overlie a saprolite of the Petersburg granite in a nonconformity.” I reproduce their sketch of this exposure as Figure 9A. Most striking are the facts that the granite is first contacted by a “thin cobble zone varying from [~8 cm to 30 cm]” overlain by “a pronounced [1.3 m] thick clay zone,” and *this bilayer conforms perfectly to the concave-upward curvature of the eastward-sloping bedrock contact!* Immediately above this bilayer, “...discoidal cobbles are oriented parallel to the contact but higher in the gravel sequence the cobbles have their maximum dimensions nearly horizontal...” (Note how carefully this revealing feature is drawn in Fig. 9A.) And, just as for the upland deposits of Southern Maryland, most of the cobbles and pebbles in this ~8-m-thick stratum were reported to comprise “massive quartz, vein quartz, quartzite, or quartz sandstone...” and are “matrix supported” by “a medium-to-coarse sand matrix.”

In Figure 9B I provide a profile of part of the Midlothian gravels based on a segment of the field trip of Goodwin and Johnson (1970) that followed U.S. Route 60 West through the tiny town of Midlothian (with all distances measured from the crater center, rather from a car odometer). This profile (hollow squares) can be compared with the southernmost radial section of Figure 7C, repeated here as the small black squares. On a westerly stretch of Route 60, the road is perfectly parallel to a crater radius; elsewhere it is slightly sub-radial. Where the two profiles overlap, Route 60 is ~8 km north of the linear, though slightly sub-radial section represented by the black squares.

We see that the Midlothian gravels rest at much higher elevations than any of the upland deposits designated in Figure 7. But they are nowhere close to being the champion. According to Schlee (1957) that honor belongs to Tysons Corners, Va., at 158 m! I find it impossible to believe that rivers could have laid down such heavy gravel loads at any one of these elevations, *much less all of them*, without leaving the slightest trail of boulders leading back to their source.

My interpretations: *The Midlothian "Carolina Bays," with the major axes of their ovate rims nearly radial to the Chesapeake Bay crater are surely results of secondary impacts*. But one may well wonder how this could be possible if the upland deposits themselves are assumed to comprise *primary* ejecta. Well, I argue that the *gravel member* of the upland deposits comprises *interference-zone* ejecta and, as such, these materials would have been ejected early and fast and thus might be expected to have been the first ejecta to re-impact. However, whereas the (coarse) gravel member comprises the base of the upland deposits of Southern Maryland and Northern Virginia, the Midlothian road-cut site of Figure 9A reveals that the coarse gravels *conformably* overlie a ~1.5-m-thick gravel-and-clay bilayer. Clearly, if the gravels are impactoclastic, then this basal bilayer must also be crater ejecta – ejecta that preceded the interference-zone materials to this site. With total confidence that rivers couldn't have emplaced *any* of these deposits, I therefore conclude that this basal gravel-clay bilayer must be a *jetting-phase deposit*.

Now returning to what would have created those "Carolina Bays," it should be noted that the very last, and slowest moving, materials to be ejected from any impact crater are the dregs of the excavation-flow ejecta (Melosh, 1989) – which should include a large melt component. For a crater the size of the Chesapeake Bay structure, these last-out ejecta would have been launched ~90 seconds later than the last interference-zone ejecta (Melosh, 1989, eq. 5.5.2) – allowing plenty of time for the pre-arriving jetting and interference-zone ejecta to have settled.

Still, Goodwin and Johnson (1970) report nothing that could be interpreted as the remains of the large ejecta fragments that would have been required to excavate secondary craters of the order of 1 km in diameter ...or do they?

Kastner et al. (1984) were the first to make the case that many impact glasses subjected to meteoric or marine weathering are likely to be converted to smectite. Therefore, I envision a coherent blob of largely-molten excavation-flow ejecta arriving at one of those Midlothian sites with just the right kinetic energy to excavate the upper sand member of the until-then-flat upland deposit all the way down to its lower, more resistant, gravel member. Whereupon, I see this blob flattening itself against the top of the gravel and rapidly quenching to solid glass. Then, over the eons this glassy lens would have been digenetically altered by rain water inevitably trapped in the "Carolina Bay" of its own creation. So, if my vision has any correspondence to reality, the remains of the original blob now rest inconspicuously as the “massive, brownish-gray, silty clay [smectite?] ...in direct contact with the underlying gravels” (Goodwin and Johnson, 1970).

ANOTHER CANDIDATE JETTING-PHASE DEPOSIT: A NORTHERN-VIRGINIA DIAMICTON YIELDING REMARKABLE CLASTS

Geologic setting. Springfield, Va., is located just south of the Washington Capital Beltway at its southerly junction with Interstate 95. It rests on a remarkably flat clay terrace, which remains within ~61-64 m above present sea level all of the way to our former home, about 10 km east-southeastward and 2 km shy of the Potomac River. According to a 1960s-vintage U.S. Geological Survey map, these upland flats used to be everywhere dotted with gravel pits. If my recollections are correct, the cobble-size distributions I’ve seen at some of these sites were commonly similar to that of Figure 3B, though frequently including a minor boulder component.

Near our home was the spring-plus-shopping-center-runoff-fed headwaters of the Paul Spring Branch of Little Hunting Creek, which empties into the Potomac River at Mount Vernon. Before turning south, Paul Spring Branch flows southeasterly through a 33-m-deep valley in a 3-m-wide quartzite-cobble-lined channel with a gradient of ~1 m/km. Our neighborhood of Hollin Hills was innovatively developed on what was widely considered to have been undevelopable terrain, thanks in part to the developer’s decision to preserve the existing natural drainage gullies as parklands. Brickelmaier Park slopes downward at ~66 m/km northward to reach Paul Spring Branch. One result of this terrific gradient has been the incision of a gorge ~3 m deep part way downslope. But this occasionally vigorous rivulet levels out near a point (38°45'37.2" N, 77° 3'56.2" W) where it has managed to cut only ~30 cm into a resistant layer of whitish-gray “clay” supporting a random assemblage of mostly quartzose pebbles and cobbles (Fig. 10A). Hence forward, I will refer to this stratum as “the Hollin Hills diamicton.” Microscopic examinations and X-ray diffraction eventually revealed that what I had initially thought to be a “clay” matrix actually comprises nearly pure quartz silt. Griscom et al. (2003a) show a photomicrograph of a quartz grain collected from this stratum which displays at least 3 intersecting sets of planar deformation features.

During my aperiodic visits to this site, I began to notice that the rivulet was in the process of exposing several very unusual clasts. Figures 10B, C, D, and E display a subset of my finds, down-selected to those with stories I think I understand well enough to relate in useful detail.

The autobrecciated chalk ball. I discovered this clast while it was still in the process of being exhumed directly from the bed of the rivulet. It is shown in the upper part of Figure 10B in its as-recovered condition. The friable orange-brown crust on the bottom of the object (spontaneously crumbling after desiccation) may have originally been present on its presently “bald” top surface as well – in which case it might have been stripped by the flowing water before I dug it out. The clasts internal to the illustrated sawed slab appear slightly darker than the pale yellowish-white matrix (a difference that I digitally enhanced in the figure). However, this effect appears to be due to higher specular reflection of light from the clasts due owing to their lower porosity, thus causing the matrix to appear lighter when the sawed face is viewed at a small angle to the illuminating light. Conversely, when illuminated and viewed at complementary acute angles, reflection from the matrix becomes equally specular and no color difference is perceived. Inspected with a hand lens, the matrix appears fossil free, whereas the a few of the clasts display one or two split ovoid objects ~1 mm, which to my amateur’s eye best match expectation for benthic forams cut parallel to semi-major axes. (I would be happy to make samples of this clast available to any paleontologist willing to render an expert judgment.)

This rock certainly could not have been emplaced by rivers because (1) it wouldn’t have survived fluvial transport for as far as 100 m and (2) there are no chalk outcrops upriver to the north or west, nor would any be expected anywhere in the Appalachians. On the other hand, with the eustatic sea level having been as high as 300 m during much of the Cretaceous and remaining as high a 140 m throughout the Paleogene up to the time of the Chesapeake Bay impact (Hallam, 1984), some pelagic limestones were almost certain to have been present in the target area. Therefore, *the clast of Figure 10B is surely a Chesapeake Bay crater impactite*.

The palagonite “cinder.” The 8.5-cm-tall object illustrated in Figure 10C matches the Glossary of Geology definition of palagonite. However, I strongly doubt that it is of volcanic origin given its composition (vide infra) and its incorporation of quartz clasts in sizes up to 1 cm. In Figure 10C, one of the larger quartz-crystal inclusions protrudes from the left hand side of the object, and another one is seen in the interior of the saw-cut slab (their locations are indicated by black outlines). This object’s grooved and pitted surface was densely packed with the same whitish-gray quartz silt in which it was found embedded. The coherency of this silt matrix cannot be overstated; even with a strong stream of water from a garden hose I failed to completely dislodge it from some of the pits. More significantly, the slice sawed from the bottom revealed that clods of the same quartz silt are encased in the *interior* of this “cinder.”

The only crystalline minerals identified by X-ray diffraction of a powdered sample of the orange-brown material selected to be as free as possible of quartz inclusions were goethite ...and quartz. Both sets of diffraction lines were relatively weak, suggesting that much of the orange-brown material may be X-ray amorphous. Careful inspection of the sawed surface reveals that the object solidified from a sol comprising at least two immiscible melts. The darker orange-brown material making up the continuous phase is substantially harder than the captive yellow-brown material.

I interpret this object as a blob of impactoclastic glass – possible melted clays – which accreted quartz during its jetting-launched flight toward Hollin Hills, 190 km distant from the crater center.

The melt-encrusted greenstone fragment. At first I believed this object to be another chalk ball. It had a similar "toasted" orange-brown exterior surface, which was underlain by what seemed to be the same chalk when I picked through the brittle outer rind and found a friable whitish material inside, which fizzed (briefly) when treated with HCl. However, when I finally got around to sawing it in half, a wholly different story emerged (Fig. 10D). The core rock turns out to be dense, hard, massive (overlooking the fine network of well cemented fractures), and crystalloblastic. My amateur interpretation is that it is greenstone, a common local basement rock – which in the present case (having become part of the jetting-phase ejecta) must have been transported prior to 35.5 Ma from the Blue Ridge to a location near the top of the ~500 m of sediments in the target area of the Chesapeake Bay impactor.

However, there is another interesting story yet to be told in the five-layer rind on the outside of this rock. In general, I interpret this rind to be the end result a glass layer created by an external temperature rise of "tens of thousands of degrees (Melosh, 1989)" during the jetting phase of the impact. Its present-day multi-layer complexity sure relates at least in part to diagenetic alterations, although the innermost layer could be a relict zone of partial melting.

We have seen that this surface-melted greenstone object and the chalk ball – both found in the Hollin Hills diamicton – both display "toasted" orange-brown crusts ...as did also a Chicxulub-impact-related carbonate spheroid from southern Mexico that I have studied (Griscom et al., 2003b). This coloration is certainly due to iron oxide, but how did it come to appear on the *surfaces* of these objects? Extrapolation of existing binary phase diagrams for the system Fe-Fe_2O_3 (Schneider, 1969) from their presently established high temperature of 1,600° C suggests that liquid oxide and liquid metal could coexist at temperatures above ~2,300° C, even in air at one-atmosphere pressure. The reasons why smelted iron would transport to the surface under those conditions are undoubtedly complex and dependent on ambient conditions thus far unknown. However, it is clear that any such metallic component would have re-oxidized soon after reimpact.

The rock of Figure 10D is actually surfaced by a brittle bilayer comprising a pale whitish orange-brown outer surface and slightly thicker (~0.5 mm) black layer below it. The black layer is unlikely to be magnetite, since it didn't respond to a magnet. The volume of this black material seems rather large compared to the small amount of light-colored (low-iron) melted rock represented by the existing rind. However, it seems likely that a large amount of the original rock actually evaporated in the jet that launched it. If so, given that iron has a lower volatility than most other constituent elements, iron would have naturally concentrated at or near the surface. Thus, a future determination of the compositions of both the core rock and the black layer might give an idea of how much of this rock was vaporized under differing model jetting scenarios...

Interestingly, this rock was shaped vaguely like an Apollo command module, with a gently convex “front” surface and a tent-shaped “superstructure.” And, lo, the (now crumbling) iron-oxide surface exhibited flow textures leading from the front surface around the periphery and starting to climb part of the way up the side of the “module.”

The doubly-fractured quartzite cobble emplaced in one piece. I found the three objects of Figure 10E partially exposed in the Hollin Hills diamicton *tightly meshed together*. So without a doubt, this quartzite cobble was subjected to three planar fractures *before deposition*, and yet it was deposited with three of its presumed four original parts still joined. It is probable that these three fragments resulted from a cobble deposited with a pair of partial fractures which completely split apart later due to freeze thaw cycles in a moist environment, since nothing else could have affected them while they remained entombed in their massive quartz-silt sarcophagus. This finding strongly supports the notion that the fractured rocks of Figure 3 were also “fatally,” if not completely, fractured *prior to deposition* – implying that their current condition was neither the work of rivers nor of any Indians or colonial farmers who might have dug them up.

On the origin of the Hollin Hills diamicton. I propose that the whitish-gray diamict stratum of Figure 10A comprises *jetting-phase ejecta* possibly correlative with the basal pebble-and-“clay” bilayer in the Midlothian uplands (vide supra). However, the location of this stratum is topographically lower by ~30 m than the base of the local upland deposits, which overlie what in this context I will call the “Springfield-Hollin Hills plateau.” Therefore, the temporal separation in the stratigraphic column of the Hollin Hills diamicton from Schlee’s (1957) upland deposits is yet to be established. That is, it cannot be ruled out that the former was deposited long before the latter and that cobbles from the latter now immediately overly a recently exhumed ancient diamicton due to mass erosion of younger deposits on the heights.

But there is another possibility – and I think the correct one – namely, that the Hollin Hills diamicton is indeed jetting-phase ejecta of the Chesapeake Bay crater, which is fortuitously preserved in its present location by virtue of this debris flow having reached zero velocity at just the right moment to settle into a well-shielded topographic low *existing at the time of the impact* (currently a 66-m/km reverse slope relative to the direction of ejecta arriving from the Chesapeake Bay impact). This “good fortune” could have prevented it from being ripped up and incorporated into the later-arriving, interference-zone-launched upland gravels. Indeed, the relative rarity of such shielded locations could be the reason why Schlee (1957) did not report similar whitish-gray-quartz silt strata at the base of the upland deposits in any of the 98 sites he sampled.

So why didn't the "rip up" effect happen at the Midlothian sites? I think the answer is that, in that case, the ejecta landed on solid bedrock of the Piedmont, so there was no soft substrate to rip up. Indeed, I propose that the concave upward unconformity of Figure 9A was originally scoured out of this Petersburg granite outcrop by the first arriving, highest energy materials jetted in that direction during the contact phase of the Chesapeake Bay impact. I want to emphasize that, while the Hollin Hills diamicton is technically not lithified, it is definitely *not soft*! I think this ultra-dense-packed stratum could easily have withstood being caught "between a rock and a hard place" at Midlothian Stop #4 (Fig. 9A). Obviously, this notion should be a sufficient justification for careful studies of the Midlothian "pebble-and-clay" bilayer to find out whether the Hollin Hills diamicton is a one-off fluke ...or is the stratotype of a vast (and highly information laden!) impactoclastic formation. N.B. Melosh (1989) states that the jetted mass should comprise "comparable quantities of both target and projectile material," so *the probability of finding intact fragments of the impactor should be large if my interpretation is correct*.

A SPALL OF LANGLEY GRANITE IN PLAIN SIGHT?

For the first 20 years or so that we lived in Hollin Hills I would sometimes balance myself on a half-buried dome-shaped rock while contemplating Nature from the top of our driveway. When eventually this rock had to be removed to make room for an herb garden, I placed it above ground as an ornamental. Then one day my eye caught the streaks on its side. Going for closer look, I happened to turn the domed side down, revealing for the first time that the other side was nearly flat and *had a raised rim around its circumference* and that the streaks ran from the domed side "backwards" in wind-tunnel-like fashion (Fig. 11A). Being familiar with button tektites (e.g., Glass, 1984), it took me only seconds to conclude that this rock must be a meteorite. But its color was all wrong for a chondrite; in fact, it looked to me like granite. I knew then that I possibly had in my lap a meteorite from a terrestrial planet (Gladman et al., 1996), maybe even our own! I became so excited that I spread the story far and wide to scientists I thought surely would be professionally interested. But, perhaps not surprisingly (Mitroff, 1975; Griscom, 1994), my story was universally met by disinterest or outright disbelief, even after some looked closely at my objects. So for the past 9 years I have been the sole researcher of this 27-kg rock. However, no one has been paying me to do this, and I've had much else on my platter ...so I must make these excuses for the paucity of results I have to report below.

The anterior (domed) surface differs from the rest of the rock by showing a faint orange coloration reminiscent of the orange or orange-brown outer crusts on the objects found in the Hollin Hills diamicton – and probably for the same reason. From inspection with a hand lens, it appeared to me that the sub-parallel streaks on its sides might be melted biotite. So, I tried and succeeded in replicating such an effect by applying a well-tuned H_2-O_2 torch (2,800° C) tangentially to a freshly fractured sample of biotite granite. By sheer coincidence, it turns out that the temperature of the Apollo command-module heat shields rose to 2,800° C on reentry (Allday, 2000). And, if any further proof of my conclusions were needed, Deer et al. (1966) state that, while biotite is often the first granitic mineral to crystallize in magmas at depth, it also has the lowest melting point at atmospheric pressure.

I had a thin section prepared from a chip taken from the raised lip on the nearly flat posterior surface of the object of Figure 11A. Figure 11B shows an eruption of amorphous material (the colors and "flow" textures were insensitive to stage rotation when viewed between crossed polarizers) protruding just beyond the outer surface of the rock. By process of elimination, this must be melted biotite quenched to an amorphous state.

I therefore propose that this extraordinary granitic rock was blasted deeply into space – perhaps at about 10 km/s and an elevation angle of about 80° to the northeast (see, Alvarez, 1996) from the interference zone of the Chesapeake Bay impact (Fig. 4) and that, upon reentering the Earth's atmosphere, it achieved an aerodynamically stable attitude with its quasi-dome-shaped surface forward. During its descent toward our front-yard-to-be, aerodynamic-shock-induced melting of the biotite took place in a near-surface region of the anterior dome. Some of this melt was extruded and aerodynamically-driven backward to form the observed streamlines following their respective paths of least resistance back up the sides. More importantly, thicker molten-biotite-solid-crystallite slurries (clearly observed in one small patch of "naked" feldspar and quartz crystallites) also moved backward, albeit more slowly, thus accounting for the observed raised lip around the posterior surface (Fig. 11A) – as well as frozen flow fronts on the anterior dome perceptible under grazing illumination.

I would love to permit a front-to-back core to be drilled through the center of this rock and for documented samples to be cut from it for examination by any interested researchers. (I propose to hold the rest of this object for possible future display by the Smithsonian Institution.) Indeed, I would like to be a part of a paleothermometry study based on electron spin resonance of paramagnetic E ′ center defects induced in α quartz by naturally occurring radioactive impurities, given that these defects are known to anneal out at around 400° C (e.g., Jani et al., 1983).

Of course, the *prime* objective of most other studies would be to determine if the object of Figure 11 is petrologically related to the Langley Granite (Horton et al., 2005). How exciting that would be if the answer should turn out to be, *yes*!

THE ANSWER, MY FRIEND, IS BLOWIN' IN THE WIND

In fairness, I recognize that back in the time of Hack (1955) and Schlee (1957) uniformitarian geology had long since ascended to the status of a quasi-religion. Loren Eiseley (1960) wrote eloquently on how Biblical catastrophism had been justly supplanted by the concept of slow processes operating in deep time, thanks to the genius and persuasion of Hutton and Lyell. Conceivably, Eiseley (1960) caught a whiff of what was yet to come as he penned his remark on Lyell's "insist[ence] upon his principle of preoccupancy" ...which if taken to the limit would have required the dinosaurs to still be here instead of us. I'm sorry that Eiseley didn't live long enough to exhilarate in the catagory-4 wind of catastrophe by bolide impact (Alvarez et al., 1980) ...which back then was but a zephyr in what Eiseley (1961) repeatedly referred to as "the wind of the oncoming future." We who presently live in that future cannot be the least bit smug in our new knowledge, given that the onrushing future will soon enough invalidate a fair share of our own philosophies. Still, as scientists we are now obliged to use the newest revelations to mop up the errors that inevitably infest the geological literature traceable to the absence of impact-stratigraphy chapters in the text books of the recent past. Apropos of that, I lament too that Thomas Mutch's life was cut so short, since I owe my appreciation of stratigraphy to Mutch (1970).

Therefore, it is with the deepest respect for all of the geologists who in the past held firm to their "uniformitarian faith" that I here make bold to posit that the well-rounded (though oft fractured) pebbles and cobbles of Devonian-age quartzites ubiquitous to the gravel member of the upland deposits rested peacefully in the target area of the Chesapeake Bay impactor one fateful day 35.5 m.y. ago. More specifically, I argue that these upland gravels must have been present in the *interference zone* (Melosh, 1989) of the Chesapeake Bay impactor, else they would have been thoroughly pulverized to Grady-Kipp fragments by the downwardly moving tensile waves (Melosh, 1989). Indeed, I propose that the gravels that experienced *that* fate were sucked into the streamlines of the excavation-flow ejecta (including a lot of erosion resistant glass) that were re-deposited as the Bacons Castle formation in an annulus closer to the crater. I further propose that the upland gravels were size-sorted in ballistic flight by atmospheric drag (Schultz and Gault, 1979; Schultz, 1992), thus resulting in the radial gravel-size gradient so meticulously exposed by Schlee (1957). Indeed, Pope et al. (1999) have already invoked atmospheric size sorting to explain features of the Chicxulub impact deposits in Belize.

Answers to some protestations

There is no evidence of Devonian quartzite in the target zone

It has been argued that my model is contradicted by the fact that drilling has not revealed the slightest sign of Devonian quartzites in the target area. My response follows:

(1) I am arguing that the gravel member of the upland deposits derives from the effective interference zone of the Chesapeake Bay impact (Melosh, 1989, p. 73), that is, a volume ~12 km in radius and ~600 m in depth (minus the central footprint of the impactor which, for example, might have been a 6-km-diameter comet moving at 30 km/s and striking at a 45° angle). It follows therefore that the immediately contiguous materials out to a radius of 45 km were completely removed by the impact and thus all drilling into lithologies pre-existing the impact is pushed outward at least 33 km from the putative location of the materials that the doubters would like to have verified. If, for example, a prime source of the quartzite gravels should have been the Jurassic wedge cartooned in Fig. 4 (Koeberl et al., 1996; Poag 1997), then only ocean-side drilling down to bedrock will ever reveal them, and this has not been done. Indeed, *the sediments immediately surrounding the Chesapeake Bay crater in its first moments of existence are likely to have transported some distance seaward during the past 35.5 m.y.*

(2) The sorting-coefficient data of Schlee (1955) for the upland gravels, when compared with the data of Emery (1955), show the upland gravels to be a near perfect match for an alluvial fan (Fig. 12). But alluvial fans develop at the base of the mountains they derive from, whereas the nearest significant sources of Devonian (and Silurian) quartzites are the anticlines capping the Appalachians *to the west of the Blue Ridge* (see Fig. 1) and the synclinorium of Massanutten Mountain in the intervening Shenandoah Valley. Since I believe it proven beyond the shadow of a doubt that rivers could not have dragged 50 to 100 km^3 of quartzite silts, sands, and coarse gravels from the Shenandoah Valley out to the Coastal Plain in the last 10 m.y., it seems to be a more fruitful line of inquiry to now search for reasons why these materials were actually present in the target zone. To that end, I propose that at the beginning of the Appalachian orogeny around 250 Ma, the very same Paleozoic quartzite formations that are now truncated at Massanutten Mountain actually extended eastward of what is now the Blue Ridge. In this scenario, those putative eastward extensions of these Silurian and Devonian quartzites would have been eroded to alluvium in Mesozoic times. This alluvium would at some point have needed to be transported still farther eastward by rivers in order to eventually become part of the presently ~600-m-thick, *non-marine* "seaward-thickening wedge of mainly lower Cretaceous to upper Eocene, poorly lithified, and mainly siliciclastic sedimentary rocks" (Koeberl et al., 1996; Poag et al., 2004) present in the target zone of the Chesapeake Bay crater impactor (Fig. 4).

It is relevant to mention that Attal and Lavé (2006) have established quartzite to be the most abrasion-resistant Himalayan rock in the Marsyandi River in Nepal, both by field observations and experimental measurements in a circular flume. Their experimental mass loss per kilometer for quartzite (resulting from an emulation of the Marsyandi River during annual peak discharge across the Lesser Himalaya) of 0.15 % per km translates to a line that plots between the pair of curves for the Rhine and Mur Rivers illustrated in Figure 2B. This is more bad news for Hack's (1955) model (if any more were needed). Additionally, Attal and Lavé (2006) found Himalayan granite and schists to abrade at rates ~3 and ~100 times faster than quartzite, respectively. This is good news for my implicit assumption that the largest clasts eroded from the "Proto-Appalachians-East" that were likely to have been transported the farthest eastward by rivers would have been quartzite. And, if these Proto-Appalachians were as high as the Himalayas, landslides might have delivered quartzite directly into the rivers as debris in diamict sizes (Attal and Lavé, 2006: dashed curves in Fig. 12).

In the latter regard, it is worth emphasizing that Schlee's (1957) data for the upland deposits shown in Figure 12 pertain to *the gravel member only*, whereas the overlying loam member is comparable in thickness and 90% comprised of quartz silt (Hack, 1955). Thus, if the gravel and loam members – which I argue were atmospherically separated in ballistic flight – were to be reunited, their combined particle-size distribution might well recapitulate one of Attal and Lavé's (2006) quartzite landslide diamictons.

Established stratigraphic columns are not subject to reshuffling

Another objection I heard at the Penrose Conference was that, whereas fossil biozones may occasionally need to be re-dated when more accurate constraints are imposed (usually by radiometric dating of volcanic-ash or impact layers; e.g., Smit, 1999), it is inconceivable that any accepted sequence of biozones in a long-established stratigraphic column should ever be compelled to exchange positions across a time plane. I will give my answer to that below.

But first the background: Figure 5 is an adaptation of a U.S. Geological Survey section of the Potomac-Patuxent peninsula extending north-northwestward, passing just eastward of the District of Columbia. The upland deposits are clearly shown as resting on their scarp-free base, gently sloping in the general direction of the Chesapeake Bay crater. The Calvert formation (darkest band to the right) and the Choptank and St. Marys formations that successively overlie it comprise the base of the upland deposits. Therefore the upland deposits must be younger than those underlying formations, thus contradicting my hypothesis if one accepts their historical dating as Miocene or more recent. However, "the plot thickens" when one begins to contemplate that there is also an *in-the-crater Calvert formation* (e.g., Poag, 1997).

In the following discussion, I will refer informally to the extensive *upland* Calvert formation as "Calvert I" and to the relatively tiny *in-crater* Calvert formation as "Calvert II." These two Calverts are diachronous in anyone's model, not just mine. In the currently accepted stratigraphic column they are diachronous across a time plane including hiatuses summing to as much as 28 m.y. (Poag, 1977). My operating premise is that *Calvert II was deposited soon after the impact* and owes its preservation to the fact that the crater floor has been continuously subsiding due to compaction of the Exmore breccias under the weight of post-impact sedimentation. Poag (1997): "The evidence suggests that the thick breccia lens inside the crater continued to compact and subside more rapidly than deposits outside the crater... and may continue even today." Thus, in my view, sediments entering the crater in the first few million years of its existence are sure to still be there, whereas sediments deposited more recently would have been subject to removal and replacement by ever younger sediments subsequent to the eventual breach of the crater rim by the many rivers that presently converge there. In particular, *a "Calvert II" envisioned as being 28 m.y. younger than the crater should not have survived fluvial erosion during the lowstands of the Quaternary.*

In my model, Calverts I and II are diachronous across a time plane of only about 2 million years. The microfossils found in both Calverts are neritic species. Therefore, in anyone's model Calvert I must have been deposited during sea level transgressions equal to or greater than ~79 m (the highest point in Calvert I that I've been able to locate), whereas Calvert II would have been deposited during an exceptional regression, almost certainly the one that occurred around the Eocene-Oligocene boundary. Hallam (1984) has estimated eustatic sea levels of about 155 m during most of the Late Eocene, vis-à-vis a comparatively brief peak of about 65 m around 10 Ma. Any doubts that the sea levels were that much lower in the Miocene than in the Late Eocene should be dispelled by recent oxygen isotope data (e.g., Zachos et al., 2001), which prove that ice-caps were small to nonexistent before the Chesapeake Bay impact, whereas a substantial Antarctic cap has been continuously in place ever since.

Summary

In addition to supplanting the untenable fluvial model for the upland deposits, my new model for the stratigraphic column of the U.S. Middle-Atlantic Coastal Plain is (1) in accord with the physics and geology of impact cratering and (2) matches far better the emerging picture of Cenozoic eustatic sea level variations than does the currently accepted one. Moreover, the diachronism that I invoke – crossing a time plane spanning a likely-hiatus-free 2 m.y. – is far less problematic on its face than the hiatus-riddled column historically predicated on a Miocene-age "Calvert I."

CONCLUSIONS RELATING TO THE LATE EOCENE EARTH

The consequences of this study for non-impact geologists whose prime concern is the Late Eocene Earth can be summarized in this way: If the essence of my thesis should turn out to be correct, then those concerned with neritic biota of the Late Eocene would then need to bring "under their tent" a whole lot of additional species, that is, those species that I argue have been misdated as Miocene. If this should be found out to be the case, then a number of vexing anomalies in the fossil record would almost certainly be cleared up in the process. Moreover, comparison of the species present in "Calvert II," which I propose to have been deposited during the Early Oligocene lowstand, with those present within "Calvert I," which I argue was deposited sometime during the Mid Eocene highstand, would finally give a the correct picture of exactly which extinctions may or may not have resulted from the Late Eocene impacts. Finally, studies of the survival rates of the "Calvert-II" species would provide an excellent test of Hallam's (1984) proposition that lowstands should bring about extinctions of neritic species endemic to highstand ecosystems.

In any event, it is worth noting that Pälike et al. (2006) have computer modeled the paleoclimatic record of the entire Oligocene preserved in the Pacific Ocean Site 1218 Ocean Drilling Project core, showing among other things that the Oi-1 glaciation can have resulted from long-term orbital-forcing cycles *only in combination with some unspecified terrestrial trigger, with this trigger indicated by their simulations to have occurred between 35.2 and 36.3 Ma* (see their Fig. 3). Since these dates bracket the radiometric ages of the Chesapeake Bay and Popigai impacts, they support the proposition of Fawcett and Boslough (2002) that this climate-change trigger – instead of being some cryptically extreme removal of CO_2 from the atmosphere – could have been the winter-hemisphere cooling resulting from the shadow of an equatorial debris ring thrown up by one or the other of the two major Late Eocene impact events.

REFERENCES CITED

Allday, J., 2000, Apollo in Perspective: Spaceflight Then and Now: CRC Press, p. 320.

Attal, M., and Lavé, J., 2006, Changes of bedload charactistics along the Marsyandi River (central Nepal): implications for understanding hillslope sediment supply, sediment load evolution along fluvial networks, and denudation in active orogenic belts: in Willett, S.D., Hovius, N., Brandon, M.T., and Fisher, D.M., eds., Tectonics, Climate, and Landscape Evolution: Geological Society of America Special Paper 398, Penrose Conference Series, p. 143-171.

Alvarez, W., 1996, Trajectories of ballistic ejecta from the Chicxulub crater: in Ryder, G., Fastovsky, D., and Gartner, S., eds., The Cretaceous-Tertiary Event and Other Catastrophes in Earth History: Boulder, Colorado, Geological Society of America Special Paper 307, p. 141-150.

Alvarez. L.W., Alvarez, W., Asaro, F., Michel, H.V., 1980, Extraterrestrial cause for the Cretaceous-Tertiary extinction: Science, v. 208, p. 1095-1108.

Baker, V.R., 2002, The study of superfloods: Science, v. 295, p. 2379-2380.

Bradt, R.C., and Tressler, R.E., 1994, Fractography of Glass: New York, Springer, 310 p.

Chesterman, C.W., and Loew, K.E., 1978, The Audubon Society Field Guide to North American Rocks and Minerals: Tokyo, Alfred A. Knopf, 850 p.

Deer, W.A., Howie, R.A., and Zussman, J., 1966, An Introduction to Rock Forming Minerals: New York, Longman,.

Dragon Run Steering Committee, 2003, The state of Dragon Run: Status of natural resources: Virginia Department of Conservation and Recreation (Report, December, 2003), 34 p..

Eiseley, L., 1960, The Firmament of Time: (reprinted 1999) Lincoln, University of Nebraska Press, 183 p.

Eiseley, L., 1961, The Man Who Saw Through Time: New York, Charles Scribner's Sons, 125 p.

Emery, K.O., 1955, Grain size of marine beach gravels: Journal of Geology, v. 63, p. 39-49.

Fawcett, P.J., and Boslough, B.E., 2002, Climatic effects of an impact-induced equatorial debris ring: Journal of Geophysical Research, v. 107, p. 10.1029/2001JD001230.

Frye, K., 1986, Roadside Geology of Virginia: Mountain Press, Missoula, Montana, 278 p..

Gladman, B.J., Burns, J.A., Duncan, M., Lee, P., and Levison, H.F., 1996, The exchange if impact ejecta between terrestrial planets: Science, v. 271, p. 1387-1392.

Glass, B.P., 1984, Tektites: Journal of Non-Crystalline Solids, v. 67, p. 333-344.

Gohn, G.S., Koeberl, C., Miller K.G., Riemold W.U., Browning, J.V., Cockell, C.S., Horton, J.W. Jr., Kenkman, T., Kulpecz, A.A., Powars, D.S., Sanford, W.E., Voytek, M.A., Deep drilling into the Chesapeake Bay impact structure: Science v. 320, p. 1740-1745.

Goodwin, B.K., and Johnson, G.H., 1970, Geology of the upland gravels near Midlothian, Virginia: Guidebook, 11th Annual Field Conference of the Atlantic Coastal Plain Geological Association.

Griscom, D.L., 1994, From E´ centers to the 'characteristic' resonance: a colleague's reminiscences on the scientific odyssey of Robert A. Weeks: Journal of Non-Crystalline Solids, v. 179, p. 22-38.

Griscom, D.L., 1999, Possible Chesapeake-Bay-crater ejecta found at 4.5 crater radii [abs]: Eos (Transactions, American Geophysical Union), v. 80, p. S211.

Griscom, D.L., 2001, Possible identification of proximal ejecta from the Upper-Eocene Chesapeake Bay crater: ESF-IMPACT Workshop, 6th, Granada, Spain, Abstracts, p.35.

Griscom, D.L., 2002, Is Washington, DC, built on ejecta deposits of the Upper-Eocene Chesapeake Bay crater?: ESF-IMPACT Workshop, 9th, Prague, Czechoslovakia, Abstracts, p. 23.

Griscom, D.L., 2007, Chesapeake Bay crater ejecta are alive and well and living in Maryland, Virginia, and the District of Columbia ...disguised as ~5,000 km^2 of "upland deposits": Wetumpka Field Forum 2007 Guidebook and Abstracts. p. 85-87

Griscom, D.L., Akiyoshi, A., Homae, T., Kondo, K., Yamanaka, C., Ueno, T., Ikeya, M., Affatigato, M., Schue, A., 2003a, Fossil natural glasses composed of ferric oxyhydroxides: Impactites of the 35.5 million year old Chesapeake Bay crater: Journal of Non-Crystalline Solids v. 323, p. 7-26.

Griscom, D.L., Beltrán-López, V., Pope K.O., Ocampo, A.C., 2003b, New geochemical insights from electron-spin-resonance studies of Mn^{2+} and SO_3^- in calcites: Quantitative analyses of Chicxulub crater ejecta from Belize and southern México with comparisons to limestones from distal Cretaceous-Tertiary-boundary sites: in Koeberl, C., and Martinez-Ruiz, F., eds., Impact Markers in the Stratigraphic Record: Berlin, Springer-Verlag, p. 229-270.

Hack, J.T., 1955, Geology of the Brandywine area and origin of the upland of southern Maryland: U.S. Geological Survey Professional Paper 267A, p. 1-43.

Hallam, A., 1984, Pre-Quaternary sea-level changes: Annual Reviews of Earth and Planetary Science, v. 12, p. 205-243.

Horton, J.W., Jr., Aleinikoff, J.N., Kunk, M.J., Naeser, C.W., and Naeser, N.D., 2005, Petrology, structure, age, and thermal history of granitic Coastal Plain basement in the Chesapeake Bay impact structure, USGS-NASA Langley Core, Hampton, Virginia: in Horton, J.W., Jr., Powars, D.S., Gohn, G.S., eds., Studies of the Chesapeake Bay Impact Structure – The USGS-NASA Langley corehole, Hampton, Virginia, and related coreholes and geophysical surveys: U.S. Geological Survey Professional Paper 1688, p. B1-B29.

Jaeger, H. M., and Nagel, S. R., 1992, Physics of granular states: Science**,** v. 255, 1524.

Jani, M.G., Bossoli, R.B., and Halliburton, L.E., 1983, Further characterization of the E'_1 center in crystalline SiO_2: Physical Review B, v. 27, p. 2285-2293.

Johnson, G.H., and Goodwin, B.K., 1967, Elliptical depressions on undissected highland gravels in northern Chesterfield County, Virginia, Abstracts : Virginia Journal of Science, v. 18, p. 166.

Kastner, M., Asaro, F., Michel, H.V., Alvarez, W., and Alverez, L.W., 1984, Did the clay minerals at the Cretaceous-Tertiary boundary form from glass? Evidence from Denmark and DSDP hole 465A: Journal of Non-Crystalline Solids, v. 67, p. 463-464.

King, D.T., Jr., Demick, J.D., Pope, K.O., and Ocampo, A.C., 1997, Surface markings on Chicxulub ejecta clasts from Belize, Central America: Geological Society of America Abstracts with Programs, v. 29(3), p. 12.

King, D.T., Jr., and Petruny, L.W., 2003, Stratigraphy and sedimentology of coarse impactoclastic breccia units within the Cretaceous-Tertiary boundary section, Albion Island, Belize: in Koeberl, C., and Martinez-Ruiz, F., eds., Impact Markers in the Stratigraphic Record: Berlin, Springer-Verlag, p. 347.

Koeberl, C., Poag, C.W., Reimold, W.U., and Brandt, D., 1996, Impact origin of the Chesapeake Bay structure and the source of the North American tektites: Science, v. 271, p. 1263-1266.

Krantz, D.E., and Powars, D.S., 2000, Hydrogeologic setting and potential denitrification in ground water, Coastal Plain of Southern Maryland: USGS Water-Resources Investigations Report 00-4051, 19 pp.

Lambeck, K., and Chappell, J., 2001, Sea level change through the last glacial cycle: Science, v. 292, p. 679-686.

McGetchin, T.R., Settle, M., and Head, J.W., 1973, Radial thickness variation in impact crater ejecta: Implications for lunar basin deposits: Earth and Planetary Science Letters, v. 20, p. 226-236.

Melosh, H.J., 1989, Impact Cratering – A Geological Process: New York, Oxford University Press, 245 p.

Michel, F.M., Ehm L., Antao S.M., Lee P.L., Chupas P.J., Liu G., Strongin D.R., Schoonen M.A.A., Phillips B.L., and Parise J.B., 2007, The structure of ferrihydrite, a nanocrystalline material: Science, v. 316, p. 1726-1729.

Mitroff, I., 1975, The Subjective Side of Science: A Philosophical Inquiry into the Psychology of the Apollo Moon Scientists: Amsterdam, Elsevier, p. 345.

Mixon, R.B., 1985, Stratigraphic and geomorphic framework of uppermost Cenozoic deposits in the southern Delmarva Peninsula, Virginia, and Maryland. U.S. Geological Survey Professional Paper 1067-G, p. 53.

Mixon, R.B., Berquist, C.R., Jr., Newell, W.L., Johnson G.H., Powars, D.S., Schindler, J.S., and Rader, E.K., 1989, Geologic map and generalized cross sections of the Coastal Plain and adjacent parts of the Piedmont, Virginia: U.S. Geological Survey Miscellaneous Investigations Series Map I-2033, 2 sheets, Scale 1:250,000.

Mutch, T.A., 1970, Geology of the Moon – A Stratigraphic View: Princeton, Princeton University Press, 324 p.

Oberbeck, V.R., 1975, The role of ballistic erosion and sedimentation in lunar stratigraphy: Reviews of Geophysics and Space Physics, v. 13, p. 337-362.

Ocampo, A.C., Pope, K.O., and Fischer, A.G., 1996, ejecta deposits of the Chicxulub crater from Albion Island, Belize: in Ryder, G., Fastovsky, D., and Gartner, S., eds., The Cretaceous-Tertiary Event and Other Catastrophes in Earth History: Boulder, Colorado, Geological Society of America Special Paper 307, p. 75-88.

Pälike, H., Norris, R.D., Herrle, J.O., Wilson, P.A., Coxall, H.K., Lear, C.H., Shackleton, N.J., Tripati, A.K., and Wade, B.S., 2006, The heartbeat of the Oligocene climate system: Science, v. 314, p. 1894-1898.

Peebles, P.C., 1984, Late Cenozoic landforms, stratigraphy, and history of sea level oscillations of southeastern Virginia and northeastern North Carolina: Ph.D. Thesis, College of William and Mary, Williamsburg, Va., 227 p.

Pettijohn, F.J., 1949, Sedimentary Rocks: New York, Harper and Brothers, p. 526.

Poag, C.W., 1997, The Chesapeake Bay bolide impact: a convulsive event in Atlantic Coastal Plain Evolution: Sedimentary Geology, v. 108, p. 45-90.

Poag, C.W., Powars, D.S., Poppe, L.J., and Mixon, R.B., 1994, Meteroid mahem in Ole Virginny: Source of the North American tektite strewn field: Geology, v. 22, p. 691-694.

Poag, C.W., Koeberl, C., and Reimold, W.U., 2004, The Chesapeake Bay Crater: Geology and Geophysics of a Late Eocene Submarine Impact Structure: Berlin, Springer-Verlag, 522 p.

Pope. K.O., Ocampo. A.C., Fischer, A.G., Alvarez, W., Fouke, W., Webster, C.L., Vega, F.J., Smit, J., Fritsche, A.E., and Claeys, P., 1999, Chicxulub impact ejecta from Albion Island, Belize: Earth and Planetary Science Letters, v. 170, p. 351-364.

Schlee, J., 1957, Upland gravels of Southern Maryland: Geological Society of America Bulletin, v. 68, p. 1371-1409

Schneider, S.J., 1969, Melting points in ceramics: in Wachtman, J.B. Jr, ed., Mechanical and Thermal Properties of Ceramics, Washington, DC, U.S. Govt. Printing Office, p. 19-39.

Schultz, P. H., 1992, Atmospheric effects on cratering efficiency, Journal of Geophysical Research, v. 97(E1), p. 975–1005.

Shultz, P.H., and Gault, D.E., 1979, Atmospheric effects on martian ejecta emplacement: Journal of Geophysical Research, v. 84, p. 7669-7687.

Smit, J., 1999, The global stratigraphy of the Cretaceous-Tertiary boundary impact ejecta: Annual Review of Earth and Planetary Science, v. 27, p. 75-113.

Warren, P.H., Claeys, P., and Cedillo-Pardo, E., 1996, Mega-impact melt petrology (Chicxulub, Sudbury, and the Moon): Effects of scale and other factors on potential for fractional crystallization and development of cumulates: in Ryder, G., Fastofsky, D., and Garner, S., eds., the Cretaceous-Tertiary Event and Other Catastrophes in Earth History, Boulder, Colorado, Geological Society of America Special Paper 307, p. 105-124.

Wentworth, C.K., 1928, Striated cobbles in the southern States: Geological Society of America Bulletin, v. 39, p. 941-953.

Wentworth, C.K., 1930, Sand and gravel resources of the Coastal Plain of Virginia: Virginia Geological Survey Bulletin, v. 32, 146 p.

Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001, Trends, rhythms, and aberrations in the global climate 65 Ma to present: Science, v. 292, p. 686-693.

FIGURE CAPTIONS

Figure 1. Map of the upland deposits of Southern Maryland, Eastern Virginia, and the District of Columbia and of the Bacons Castle formation of Eastern Virginia, displayed in relation to the buried Chesapeake Bay crater. Mapped land surfaces were taken from Schlee (1957) and Frye (1986). The gravity map of the impact structure with the crater rim indicated by dashes was taken from Koeberl et al. (1996). Dashed circle at 115 km radius from the crater center at Cape Charles, Va., is provided for visual reference.

Figure 2. A: Moving average modal-gravel-size isoliths for the gravel member of the upland deposits of the District of Columbia and the Potomac-Patuxent peninsula of Southern Maryland adapted from Schlee (1957). B: Modal gravel size plotted on a semi-log scale versus distance for four separate paths chosen to be nearly as possible perpendicular to the isoliths of A (replotted from Schlee, 1957) with comparisons to the cobble-size reduction rates for the modern Rhine and Mur Rivers (Pettijohn, 1949). Arrows, circles, and commentary external to the boxed graphs added by the author.

Figure 3. A: Pile of quartzite pebbles and cobbles from the upland deposits that were sorted by size and fracture frequency. B: Histogram of the results. C: A dramatically fractured cobble from the same residential lot in the Hollin Hills subdivision of Faifax County, Va. D: Cartoon typical of the fracture patterns of fused silica optical fibers broken under high tensile stress.

Figure 4. An idealized profile of the Chesapeake Bay crater ejecta blanket tens of minutes after the impact calculated by means of equation 1, using the coefficient derived in equation 3 and the notional assumption that the Eocene Coastal-Plain sediments were sloping seaward at 0.5 m/km at the time of the impact. The lithology of the target rocks follows Koeberl et al. (1996) and Poag (1997), and the overall sediment depths were scaled with reference to the depth at which the USGS-NASA Corehole intercepted basement granite (Horton et al., 2005). Also indicated is the effective interference zone (Melosh, 1989) from which all materials, excluding the impactor footprint, would have been ejected at high speeds with minimal shocking.

Figure 5. Stratigraphic section of Southern Maryland reproduced from Krantz and Powars (2000), with comparison to an idealized calculation of the original Chesapeake Bay crater ejecta blanket. Note the gently sloping, concave-upward base of the upland deposits and the depths of these deposits that a comparable to the calculated idealized ejecta blanket depth.

Figure 6. Map of the upland deposits and Bacons Castle formation showing the sections for which elevation profiles are shown in Figures 5 (Southern Maryland) and 7 (Eastern Virginia: Fig. 7A, B, C, and D) as correspondingly marked here. Note that only sections D are truly radial to the center of the crater. The other sections radiate from points near the crater center selected

to intercept long, mostly contiguous stretches of the upland deposits, as well as the Bacons Castle formation.

Figure 7. Elevation profiles of the upland deposits and Bacons Castle formation in Virginia along the sections illustrated in Figure 6. A: the Potomac-Rappahannock peninsula. B: the Rappahannock-York peninsula. C: the York-James peninsula. D: the quadrant south of the James River. Multiple profiles on the same peninsula are arranged so that the southern most is represented by small black squares, the central one by large hollow circles, and the northernmost by a dash-dot line. In A and B, calculated profiles of the original Chesapeake Bay crater ejecta blanket have been overlain. The spatial resolution of these elevation data varies in respect to the eye altitude "flown" on Google Earth. Thus, data showing high spatial frequency of hills and valleys were "flown" at an eye altitude typically ~10 km, while those recording only low-frequency elevation changes were "flown" at eye altitudes of ~50 km or more. In D, high-altitude data were supplemented low-altitude data to sharpen river valleys and scarps.

Figure 8. A: Location of the Dragon Run watershed (Dragon Run Steering Committee, 2003) relative to the Chesapeake Bay impact structure (the latter being indicated by means of the gravity map of Koeberl et al., 1996). B: A linear string of secondary impacts associated with the lunar crater Copernicus (Lunar Orbiter photograph LO V M-144). The Dragon Run watershed is a similar raised rim depression. Its boundary in green is a fluvial-transport divide; all streams inside it flow inward. The dashed arrow indicates a proposed circumferential paleochannel of the York River before it succeeded in breached the crater rim.

Figure 9. A: Sketch of an exposed outcrop of Midlothian upland gravels (Goodwin and Johnson, 1970). B: A profile of the Midlothian gravels along U.S. Route 60 west of Richmond, Va. (hollow squares), with comparison to a straight-line profile roughly 8 km southward (black squares). All distances are relative to the center of the Chesapeake Bay impact structure. The dotted areas represent the author's general understanding of the approximate lateral and vertical extents of the upland deposits along these profiles, based on maps (Frye, 1986) and written accounts (Goodwin and Johnson, 1970).

Figure 10. A: The Hollin Hills diamicton: an unsorted whitish-gray siliciclastic stratum ~30 cm high above the stream bed and an unknown, but probably not great, depth below it. B: An autobrecciated chalk ball recovered from the stream bed at site A. C: A massive "cinder" composed of iron oxide, possible glass, quartz silt, and 5-10 mm quartz clasts, recovered near site A. D: Rock fragment (tentatively greenstone) with a multi-layer external coating also recovered near site A. E: A multiply-fractured quartzite cobble found still nested in this stratum.

Figure 11. A: A 27-kg granite boulder recovered by the author in his front yard in Hollin Hills, Fairfax County, Va. B: Thin-section photomicrograph of a chip of boulder of A showing

amorphous biotite extruded from the interior to the exterior (from the lower left to upper right). C: Thin-section photomicrograph of a feldspar grain showing planar deformation features in the interior of the chip. Images B and C were both recorded under cross-polarized light. Object A is suspected to be interference-zone ejecta from the 35.5-Ma Chesapeake Bay impact.

Figure 12. Cumulative particle size distributions for various erosive environments. The solid black data points are after Emery (1955) and the open circles are due Schlee (1957); in both of these cases the distributions were recorded percentages of the total number of samples. Schlee’s samples were pre-selected to have median sizes larger than 10 mm. Attal and Lavé (2006) sorted by lithology and then weighed particles coarser than 10 mm in the field; the size distribution of the finer fraction was determined in the laboratory.

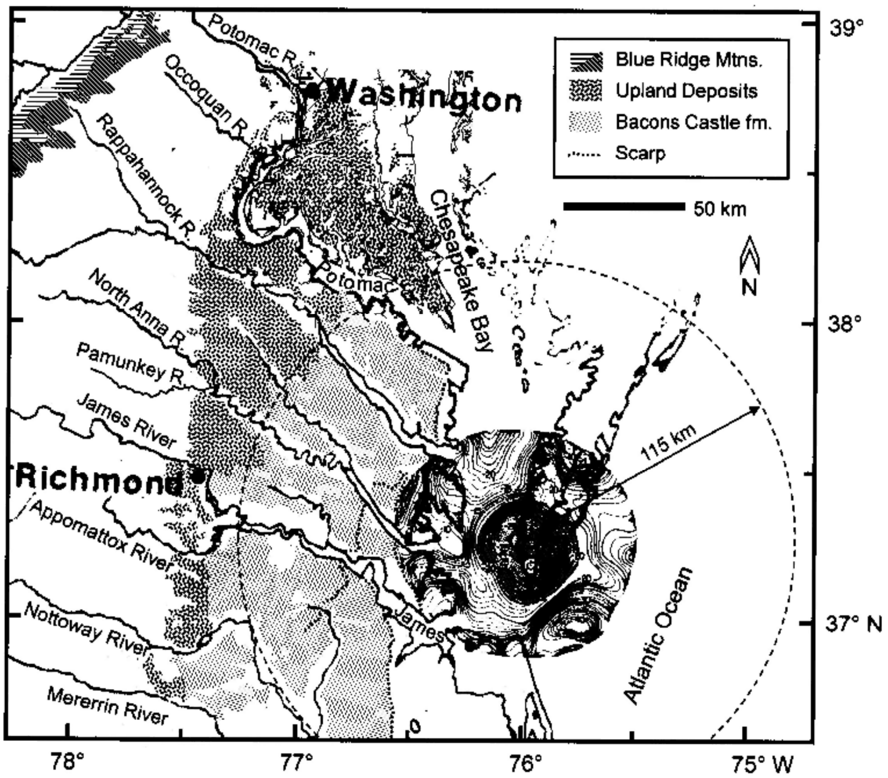


Figure 1

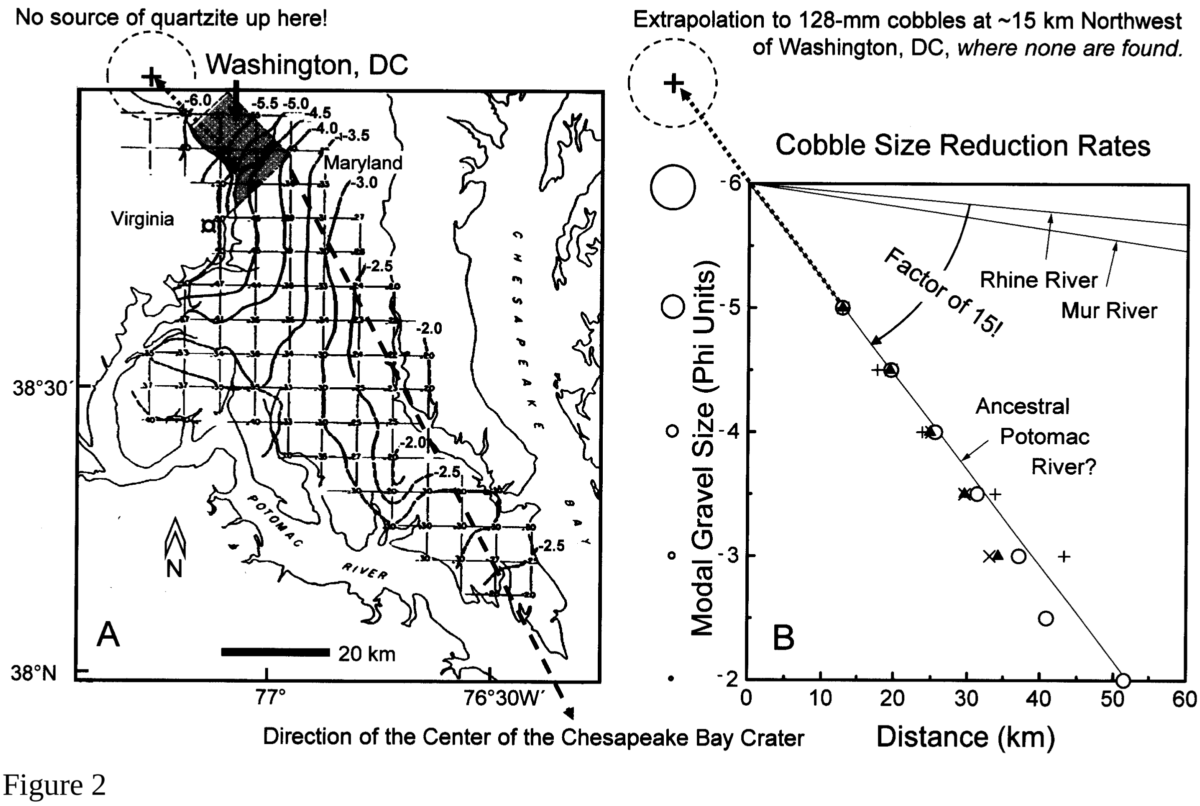


Figure 2

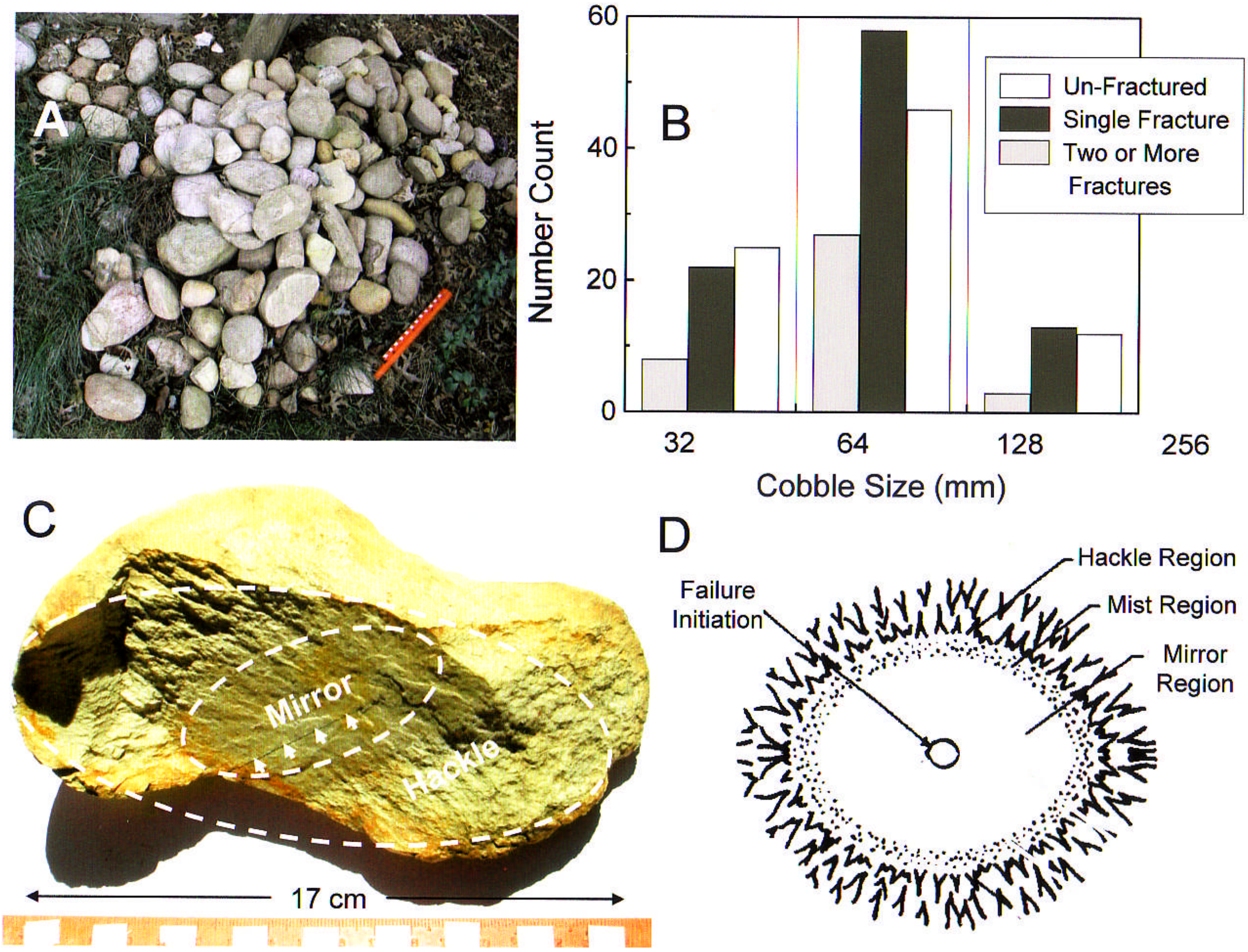

A
B
Number Count
60
40
20
0
32
64
128
256
Cobble Size (mm)
Un-Fractured
Single Fracture
Two or More
Fractures
C
Mirror
Hackle
17 cm
D
Hackle Region
Failure
Initiation
Mist Region
Mirror
Region

Figure 3

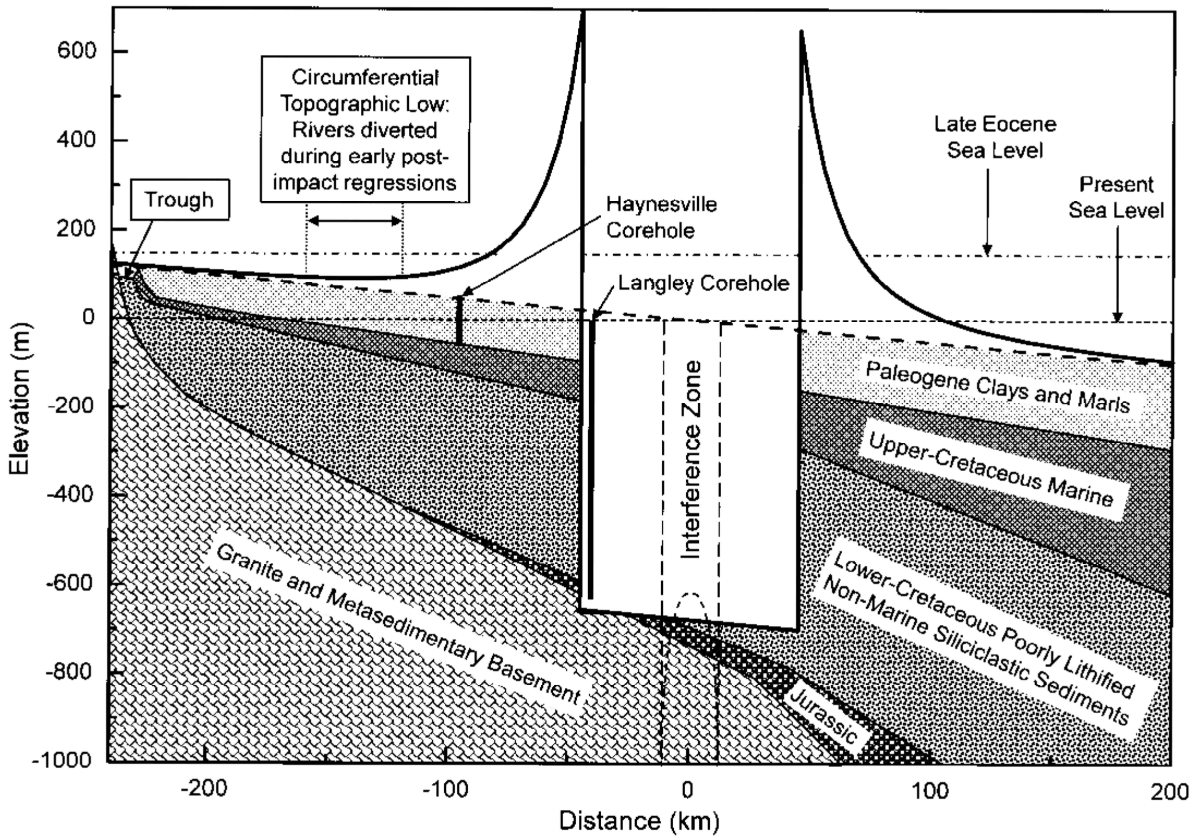


Figure 4

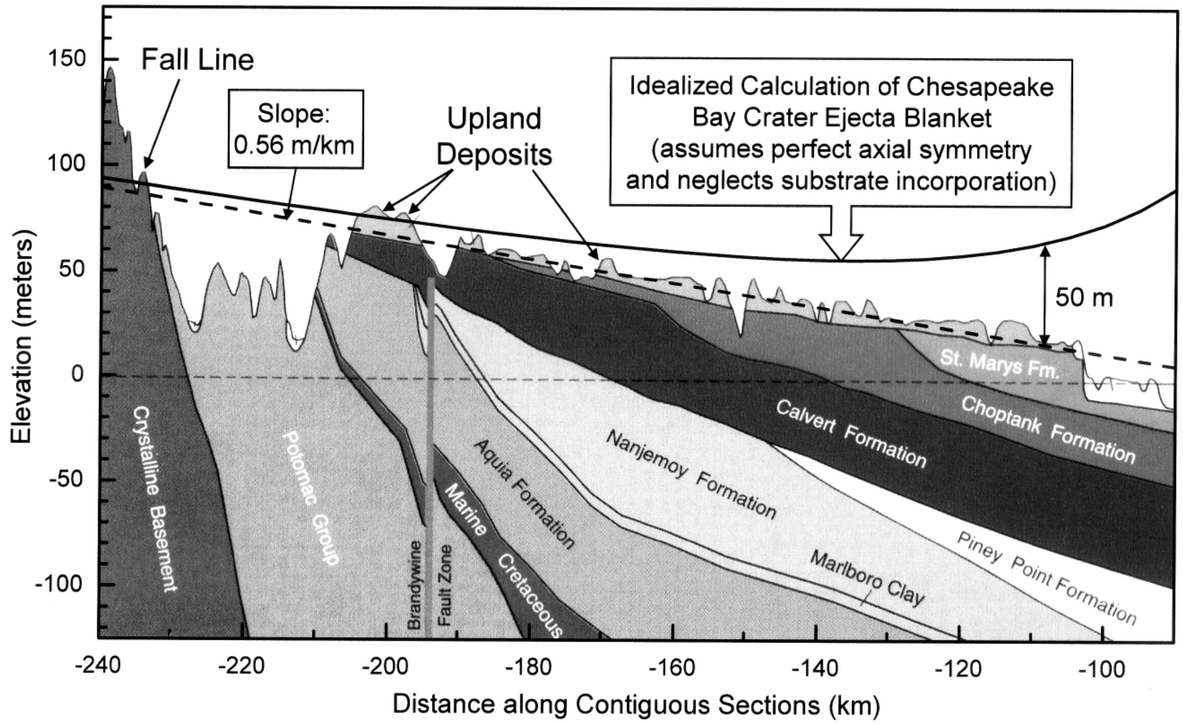
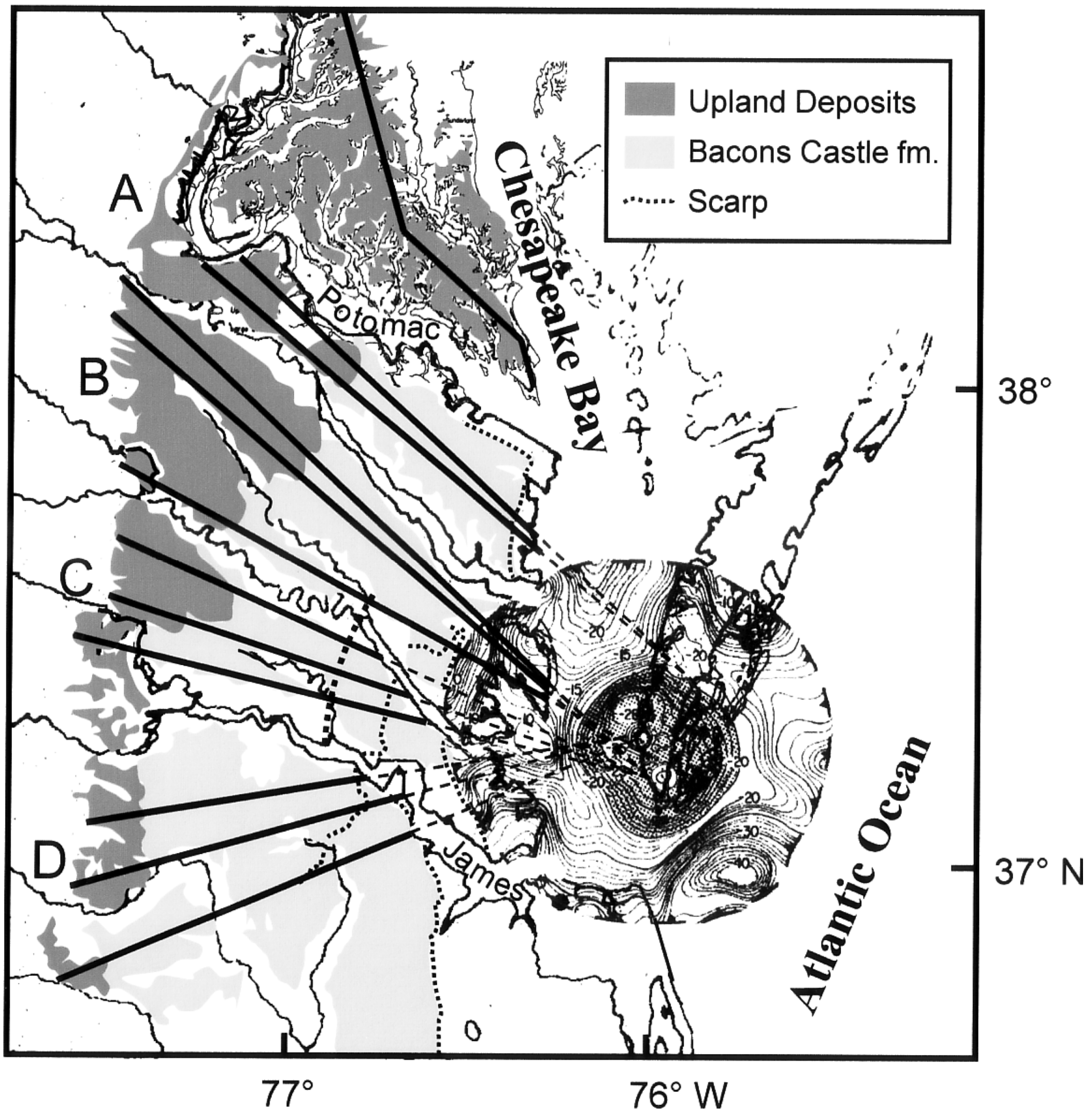


Figure 5



Upland Deposits
Bacons Castle fm.
Scarp
A
B
C
D
Potomac
Chesapeake Bay
James
Atlantic Ocean
38°
37° N
77°
76° W

Figure 6

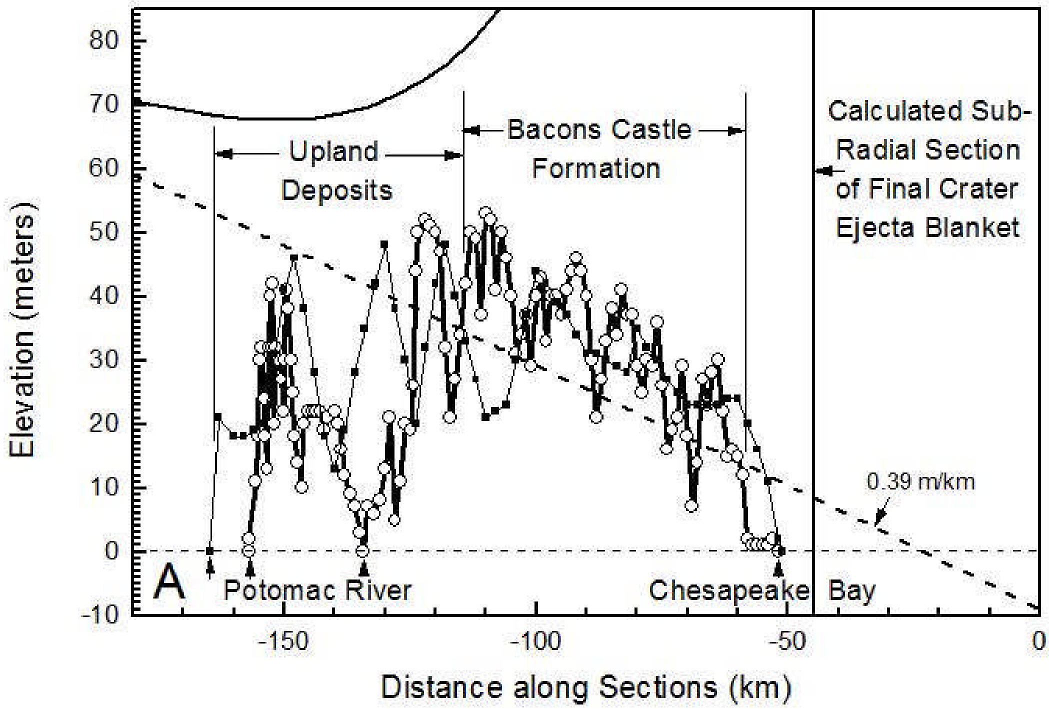


Figure 7A

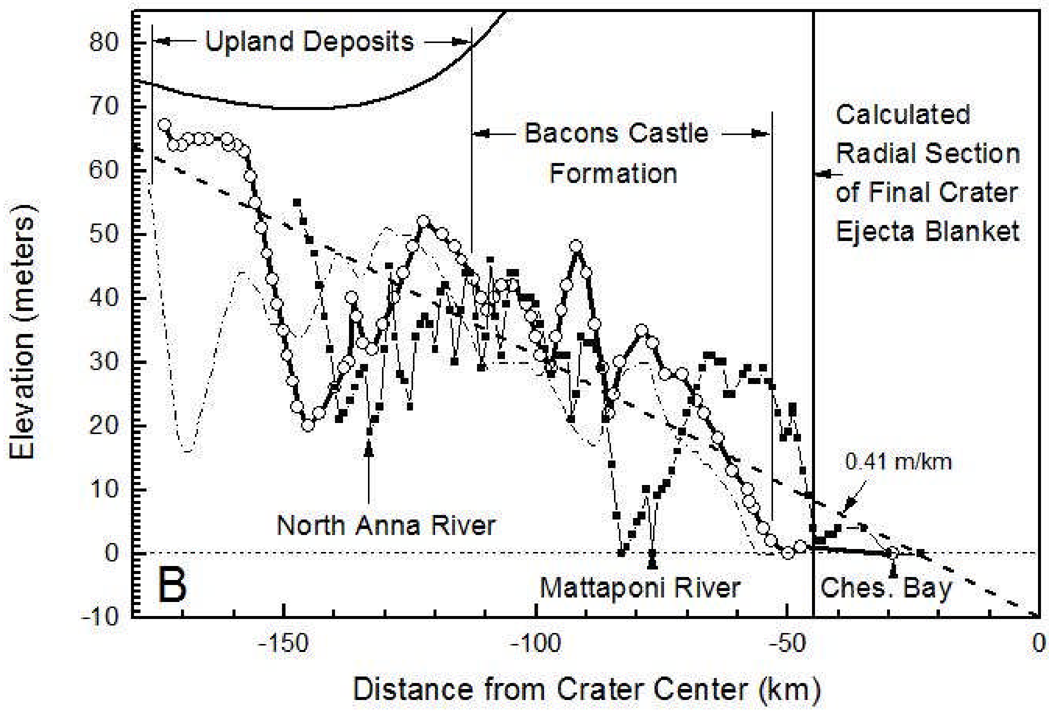


Figure 7B

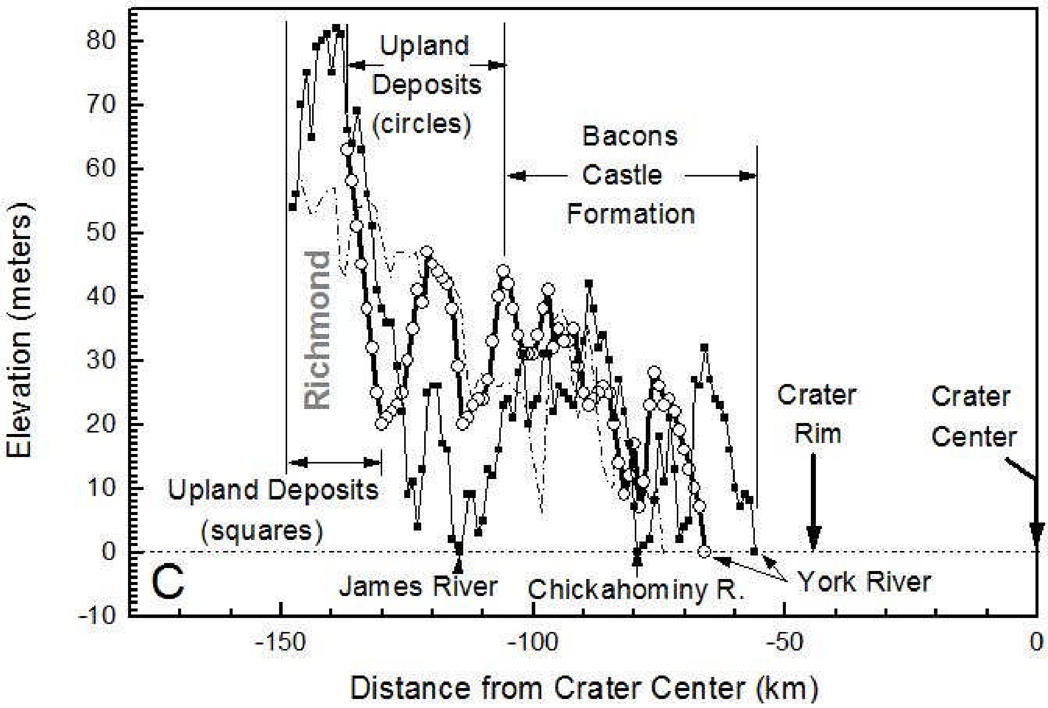


Figure 7C

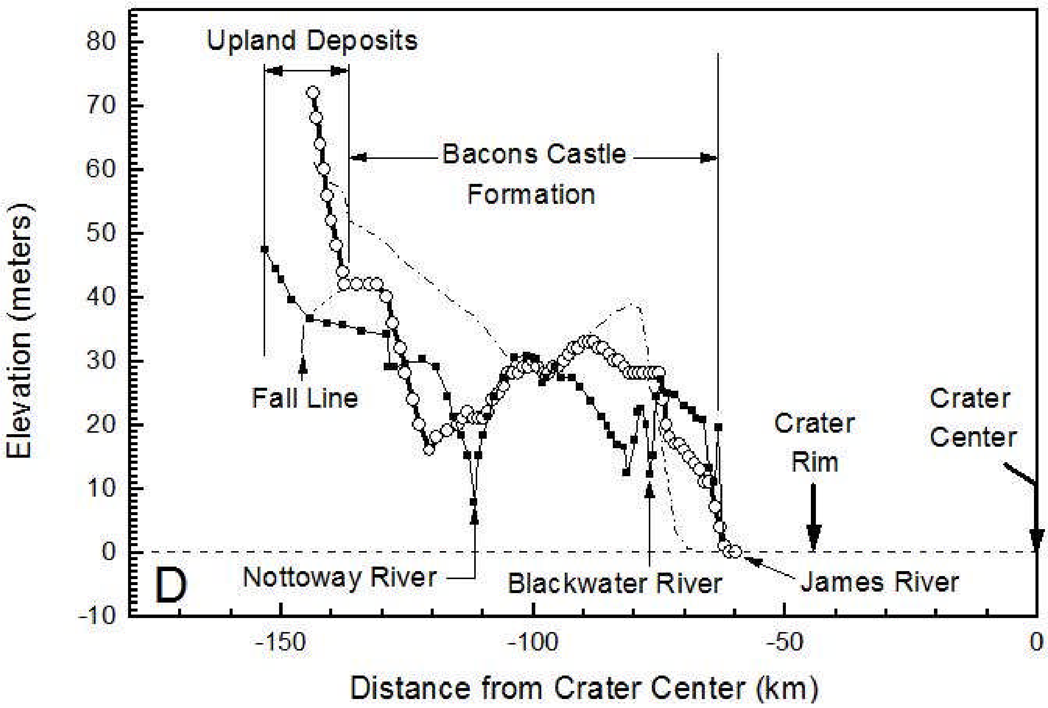


Figure 7D

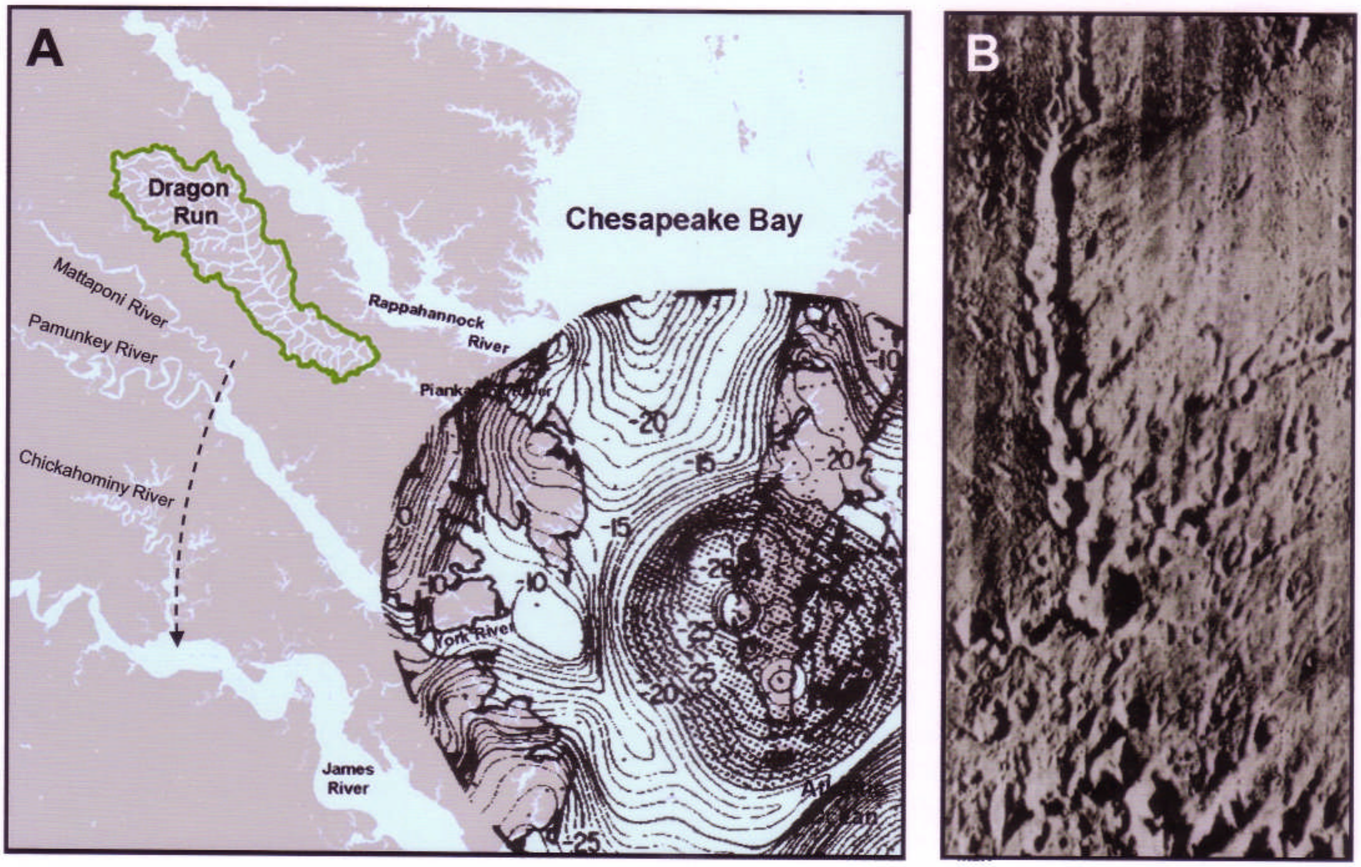

A
Dragon Run
Chesapeake Bay
Mattaponi River
Pamunkey River
Rappahannock River
Chickahominy River
York River
James River
B

Figure 8

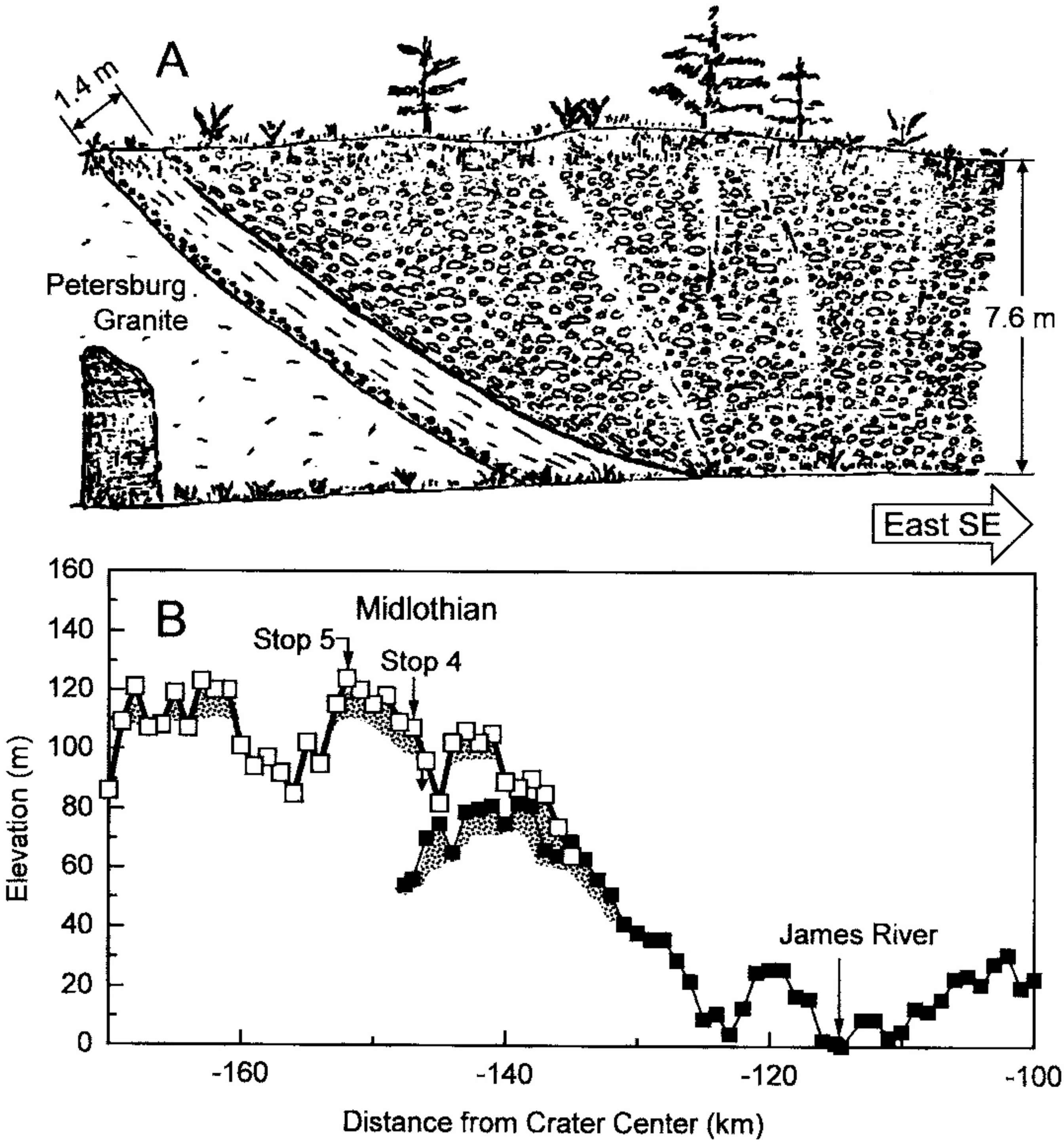

A
1.4 m
Petersburg
Granite
7.6 m
East SE
B
Midlothian
Stop 5
Stop 4
James River
160
140
120
100
80
60
40
20
0
Elevation (m)
-160
-140
-120
-100
Distance from Crater Center (km)

Figure 9

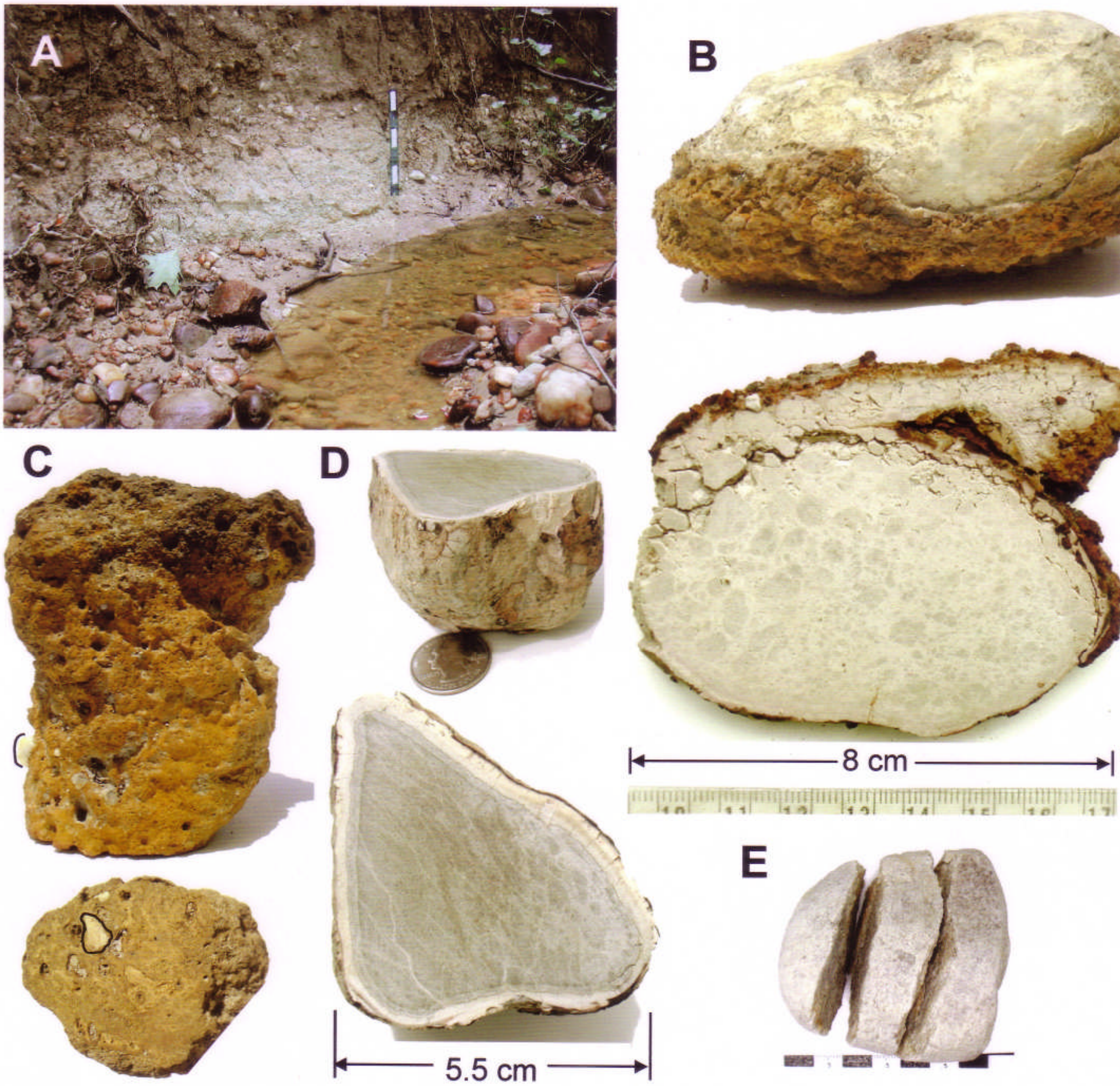


Figure10

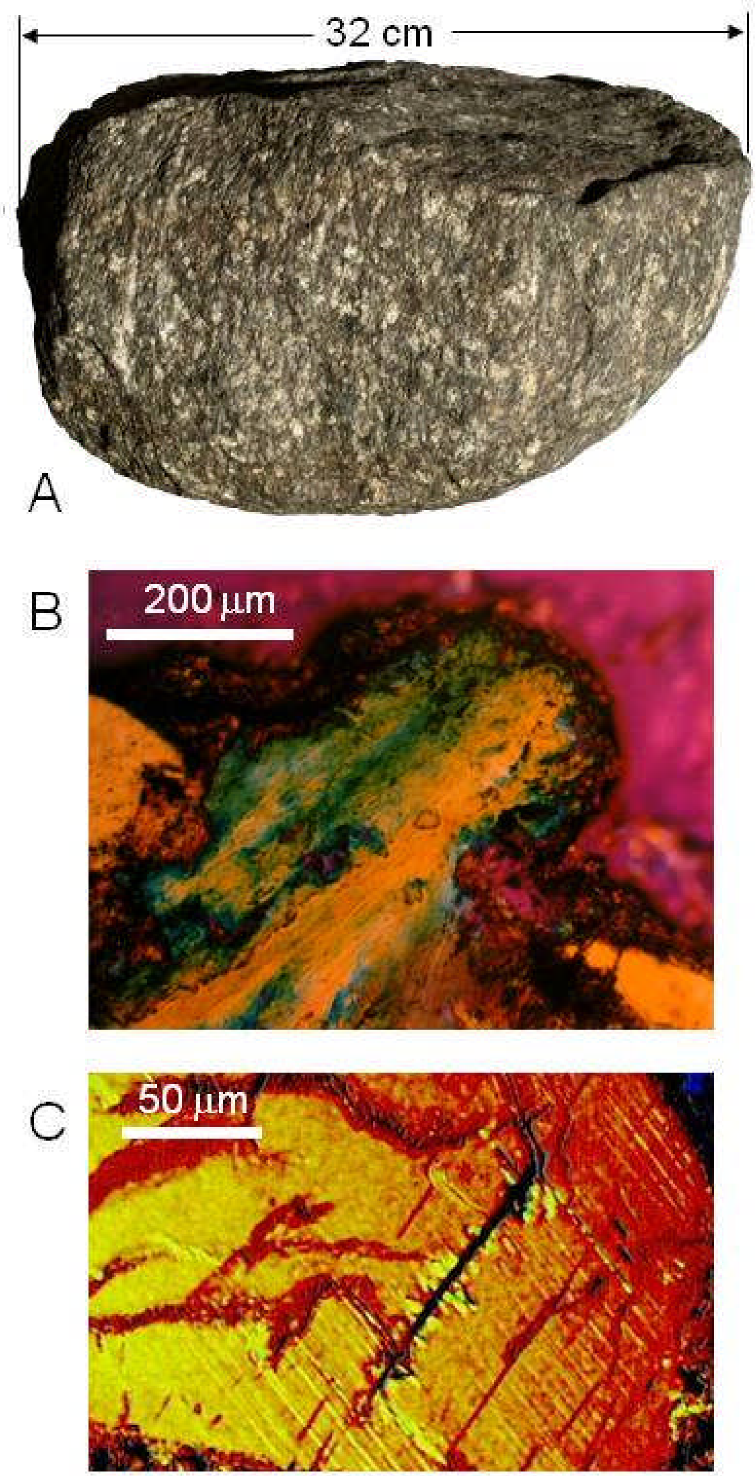
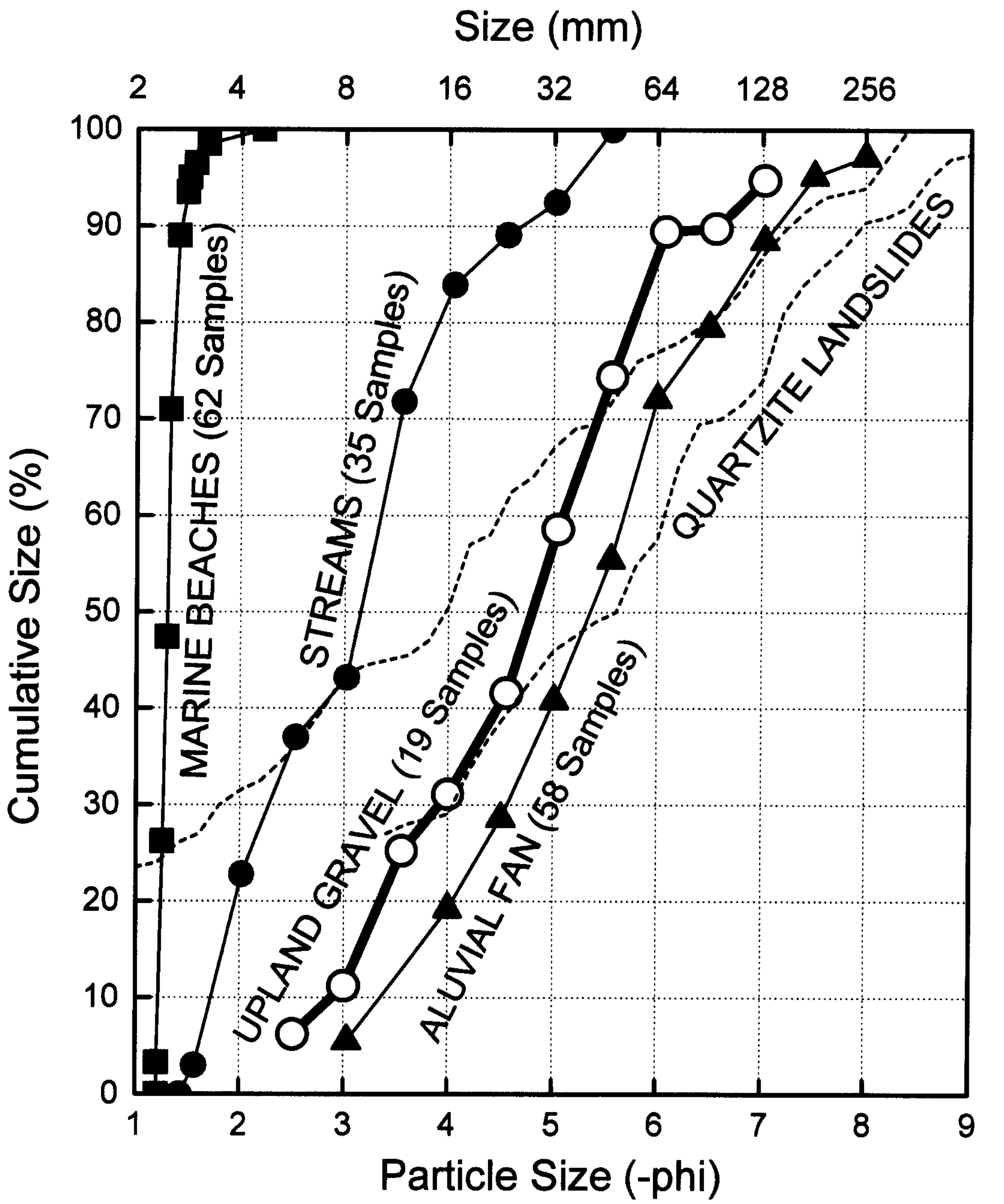


Figure 11



Size (mm)
2
4
8
16
32
64
128
256
Cumulative Size (%)
100
90
80
70
60
50
40
30
20
10
0
MARINE BEACHES (62 Samples)
STREAMS (35 Samples)
UPLAND GRAVEL (19 Samples)
ALUVIAL FAN (58 Samples)
QUARTZITE LANDSLIDES
1
2
3
4
5
6
7
8
9
Particle Size (-phi)

Figure 12