A3 AND B3: SALAMANCA ("LITTLE ROCK CITY") CONGLOMERATE TIDE-DOMINATED & WAVE-INFLUENCED DELTAIC/COASTAL DEPOSITS UPPER DEVONIAN (LATE FAMMENIAN) CATTARAUGUS FORMATION

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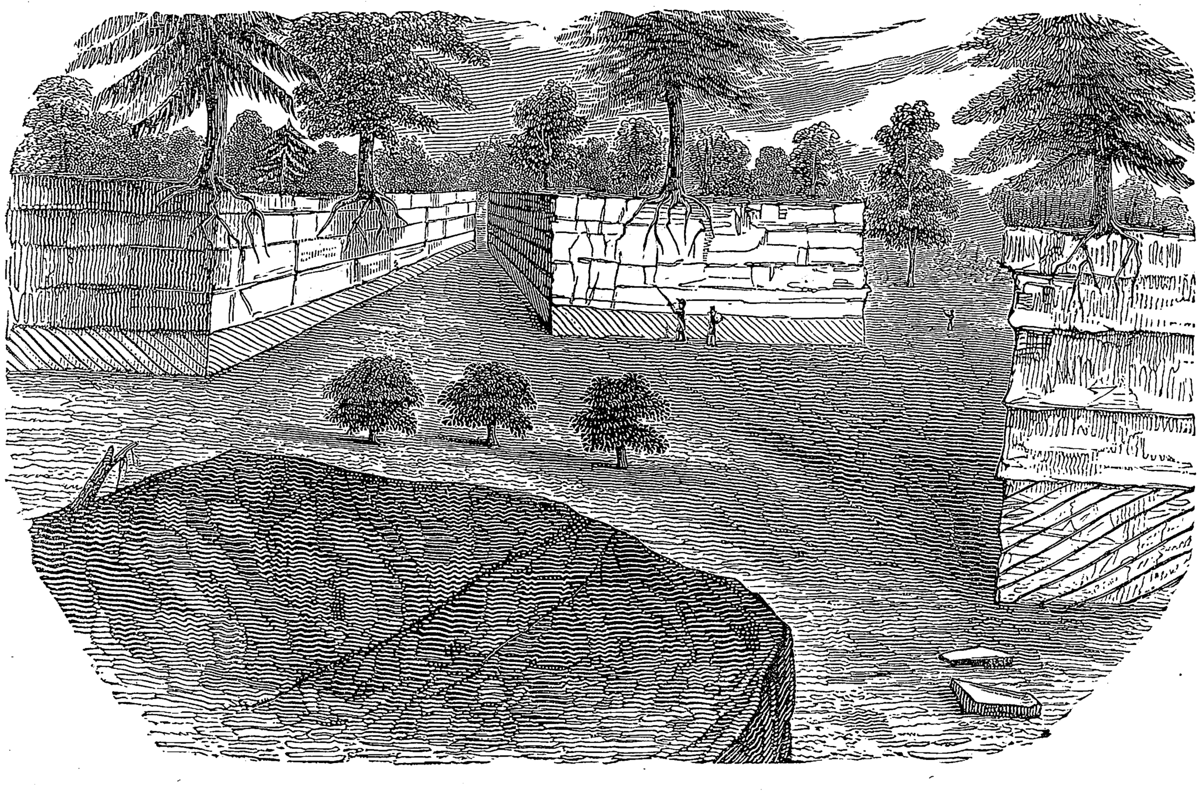


View in Rock city, seven miles south of Ellicottville. From a sketch by Mr. E. N. Horsford.

(*Geology of New York*, 1843)

INTRODUCTION

On a hilltop in Rock City State Forest, three miles north of Salamanca, New York, the Salamanca Conglomerate outcrops in spectacular fashion. Part of the Upper Devonian (late Fammenian) Cattaraugus formation, the quartz-pebble conglomerate forms a five to ten-meter high escarpment and topographic bench at ~ 2200 feet elevation amid a mature cherry-maple-oak forest. In places, house-sized blocks have separated from the escarpment along orthogonal joint sets and variably "crept" downhill. Where concentrated, a maze of blocks and passageways may form so-called "rock cities", an impressive example of which is Little Rock City. The well-cemented blocks permit extraordinary 3-D views of diverse and ubiquitous sedimentary structures and features.

Six outcrop areas with the most significant exposures were logged over a four-kilometer north-south traverse. The traverse largely follows the east-facing hillside which roughly parallels the presumed paleoshore of the Devonian Catskill Sea. Extensive "bookend" outcrops at the north face (off the Rim Trail) and at the southeast perimeter ("Little Rock City" along the North Country-Finger Lakes Trail) and vertical (caprock) control allow a nearly continuous look at spatial and temporal changes in sedimentary deposits along a four-kilometer stretch of inferred late Devonian seacoast.

We'll examine several outcrops which reflect a high-energy and varied coastline as summarized below.

Summary of Findings

Three major depositional environments are interpreted from north to south:

Shoreface to foreshore (beach) coarsening-upward sequence - ("north face" **Outcrops #2 & 3** from base)

- ~ 1 m of thin-bedded (5-10 cm) wave cross-laminated strata; mostly buff, medium sand with some coarse sand, granules, and a few fine pebbles.
- ~ 3 m of amalgamated coarse-grained, large (10-20 cm x 50-100 cm), smooth-crested wave ripples with abundant pebbles (some apparent 3-D forms seem without analogues), interbedded in places w/thin fine-grained (rolling-grain) wave ripples; some trough/planar cross-beds near top.
- ~ 3 m of parallel/low-angle strata of gray interbedded coarse sand and pebbles.

Prograding tide-dominated delta (w/coarse-grained distributaries, tidal channels, bars, and shoals) **(Outcrops # 4, 5, & 6)**

- abundant channels (abundant pebbles, meters to tens of meters wide, ~1-2 m deep) and channel point bars (coarse sand to pebble lateral-accretion deposits of tidal, delta distributary/fluvial channels). One channel complex directly overlies fine-grained wave ripple-laminated (marine) sandstones.
- cross-bedded strata of various dimensions (~ 0.05 m to +1 m) commonly arranged in cosets, some bidirectional.
- current indicators mainly directed shoreward (E-SE) but bi-directional cross-beds common; some truncation surfaces show wave influence.

Sub-aqueous tidal dune field – **Outcrop area # 7** ("Little Rock City")

- very large scale (up to 5+ m thick) planar 2-D cross-beds with fine-to-coarse sand and abundant granule/fine pebble concentrations and occasional larger pebbles; dune foresets mostly inclined 20-30^{o} and ~ 5-10 cm thick; granule layers usually thicker; some dunes are traceable up to 150 m across several blocks.

- foreset azimuths (50^o to 150^o) show dunes migrated parallel with and toward the paleoshore with no major reactivation surfaces; most toesets are tangential; planar truncation surface at the top of the dunes shows wave influence.
- a complete 2-3 m dune bedform (“form-set”: foreset, topset, stoss preserved); core shows directionally-opposed cross-strata which aggraded vertically until one flow direction (100^o – apparent flood tides) prevailed and the ~ 3 m dune began to migrate by periodic foreset deposition.

The entire sequence is overridden by ~ 2-3 m thick channel/lateral accretion deposits with some reddish, well-oxidized strata and plant remains. The caprock varies spatially and is generally similar to the underlying deposits with some reworking evident. Within the deltaic sequence, the uppermost caprock contains large (average 2-4 cm; up to 7 cm) densely/randomly-packed flat-lying vein-quartz pebbles (and some red and brown sandstone, a jasper? clast w/quartz veins, and red mudstone rip-up clasts not seen elsewhere) with abundant aligned plant remains. And finally, with diligent search, wave-ripple laminated buff-colored sandstones with abundant marine fossils (not seen elsewhere) can be found draping the caprock in this area.

The orthoquartzitic Salamanca conglomerate evidently records a high-energy Upper Devonian seacoast, with at least meso-tidal range, as indicated by a pebbly beach, a tide-dominated delta prograding over marine wave-rippled fine sands, and a sub-aqueous large-scale dune field formed by strong flood tides. Most of the sequence records delta progradation and sediment transport/redistribution along shore to dunes and beaches by tides and waves. Well-exposed channel deposits at the top (which overlie wave-truncated dunes and beach deposits at a similar elevation) suggest either expansion of the delta/delta plain or a transition to a coastal plain terrestrial environment (perhaps including a major flood event as suggested by localized large clasts of quartz, sandstone, mud rip-up clasts, and abundant plant fossils) followed by an apparent abrupt rise in relative sea level and a transgression as indicated by subsequent fine-grained wave-formed strata with an abundant marine fossil fauna.

Location and Physiographic Setting

The conglomerate beds of southwestern New York have long been a source of wonder. Appearing in widely-scattered and limited outcrops and more often, as isolated “float” blocks, these beds may more rarely form accumulations of large joint-separated blocks (“buildings”) and passages (“streets”) dubbed “rock cities”. Examples include “Rock City Park” south of Olean (Pennsylvanian age), “Thunder Rocks” (Mississippian? age) atop Allegany State Park, “Panama Rocks” (Upper Devonian age) and “Little Rock City” (the Upper Devonian Salamanca Conglomerate), the subject of this study and perhaps the finest example of a rock city in an unrivaled and freely-accessible setting.

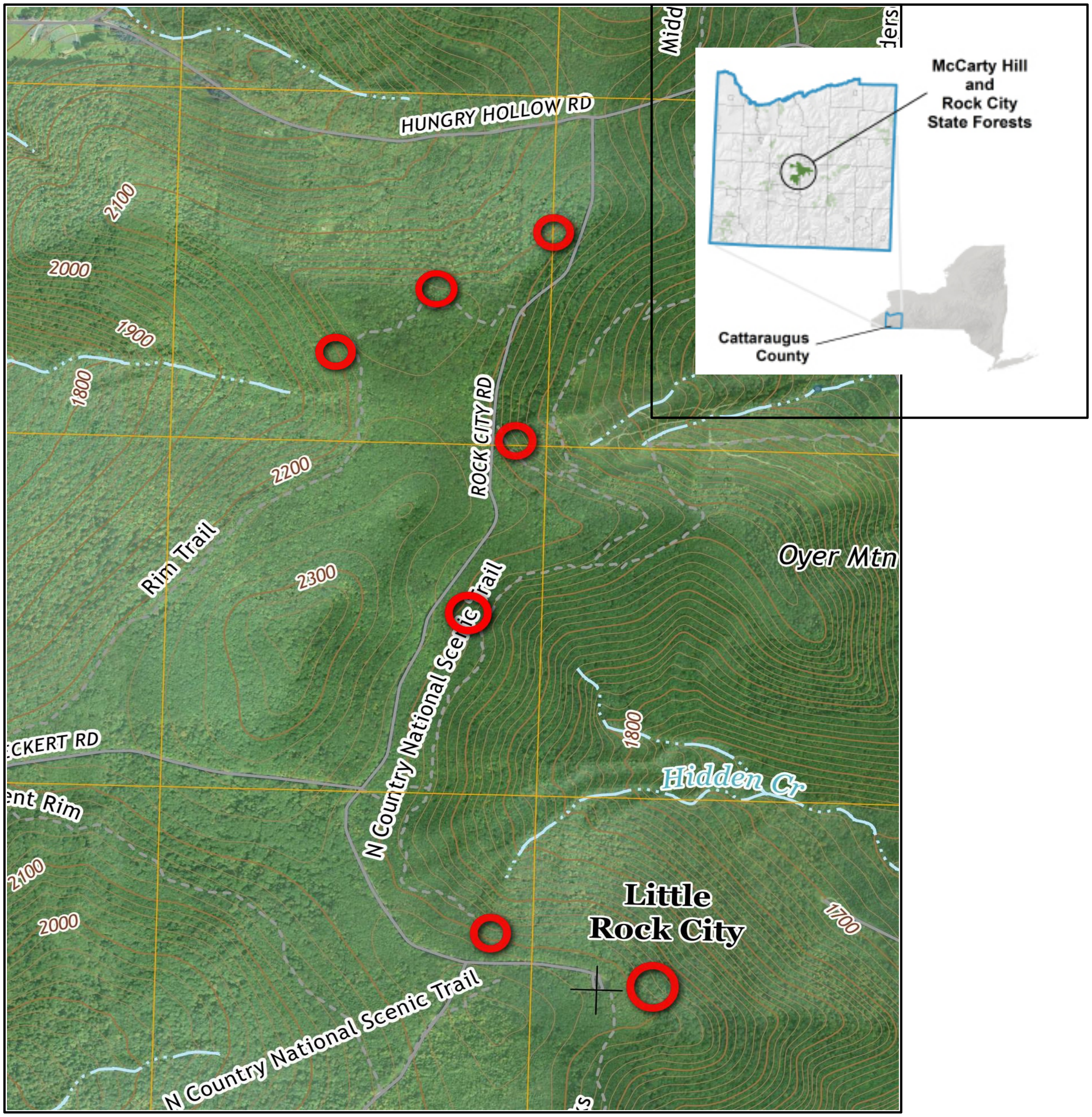


Figure 1. **<u>Location Map</u> – Outcrops in Red (#1 - #7 - North to South)**

<u>Scale</u>: 1 cm = 200 m <u>Source</u>: USGS – Salamanca Quadrangle (2016)

The Salamanca Conglomerate outcrops prominently (up to a 10m escarpment) and forms a locally-widespread plateau (~ 2200 feet elevation) in Rock City State Forest and adjacent McCarty Hill State Forest (http://www.dec.ny.gov/lands/77184.html). This mature forest of cherry, maple, and oak blankets nearly ten square miles of Appalachian Plateau uplands between Ellicottville and Salamanca, NY. "Little Rock City" (LRC - outcrop #7) at the southeast outcrop perimeter is the type locality (Tesmer, 1975). With perhaps the most exquisite exposures, LRC has been an attraction since the early 1800s (e.g., Hall, 1843). Much of the outcrop belt is partially obscured by vegetation, rubble, and in places, glacially-deposited debris but it is readily traceable around the entire hill perimeter as facilitated by a network of hiking trails such as the North Country National Scenic Trail, the Finger Lakes Trail, and the Rim Trail. The outcrop and separated "blocks" are also readily visible with online orthoimagery (https://orthos.dhses.ny.gov/) and can be traced to nearby hillsides.

Glaciation – Evidence and Effects

The study area is mapped within the Salamanca Re-entrant, which is part of the unglaciated Appalachian Plateau and northernmost unglaciated area in the eastern United States. Muller (1977) placed an "uncertain" glacier margin at about 1800 feet elevation at roughly one to five kilometers north of the outcrop belt. However, evidence of glaciation in the study area includes: (1) "drab" glacial till (chaotically-oriented thin-bedded sandstone in a gray clay matrix) exposed in a small ephemeral stream east of Eckert Road at 2200 feet AMSL (the only stream found at this elevation), (2) some large quartz clasts which appear shattered and sheared-off level with the top surface of caprock, and (3) a stretch of outcrops disrupted and largely "buried" with float that includes, in places, abundant slabs of thin-bedded wave-rippled fossiliferous sandstone. This evidence is clustered in the area of subdued outcrops (limited exposures of ~ 2m), from Outcrop #5 to Salamanca Road that includes a topographic col/saddle which may have focused ice movement albeit in east-west directions. In addition, Smith and Jacobi (2006) reported an upside-down house-sized block atop another block as evidence of glacial activity.

Other areas appear largely unaffected such as the isolated and well-weathered "sentinel" blocks (outcrop #1) and the isolated erosional remnants (~ 3m "cubes") perched on the escarpment at outcrop #4 and some extensive block "fields" at the NW and SE corners; outcrops #3 and #6). It appears then that direct glaciation affected this area variably but periglacial effects such as permafrost, prolonged freeze-thaw cycles, and ice wedging were likely intense. Such conditions likely enhanced block separation, undermining/slump, and downslope movement due to solifluction ("soil flow"/creep due to saturated conditions) and genifluction, (creep in contact with ice/permafrost; e.g., Millar & Nelson, 2001). And the general process of soil creep continues, typically the slowest (~ mm/year on average) but geologically the most significant mass movement process (Allen, 1982).

Structural Geology

The regional dip is gently southward (about 30 feet/mile – S/SW; Glenn, 1902 and 20-50 feet/mile – South; Tesmer, 1963). No surface expression of folds or faults were observed but Glenn (1902) reported small folds in Cattaraugus County and the Clarendon-Linden fault complex is nearby in Allegany County (Smith and Jacobi, 2006). Jointing is the most obvious structural feature as it controls the similar block dimensions and the extraordinary rock exposures on the sides of the blocks. The vertical joint sets are generally orthogonal, spaced ~ 10-20 meters apart, and trend NE-SW (30^{o}-45^{o}) and NW-SE (125^{o}-140^{o}). Per Engelder (1986), the NW-oriented "cross-fold" joints are extension fractures formed by abnormal pore pressures in response to NW-directed tectonic compression during the Alleghanian Orogeny. The orthogonal strike ("release") joints are thought to develop later during regional uplift aligned with NE-oriented residual compressive fabric. The NE strike-joint set may not be as well developed and may waver more in direction and linearity as seen in the gentle sinuous patterns at outcrop #5 and along the “streets” of Little Rock City. Joints can also be affected by changes in lithology and bedding as suggested by the frequent overhangs of the upper channel deposits at the top of the blocks. Apparently these joints either did not readily propagate through the more varied (more permeable?) channel bedding in places or did so at a different spacing and/or direction. Similar effects are can be seen in shale/siltstone/sandstone sequences elsewhere (Engelder, 1986).

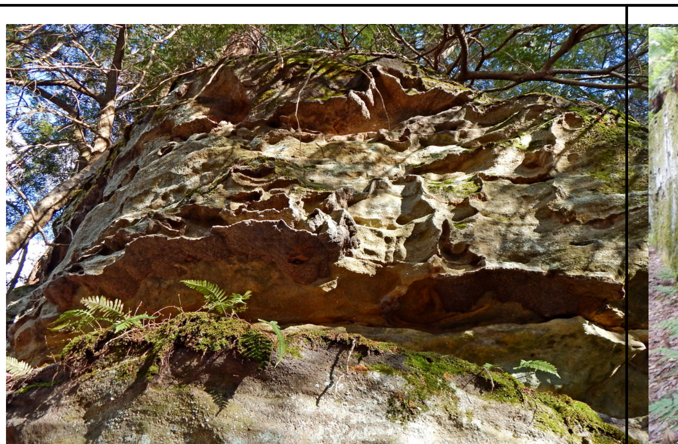
An interesting observation during collection of paleocurrent data was the frequent alignment (within a few degrees) of the true (maximum) dip of cross-bedding with the trend of the main joint set. What seemed like a helpful coincidence can be explained by the alignment of the NW joint set and its formative compressive stress field (as noted above). Tectonic strike (NE trend of plate boundaries/orogeny/mountains) would be roughly normal (like strike joints) to the maximum tectonic compression as should the corresponding paleoshoreline/basin. Fluvial/tidal channels and deposits generally trend normal to shore (down-paleoslope) and are often exposed in cross-section on NE-oriented strike joint surfaces. Marine (tidal and wave) deposits usually trend toward shore, generally SE which is a dominant flow direction as often displayed by cross-beds on cross-fold joint surfaces in this sequence.

While these Upper Devonian joint sets formed during the Pennsylvanian Period, similarly-oriented tectonic plate collisions (e.g., Taconic and Acadian orogenies) earlier in the Paleozoic yielded similar tectonic strikes/mountain ranges, alluvial plains, shoreline strikes, and depositional basins. Pettijohn (1975; p. 520) noted the stability of many paleocurrent systems through time and in particular, from Ordovician to Pennsylvanian time in the Appalachian basin. Numerous paleogeographic studies document fairly consistent paleocurrent directions and paleoslopes, generally NW-oriented as first recorded by Hall (1843) in the Ordovician Medina sandstone. In pioneering paleogeographic work, Hall deduced beach deposition, strandline orientation (NE-SW as noted is common throughout most of the Paleozoic Appalachian basin), and ocean/wave direction (NW) from oriented fossils, current scours, heavy mineral “clouds”, ripple marks, and swash marks on bedding planes in a building stone quarry (once great “outcrops” but all but extinct) near Lockport.

Iron Seams

The “iron ore” (hematite) seams of Hall (1843) are red to black in color, 1-3 cm thick, usually sub-horizontal but often smoothly contorted and commonly crosscut bedding. The seams appear most common higher in the sequence and in close association with fluvial/deltaic channels/redbeds and plant remains (Fig. 2 - outcrop #1). In the dune field (outcrop #7), iron seams cover several vertical joint surfaces (Fig. 3). And rare cylindrical shapes (10-20 cm in diameter) are suggestive of hollow logs.

With terrestrial input of iron via streams, precipitation of hematite where reduced iron-rich porewater mixed with oxygenated water might account for some seam occurrences such as in redbeds. The joint plane seam occurrences must have formed during or after joint formation. Jointing involves extension fracturing via abnormal pore pressure generated by tectonic compression (Engelder, 1986). Iron-rich porewater seems possible but the source of large amounts of reduced iron is unknown. However, the high porosity and permeability of this conglomerate could have facilitated later fluid migration. Together with hematite replacement of quartz cement in places, iron seam formation at depth after lithification, and during or after joint formation is indicated.




Iron Seams (Fig. 2 - many sub-horizontal seams in redbeds; Fig. 3 - vertical on joint planes in dune field)

Decaying plants provide localized reducing environments in sediments where dissolved metals such as iron may precipitate. Berner (1980) noted that whereas fresh/brackish waters (e.g., estuaries/deltas) are low in sulfur, plants are a source of sulfur for pyrite precipitation. Klein (2017; pers. comm.) noted that Devonian plant remains are typically pyritized and upon weathering, yield limonite/goethite and hematite as seen at several outcrops. Since pyrite is stable under reducing conditions at depth and oxidized iron (Fe^{+3}) is not readily reduced or mobilized, a direct iron source from plants appears unlikely. However, iron-rich solutions are evident; iron was sequestered as pyritized (FeS_2) plants and at some point, as hematite seams (Fe_2O_3) formed perhaps in response to undefined chemical and/or pressure gradients. In somewhat similar but more varied occurrences in Jurassic sandstones, Chan et al. (2000) thoroughly reviewed iron mobility and reactions (including Fe reduction reactions with hydrocarbons) and proposed that mixing of fault-related saline brines with shallow, oxygenated groundwater accounted for the precipitation of iron and manganese. With the Salamanca iron seams, the apparent formation at depth after lithification suggests unusual conditions perhaps related to tectonic stresses, faulting, and/or brine migration. Careful mapping and petrographic, chemical, and x-ray analyses of the iron seams may provide clues on their origin.

Previous Work and Stratigraphy

James Hall provided the first scientific descriptions of these rocks ("the conglomerate") as part of the multi-year Geologic Survey of New York (1839-1843). Working in western and central NY (the 4th district), Hall's descriptions and interpretations of some sedimentary structures (e.g., "diagonal lamination" and "ripple marks") and depositional environments (e.g., Medina Sandstone beach) were among the earliest recorded in scientific literature.

Hall's (1843; p. 285-290) conglomerate description (which is difficult to improve upon other than adding "well-rounded" to pebbles) of what at the time was apparently the premier rock city (and perhaps still is) follows below:

> *"The conglomerate consists of a mixture of coarse sand and white quartz pebbles, varying from the size of a pin's head to the diameter of two inches. They are generally oblong, or a flattened egg shape. Some of these are of a rose tint when broken, but white upon the exposed surface. Pebbles of other kinds are very rare in the mass, though red and dark colored jasper are sometimes found.*
>
> *This rock in the Fourth District occurs in outliers of limited extent, capping the summits of the high hills toward the southern margin of the State…From its position, it has been much undermined; and*

separating into huge blocks, by vertical joints, which are often many feet apart, the places have received the name of ruined cities, Rock city, etc.

There are several points in Cattaraugus County where the conglomerate is very well exposed upon the tops of the hills. The best known of these is the "Rock City," about seven miles south of Ellicottville (present-day Rock City State Forest)*….The sketch* (shown above on the title page) *represents a few of the immense blocks at this place, with the passages between them. The large trees which stand upon the top, have often sent their roots down the sides, where they are sustained in the deep soil, supporting the huge growth above upon an almost barren rock.*

The masses present the same features as before described, and offer fine exhibitions of the diagonal lamination and contorted seams of iron ore. The rectangular blocks are from thirty to thirty-five feet in thickness, and standing regularly arranged along the line of outcrop, present an imposing appearance, and justify the application of the name it has received."

The Salamanca Conglomerate is one of several conglomerate members of the Cattaraugus Formation of the Upper Devonian (late Fammenian) Conewango Group (Tesmer, 1963, 1975). First described by Hall (1843) as a single widespread unit, "the conglomerate", Carll (1880) named the Salamanca conglomerate and proposed correlation of several similar beds. Glenn (1902) likewise correlated several conglomerate beds and traced the Wolf Creek conglomerate (a very similar cross-bedded unit of sand and discoidal pebbles overlying "Chemung" beds) and the Salamanca conglomerate from the Portville/Olean area into the Salamanca quadrangle. Clarke (in Glenn, 1902) in a very prescient interpretation, cautioned Glenn about unconformities that rings true today: *"...these sand reefs constantly display indications of deep decapitation due to shifting of bars and change of directions of currents, or a modification by heavy tidal flow on a shelving coast."* Other stratigraphic work (e.g., Caster, 1934) was summarized comprehensively by Tesmer (1975) who concluded that conglomerate correlation is difficult and uncertain due to limited and separated outcrops, glacially-derived cover, probable facies changes, and possible structural complications (e.g., slight dip changes/folding/faulting). Tesmer (1975) tentatively placed the Salamanca member in the middle of the Cattaraugus formation, following Glenn (1902) who had mapped the Salamanca member well above (~ 60-70 m) the basal Wolf Creek member in the Olean area.

Baird and Lash (1990) noted some progress with correlation of the Panama Conglomerate member with the LeBeouf Sandstone in Chatauqua County and also the need to locate and observe the upper and lower contacts of these conglomerate units in order to place them in geological context (this study offers glimpses). Smith and Jacobi (2006) placed the Salamanca conglomerate at the base of the Conewango Group which puts it between the Wolf Creek conglomerate (type section near Olean, NY) and the westernmost Panama conglomerate (type section at Panama, NY). Collectively then, these units may represent the furthest preserved shoreline advance into the Devonian Catskill Sea of New York.

<u>Discussion</u> - Considering another ~ 200 meters of mostly marine Cattaraugus sedimentation above the basal conglomerate(s) (e.g., Wolf Creek), and the repetitive oscillating lithofacies (apparent shorelines) of Tesmer (1963), and the repeated T-R cycles of Smith and Jacobi (2001) in the Canadaway Group, and general observations at LRC, perhaps Tesmer's (1975) caution over conglomerate correlation is well-founded. If these very coarse, relatively thin (5-10 m, except for Panama; up to 20-25 m) units with sharp contacts represent the main transporters of sediment to the aggrading basin (e.g., as rapidly prograding deltaic complexes), one might expect these units to be vertically staggered throughout the section due to fluctuating sea levels. Larger more stable deltaic systems from larger streams (e.g., the deltas of the Cretaceous seaway – vanChappelle et al., 2016) would likely be much thicker and less influenced by small-scale relative sea level changes and more correlatable. But it seems likely that these conglomerate units scale with the streams/distributaries that formed them (fairly small channels; tens

of meters wide and at most a few meters deep) and offer glimpses of the shoreline as it oscillated through time and space with changes in relative sea level and sediment supply. Rapid progradation, evident in this study, moved the shoreline west-northwest which was followed by a marine transgression and shoreline retreat. Thick marine deposits appear to sandwich the Salamanca as glimpsed at LRC and in scattered marine outcrops nearby. Shoreline re-advance, if it occurred/reached this area after the inferred marine transgression, would likely have been well-separated vertically from the Salamanca.

By some estimates (e.g., Dennsion, 1985), the Catskill Sea shoreline beat a transgressive retreat back toward the Olean area by the close of the Devonian perhaps completing a halting but largely continuous loop through late Devonian time. Glenn (1902 map) depicts the Wolf Creek (first apparent Cattaraugus shoreline advance; type section near Portville) and the Salamanca placed about 60 - 70 m above it and overlain by the Oswayo shale (the uppermost Conewango Group/last Devonian formation) and the Olean conglomerate (Lower Pennsylvanian; w/larger, more spheroidal quartz pebbles (wider quartz veins unroofed?...a story for another day). Glenn (1902) described the Salamanca here as a hard gray sandstone (10-15 feet thick) which becomes coarser and thicker and passes westward into a massive conglomerate (such as at LRC). Once quarried extensively near Olean ("Mt. Hermon Sandstone"), it is a medium to coarse-grained buff to gray sandstone with occasional small quartz pebbles, medium-bedded (20 – 40 cm) with some layers speckled with oxidized iron (pyrite?) and pierced by many prominent vertical "fucoid" trace fossils. Given such a drastic facies change (and opposite trend to that of the prograding Wolf Creek which is usually very coarse and massive in this vicinity), the Salamanca here may represent a retrograding/stalled shoreline, perhaps the last Devonian shoreline in New York. Ongoing work will attempt to elucidate the relationship of these Devonian conglomerates and the paleoshoreline through time.

Relative Sea Level Changes – Gradual & Abrupt Examples

Tesmer (1963; Fig. 16) portrayed a coarsening-upward Cattaraugus formation with about six alternating repetitions of lithofacies (gray siltstones - buff sandstones – conglomerates). While generalized, his lithofacies curve suggests small-scale oscillations in relative sea level (T-R cycles) that are more pronounced and frequent than in underlying formations. Smith and Jacobi (2001), with detailed sedimentological work at 1200 outcrops, refined parts of the Upper Devonian T-R curve based largely on shoreface occurrences in the Canadaway Group of Allegany County. They were able to show small-scale sea-level fluctuations resulting from the separate effects of eustasy, syndepositional faulting within the Clarenden-Linden fault complex, and general tectonic subsidence with the conclusion that sea level curves inferred from local foreland basins may have a stronger tectonic signal than previously understood.

The Salamanca shows an apparent abrupt deepening at the top of the caprock. At outcrop #5, very coarse (up to 60+ mm) fluvial/deltaic-deposited pebbles are overlain by fossiliferous wave-rippled thin-bedded sandstones (the first fossil occurrences in the sequence; molds of *Productella sp., Camarotechcia sp., and Crytospirifer sp.*). Whereas wave/tidal ravinement (largely limited to reworking) along the caprock is evident in places, mud rip-up clasts (a terrestrial/fluvial source-the only mud observed in the sequence) within the matrix of some of the largest clasts (including some large sandstone and metamorphic clasts; their first significant appearance) observed within the caprock precluded significant disturbance. Given then the absence of a preserved transgressive foreshore/shoreface sequence, a fairly rapid depth change of at least 10 m is interpreted. Subsidence was likely the main cause especially in light of the copious sediment supply/vigorous fluvial input (which appeared to be expanding at the top of the sequence), energetic marine processes (tides and waves), and active progadation which could likely keep pace with relatively small-scale and slow eustatic fluctuations. Bishuk et al (2003) noted a similar abrupt deepening (with inferred 10-15 m of subsidence) in a Sonyea Group (Frasnian) coastal sequence where shallow-marine hummocky cross strata overlie terrestrial paleosols. Other examples of inferred rapid subsidence include abrupt transitions of hummocky cross strata (upper shoreface) to interbedded mudstones/thin siltstones (offshore) in the Frasnian West Falls Group (Craft and Bridge, 1987) and stacked shorefaces in the Canadaway Group (Smith and Jacobi, 2001, 2006).

PALEOGEOGRAPHY - GEOLOGIC SETTING

A paleolatitude of 25-30 degrees south (Fig.4) and a warm, seasonally wet-dry climate has been posited for the Upper Devonian of New York (e.g., Woodrow et al., 1973; Scotese, 2000). Southeast trade winds likely prevailed but the Acadian highlands presented a rain shadow (Woodrow, 1985). However, abundant rainfall would be expected from postulated monsoonal circulation (Witzke, 1990, Streel et al., 2000, Smith and Jacobi, 2006) perhaps similar to the present-day Indian Ocean/Indian subcontinent. Climate is a primary control on source-to-basin sediment flux and in warm climates, siliciclastic flux is greatest under highly seasonal rainfall (Cecil, 1990).

Given the likelihood of monsoonal rainfall, frequent floods, episodic hurricanes (Duke, 1985; Craft and Bridge, 1987; Baird and Lash, 1990; Smith and Jacobi, 2006) with possible storm-flood (Collins et al., 2016) and storm-tide coupling, and evolving plants which paradoxically may have increased weathering rates in places (Berner, 1997), significant weathering and transport of sediment to the Catskill Sea would be expected. In addition, pulsed orogenesis in the source area (the third collisional tectophase of the Acadian Orogeny; Ettensohn, 1985) likely increased stream/erosional gradients, significant fluxes of sediment into the basin, and basin subsidence in response to tectonic/sediment loading on the crust.

Figure 4. Late Devonian paleogeographic map – study area highlighted (modified from Blakey, 2017 and Zambito, 2011; shoreline extended into western NY; all boundaries approximate)

Basin Subsidence/Deposition Rates

The average subsidence rate in the Catskill foreland basin was nearly an order of magnitude higher in the upper Devonian than the middle Devonian (Faill, 1985) with deposits as thick as ~ 2000-3000 m over the ~ 14 million year duration of the Upper Devonian. In western New York, Faill (1985; Fig. 7) showed an estimated subsidence/deposition rate of ~ 100 m/million years (or roughly 2 million years to deposit the ~200 m Cattaraugus Formation). Coupling that rate with an average shoreline advance of ~ 30 km/million years (Dennison, 1985; Fig. 4), gives a rough volumetric deposition rate of three million cubic meters per million years per linear meter of prograding shoreline (or 3 cubic meters of sediment/meter of prograding shore/year). Very approximately then, for a 4 km swath of paleo-coast (such as at LRC), an average of 12000 cubic meters of sediment might be deposited per year. Sediment influx was likely much greater, on average, at prograding deltas which was then partly or mostly redistributed along and offshore by marine processes.

Shoreline

The coastal zone of the Devonian Catskill Sea varied in space and time as shown by the varied interpretations of shoreline deposits. Coastal paleoenvironments included deltas, distributary channels/mouth bars, tidal channels/flats, mud flats, and beaches (e.g., compilation in Sevon, 1985). Despite a number of early tidal interpretations, an assumption persisted that the probable tidal range/energy was low. But seminal work such as Johnson and Friedman (1969; tidal channels/flats) and Rahmanian (1979; tide-dominated delta) and tidal modeling by Slingerland (1986) and Ericksen et al. (1990) which suggested at least mesotidal (2 m -4 m) range, recognition of tidal coastal deposits increased over time (e.g., Bridge and Droser, 1985; Bridge and Willis, 1988; Bishuk et al., 1991, 2003; Duke at al., 1991; Willis & Bridge, 1994; Prave et al., 1996).

In a study with specific applicability to LRC, Slingerland and Loule (1988) documented a tide-dominated shoreline (tidal channels/flats/shoals/estuaries) with a wave-dominated (sand ridges) offshore in a shore-parallel, time-equivalent (mid-Frasnian) transect through central Pennsylvania. They posited that nearshore circulation was to the SW (clockwise), estimated tidal range was high mesotidal, and that three major clastic dispersal systems (drainage basins) existed across Pennsylvania. They also noted that meandering fluvial deposits capped all sections studied and that a lack of mouth bars and levees was attributed to strong tidal currents (like at LRC).

In a comprehensive synthesis of Devonian Catskill alluvial and coastal deposits, Bridge (2000) noted several coastal features in common: *"(1) sandy, tide-influenced channels; (2) shallow bays and tidal flats where mud and sand were deposited; (3) rarity of beaches; (4) storm-wave domination of the marine shelf. Much of the variability in the deposits across the area could be explained within the context of a wave- and tide-influenced deltaic coastline with a tidal range that varied in time and space."*

Regarding variations in tidal ranges, Reynaud and Dalrymple (2012) noted that since tides interact strongly with shelf and coastline morphology, changes in relative sea level can have a profound effect on tidal currents and deposits. Tidal resonance (amplitude strength) varies with shelf width (i.e., highest at increments of one-quarter of the tidal wavelength) and is directly affected by changing sea levels. They stated that: *"The increase in tidal influence can be geologically instantaneous in situations where the geomorphology changes rapidly. This was the case in the Gulf of Maine-Bay of Fundy system, which changed from microtidal to extreme macrotidal over a period on only a few thousand years."* Short-term changes then (e.g., tectonic or climate-driven sea-level variations) can bring about rapid change, *"potentially causing an alternation between tidal and non-tidal deposits"* and *"different parts of the transgressing sea can become resonant at different times"*. Also, once tidal resonance has been reached, further increases in sea level often result in a decrease in tidal influence. They suggested, as possible examples, abandoned tidal dune fields preserved beneath North Sea muds and tidal sandbodies in the Devonian Castkill Sea. They cited Ericksen et al. (1990) for the latter, who did not provide specific examples but the Salamanca tidal dune field at LRC is a possible example of decreasing tidal influence with wave-truncated dune tops overlain by channel deposits.

Sediment Sources & Dispersal Systems

Based on the inferred position of the Acadian orogen (Faill, 1985), source areas were likely located about 400 km to the southeast (cf. Pelletier, 1958) during Fammenian time. Weathering and erosion of actively-rising mountains produced detritus (including tabular vein quartz gravel) that was conveyed by streams to the foreland basin. As the shoreline advanced, drainage networks continually expanded and likely interacted to varying degrees. Sevon (1985) depicted up to six "sediment dispersal systems" which could have affected western NY, Slingerland and Loule (1988) noted three major drainage systems, and Boswell and Donaldson (1988) posited five stable drainage systems with large trunk streams for the Fammenian of West Virginia. The size of these drainage basins and streams are difficult to gauge but given an alluvial plain of at most 400 km, these were not the large continental rivers and deltas of today. Bridge (2000) noted that Catskill river channels were smaller near the coast (i.e., sinuous, single-channel rivers, tens of meters wide, maximum depths of 4 - 5 m, sinuosity of 1.1-1.3, mean bank-full flow velocity of 0.4 - 0.7 m/s) and perhaps distributive (delta-related). With increasing distance from the coast, slopes increased, rivers became wider (up to hundreds of meters), deeper (up to 15 m), coarser grained, and possibly braided.

Sediment - Sand & Pebbles

Other than localized rip-up clasts, no mud-sized sediment was observed. Quartz sand ranges in size from fine to very coarse, is sub-rounded to sub-angular, and composed largely of clear monocrystalline quartz. Clear quartz is mainly derived from intrusive plutonic rocks such as granite; such crystals are generally < 1 mm and are the source of most quartz sand. Cloudy polycrystalline quartz (the stuff of pebbles) predominates in coarser (1-2 mm) grains. Sand lithology is +95% quartz with occasional opaque grains including magnetite. Fine-grained magnetite comprises a very minor overall component (<<1%) of sand but it may concentrate locally along laminations in places. Bagged samples of disaggregated sand obtained from nearshore marine, beach transition, and channel deposits were magnetically-separated; all showed trace amounts of magnetite with channel deposits containing somewhat higher amounts. Since the specific gravity of magnetite (5.18 g/cm^3) is nearly double that of quartz, fine (0.125-0.25 mm) grained magnetite sand is roughly the hydraulic equivalent of medium (0.25-0.5 mm) quartz sand. At the shoreface/foreshore transition and in foresets of some dunes, dark-colored laminations and streaks occur. However, where samples could be obtained (e.g., moss-weathered outcrops), magnetite was rare. Magnetic separation showed partial black coatings on quartz grains and separated black flakes (magnetic attraction varied but mostly slight; possible hematite?).

Cross-stratified medium to very coarse sand dominates much of the sequence and is frequently interbedded with single or multiple layers of discoidal pebbles which usually conform to bedding and accentuate sedimentary structures.

PEBBLES

Perhaps the most interesting geological feature of the Salamanca conglomerate (in addition to cross-bedded monoliths the size of houses) are the ubiquitous well-rounded discoidal vein-quartz pebbles. The Salamanca is classified as an orthoquartitic conglomerate since the pebbles are lithologically and texturally mature. The milky polycrystalline quartz pebbles range from ~ 2 mm to 60+ mm (very fine to very coarse pebbles), average ~ 8-10 mm in size and are oblate ("flattened") ellipsoids in shape. Pebble lithologies are +98% quartz with minor amounts of red jasper and rock fragments. Shallow pits and fracture traces are evident on the surface of many pebbles. Most pitting is likely related to point-contact pressure solution upon burial which probably provided much dissolved silica for this well-cemented unit. Some surface ornamentation may be impact-related such as possible percussion marks on beach clasts (Allen, 1970) and V-shaped pits. The milky/cloudy nature of the polycrystalline quartz pebbles derives from microscopic fluid inclusions which disperse light. Fluid inclusions are consistent with a hydrothermal origin where silica-rich fluids were likely emplaced under pressure and crystallized rapidly in fractures of an active orogen source zone. Uplifted older vein quartz deposits, formed in the same fashion, are also possible.

Pebbles often conform with and accentuate sandy stratification and hence are very helpful in defining sedimentary structures and paleoflow directions, and assessing paleohydraulics. However, in beds where pebbles dominate (e.g., minor sand matrix, clast-supported "open framework gravels" such as common in channel bars and fills), stratification may be crudely developed and difficult to interpret. Pebble imbrication can be helpful such as the common orientation of oblong pebbles transverse to flow but pebble inclination may be ambiguous. Jumbled/chaotic/unstable pebble orientations are common especially in channel deposits which suggests disequilibrium with waning, rapidly depositing flows ("unsteady" tidal currents) and sediment-choked channels.

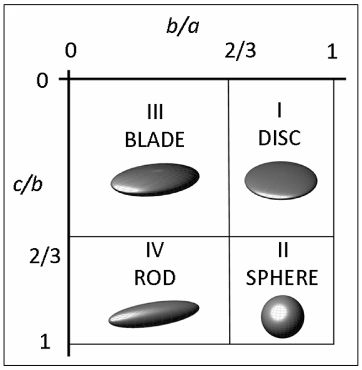
Pebble Shape/Origin

The origin of the distinctive discoidal pebble shape has been ascribed to beach-shoreface abrasion since the Salamanca was named (Carll, 1880) and accepted over time by Glenn, 1902, Tesmer, 1975, and Miller, 1974, as cited by Baird and Lash 1990). While appealing, it is not clear how an entire population (billions?) of extremely durable quartz pebbles could be systematically and symmetrically abraded/flattened to yield co-planar sides and with a probable concomitant mass loss of up to 80-90%. Prolonged abrasion experiments show little mass loss for quartz pebbles after initial edge rounding (e.g., Krumbein, 1941; Keunen, 1956; Attal and Lave, 2009; Domokos, 2012). Also, at this locality the majority of Salamanca pebbles are found in channel deposits; beach deposits are not common. It seems clear then that the cloudy pebbles of vein-quartz derived their tabular shape from their origin in tabular quartz-filled fractures (veins) in the source area and rounding/smoothing during stream transport. As Pettijohn (1975) noted, the end-shape of sedimentary quartz is an expression of its initial shape.

Pebble Dimensions - Fractures/Fragmentation

Clast thickness is largely determined by the dimensions of tabular quartz veins in the source area. Caliper triaxial measurements of ~ 100 pebbles spanning the available size range yielded a C-axis (the short ellipsoidal axis) range of 1.5 mm to 16 mm which suggests veins of that size range in the source area. And that rather restricted thickness range suggests a rather consistent source area of narrow tabular veins of quartz (no rogue spheroidal clasts; other lithologies are less durable). And hydraulic (size) sorting during extended fluvial transport likely restricted the upper size limit (the majority of pebbles are < 15 mm; large clasts are sequentially sorted out; Pelletier 1958 showed an exponential decline in the Pocono Group). The sudden appearance of some very coarse pebbles (up to 60 mm and 25% different lithologies; sandstone and metamorphic clasts) at the top of the caprock suggests an unusual event or process.

The triaxial pebble measurements yielded axial ratios (B/A and C/B) that plot, for the most part as expected, within the Disc zone of the Zingg shape diagram (1935; in Pettijohn, 1975) (Fig. 5). However, smaller size ranges (< 8 mm) trend toward, and in particular, many 2-4 mm ("granules") pebbles, plot within the Sphere zone. The C-axes, while thin (1.5 mm – 3 mm), are still recognizable as parallel which suggests the same vein origin. These more equant shapes result when the A - B axes approach the C-axis in dimension.



← Figure 5

Figure 6 →

So the C-axis is essentially fixed (vein-pebble thickness = lowest common diameter); the A and B axes can get smaller due to breakage normal to C. The suggested mechanism is the greater susceptibility of thinner veins and clasts to weathering and fragmentation at the outcrop and in early transport in high-gradient streams. Fracture traces, often outlined by iron-oxide staining, are common and are generally normal to the two co-planar sides (A & B axes). Some pebbles show smoothing/rounding of fracture-parallel edges (missing chunks) which suggests fragmentation/smoothing occurred during transport (Fig. 6). Other pebbles show sharp-edged breaks which, if natural, suggest little transport after fragmentation. So planes of weakness would tend to focus breakage along the short "C" axis and the thinner the veins/pebbles, the higher the expected rate of disintegration (lots of thin veins/pebbles = lots of milky granules and odd shapes, e.g, irregular or roughly triangular, appear more common in small pebbles). The highest fragmentation rates during transport would be expected in near-source high-gradient streams where strong flows, a wide size range of particles in motion, and high impact velocities which, along with existing planes of weakness, would promote fragmentation (e.g., Attal and Lave, 2009).

Rounding

Experiments have shown that lithology controls abrasion rates (e.g., Keunen, 1956, Domokos et al., 2014). Quartz pebbles are extremely durable with fairly rapid rounding (Attal and Lave, 2009; Domokos et al, 2012) but little overall change in clast diameter ("virtually indestructible", Pettijohn, 1975; Southard, 2006). In an elegant series of experiments, modeling, and field studies (Domokos et al,. 2012; Miller et al. 2014) demonstrated that *"abrasion occurs in two well-separated phases: first, pebble edges rapidly round without any change in axis dimensions until the shape becomes entirely convex; and second, axis dimensions are then slowly reduced while the particle remains convex." The first phase occurs mainly in high-gradient, source-proximal streams, the second, in lower-gradient alluvial plains where size sorting due to stream hydraulics and lithologically-controlled abrasion prevails."* Coupled with the fragmentation process noted above, most sizing and shaping (fragmentation) and rounding (convex shaping) of pebbles likely occurs in near-source high-gradient streams whereas most hydraulic (size) sorting and abrasion (slight for quartz) occurs in lower gradient alluvial plain streams.

The Granule "Problem" - Pettijohn (1975) and Southard (2006) noted a general scarcity of very coarse quartz sand and granules (1 mm - 4 mm) in the rock record. The cause appears related to the fact that the most common sizes of quartz crystals in plutonic rocks, the source of most quartz sand, are mostly < 1 mm. The Salamanca and other Upper Devonian conglomerates have an abundance of coarse sand, granules, and fine pebbles likely due to an abundance of vein quartz and the processes noted above. These size ranges are more spherical than larger pebbles and readily transported by nearshore currents and are abundant in dunes and nearshore marine deposits. The general source of quartz grains can be roughly distinguished in field: Clear = monocrystalline plutonic sources vs. Milky = polycrystalline vein sources. The tannish granule/pebble layers ("grain striping") within the light gray sands of large dune foresets is a macro-example.

Paleohydraulic Estimates

By some measures, such as a simple fluid/particle force balance and frictional considerations, low-profile discoidal pebbles should be more difficult to entrain and transport. However, once entrained ("mobilized") in a current, the tabular clasts would likely settle slower as indicated by calculation of the Maximum Projection Sphericity (Sneed and Folk, 1958). Also known as Maximum Settling Sphericity, a range of pebble sizes averaged about 0.5 or equivalent to about twice the cross-sectional area of a sphere of equivalent volume which suggests slower settling of discoidal pebbles. Bradley et al. (1972) studied the effect of shape both in the field (Knik River, Alaska ; high-gradient glacial-meltwater stream) and in the laboratory. They detected downstream sorting of shapes, with platy pebbles being the most easily transported, then elongate pebbles (rollers), and more equant pebbles being the least easily transported. The different shape-sorting effects were attributed to particles moving by traction and by -suspension and hence closely related to flow strength and particle size.

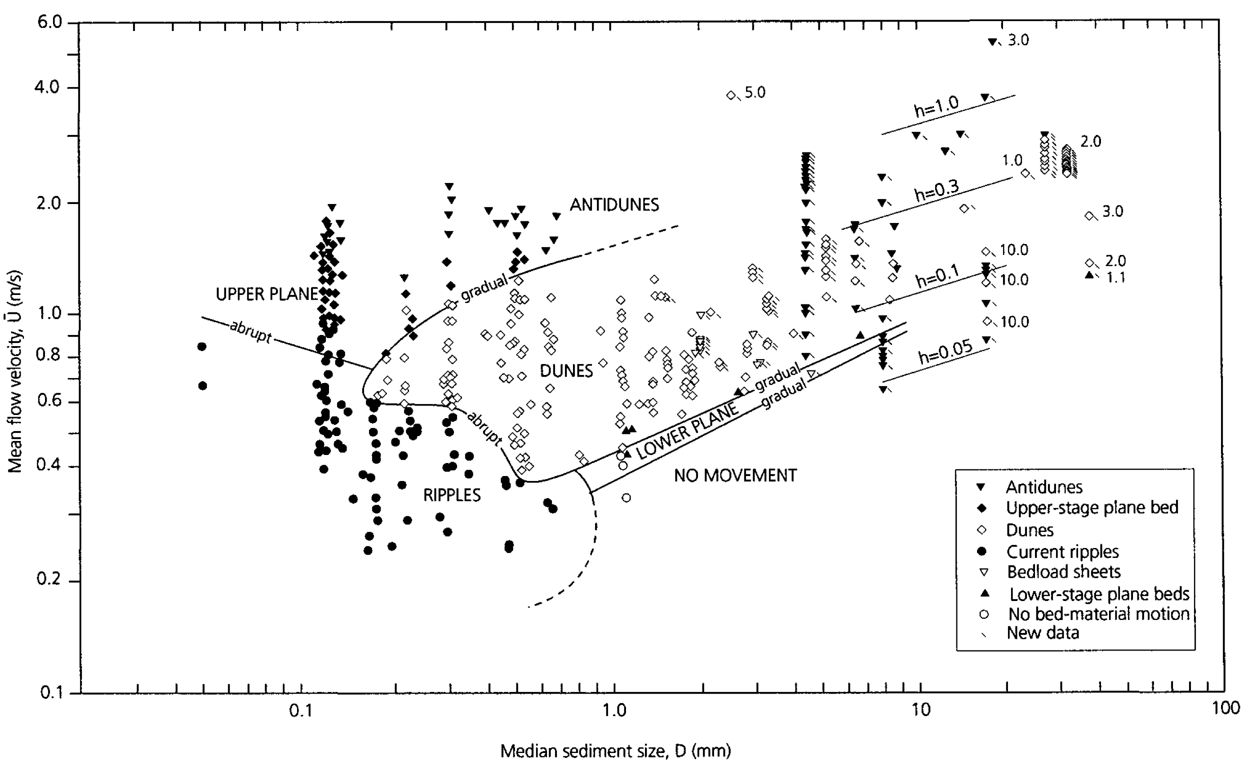


Figure 7. Bedform Existence Fields for Unidirectional Flows (Carling, 1999; redrawn after Southard and Boguchwal, 1990 and extended to gravel sizes, D ~ 33 mm). For general background, see Middleton (1977), Allen (1982), and/or Harms et al. (1982).

The bedforms most applicable to conditions at LRC are Dunes and Bedload Sheets. Dunes of various dimensions produce cross-strata which incline downcurrent and scale with depth, flow strength, and grain size. Ripples, smaller-scale (<= 4 cm) dune-like bedforms, require sand of < ~0.6 mm which is uncommon at LRC; ripples have not been observed. For sand sizes, a current of roughly 0.4 m/s to 1.0 m/s would be required to form dunes. For dunes composed of granules and fine pebbles (2 mm to 8 mm), a current of 0.6 to 1.5 m/s is indicated. Flows required for coarser pebbles (8 to 32 mm; limit of graph) are less clear given prominent disc shapes and data scarcity, but currents of 1 m/s to +2 m/s appear likely. Bedload sheets, low-amplitude bedforms which can transport a range of sediment are likely common at LRC but difficult to definitively identify; their existence field appears to coincide with dunes.

Possible Vein-Quartz Source-Area Analogues

Pettijohn (1975) and Baird and Lash (1990) noted that large vein quartz accumulations imply the destruction of large volumes of source rocks since quartz veins make up a just small percentage of normal lithosphere. However, vast amounts of vein-quartz pebbles transported within fairly limited drainage basins for at least 2 million years suggest an unusual lithosphere (an abundance of quartz veins) within areally-limited source areas. Hack's (1957) law indicates that a stream ~ 400 km long would have a drainage basin of ~ 20,000 km^2; the near-source catchment width is uncertain but likely on the order of 100 km. A similar sedimentation pattern (Olean/Pocono) continued in the Pennsylvanian Period from a similar source area (Pelletier, 1958; larger more equant pebbles suggest unroofing of thicker quartz veins).

A possible analogue for a vein-quartz source terrain is the Ouachita Mountains where more than 8000 meters of Paleozoic strata were folded during the Mid-Pennsylvanian Ouachita Orogeny. Innumerable steeply-dipping fractures, related to the major folds and faults of the region, controlled the emplacement of hydrothermal quartz (Miser, 1943; Engel, 1951). Another example and possible analogue of an abundant source area of vein-quartz as well as long distance transport of discoidal pebbles is the Miocene uplift in the southern Appalachians. As reported by Missiner and Maliva (2017), pulsed tectonism resulted in a surge in coarse siliciclastic sediment (including abundant discoidal vein-quartz pebbles of up to 40 mm in diameter) and long distance (up to 1000 km) fluvial transport. And the famous Witwatersrand gold deposit in South Africa (source of 50% of the world's gold for over a century) is a Precambrian fluvial conglomerate with discoidal vein quartz (~ 30 mm) pebbles (*let's go with this one!*).

DESCRIPTION OF CROSS-STRATA

Large-scale cross-stratification, the dominant sedimentary structure, is examined in detail and the overall depositional environments are reviewed below. But what's missing bears emphasis: unusual characteristics of this sequence are the near absence of preserved mud-sized sediment or trace fossils (perhaps small burrows in channels) or body fossils (but many fossils are present on and above the caprock). Strong currents, substrate mobility, and the apparent lack of organic matter likely presented an inhospitable environment, poor habitat, and poor preservation potential. The inferred high-energy coastal environment likely prevented deposition of fine-grained sediment which was transported offshore. Protected lower-energy coastal areas, such as back-barrier lagoons, muddy tidal flats, or fine-grained overbank fluvial/deltaic deposits, were not recognized in this sequence. Wave influence is pervasive but often subtle throughout this sequence.

Fossils – Plant remains are common in channel and associated deposits particularly within the deltaic sequence. With the exception of possible escape burrows in a channel base, trace and body fossils were not observed within the Salamanca conglomerate. However, abruptly overlying the caprock are finer-grained buff sandstones that contain an abundant brachiopod fauna and rich marine faunas are common in shallow marine deposits nearby. For example, an intact *Productella sp*. was found lying directly on the caprock seemingly in life position (Fig. 24).

Hall (1843) noted that fossils are extremely rare within the "conglomerate" citing 3 brachipod species in a sandy correlative of the Panama member. Tesmer (1975), citing the work of Butts (in Glenn, 1902) in the nearby Olean quadrangle, noted two brachiopod species in the Salamanca, Camarotoechia contracta and Crytosprifer sp? along with 13 pelecypod species, an ammoniod, and a gastropod.

Large-scale Cross-stratification

Cross-stratification, the most abundant sedimentary structure, dominates most outcrops. Individual set dimensions range over two orders of magnitude in scale (~ 0.05 m to +5 m). Most cross-strata are planar (straight- to slightly-sinuously crested = 2D type) with some trough (sinuously crested = 3D type) evident in channel deposits and upper shoreface/lower foreshore depsoits. Smaller forms may display bi-directional foresets in places (Fig. 8) but are more commonly organized in stacked co-sets (Fig. 9 ; cosets are ~ 0.75 m thick and show wave influence at the tops, e.g., centered on the 3' tape; Fig. 10 shows fine sand drapes, possible tidal influence and at the interface, wave ripples occur, then a 1 m thick cross-strata with angular toesets; small and large x-strata align shoreward; outcrop #1). Foresets are largely composed of grayish medium to coarse sand with variable interbeds of milky granules and pebbles ("grain striping" within large-scale foresets). Larger sets are generally coarser. Some channel bars and fills are composed in large part with pebbles (open framework gravel) that are crudely stratified or imbricated. The vast majority of paleocurrent data are cross-bed inclinations and range from 90^{o}– 150^{o} (mostly) with minor clusters at 40^{o} – 60^{o} and 220^{o} – 240^{o}.

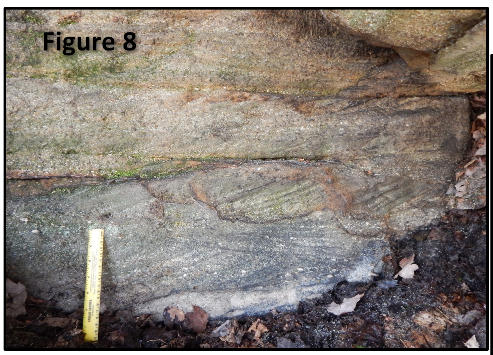


Figure 8

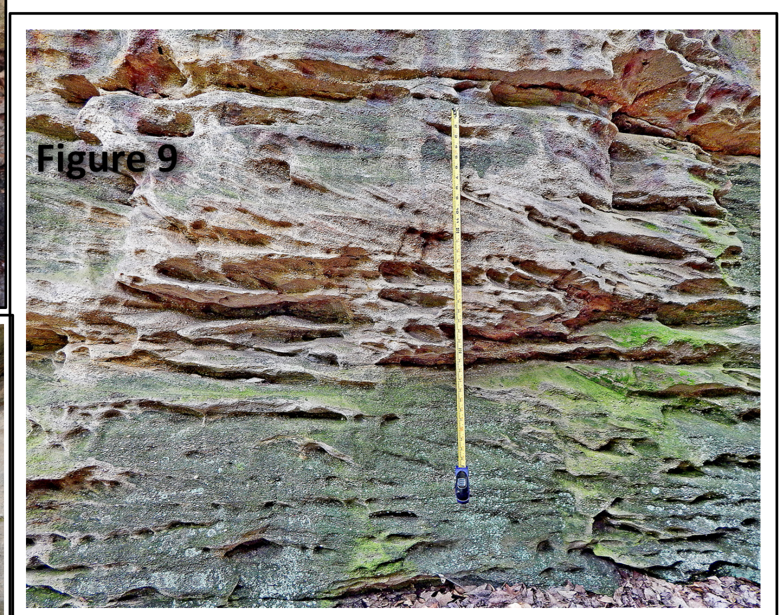


Figure 9

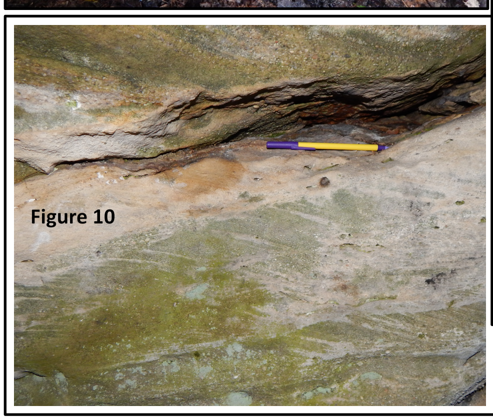
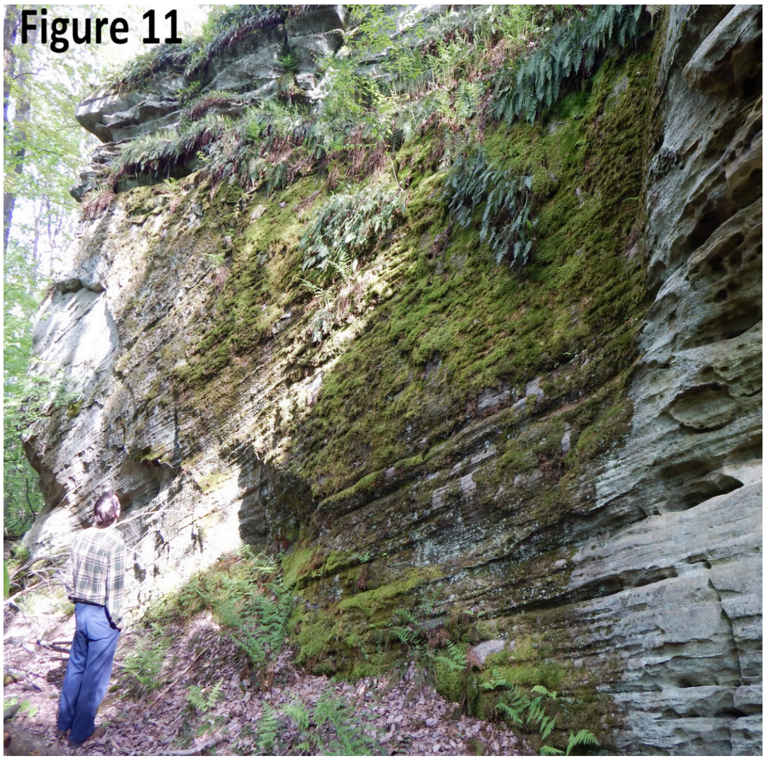


Figure 10

The largest cross-strata (0.50 – 5 m) increase in size and abundance from north to south across the outcrop belt. At the southernmost outcrops at "Little Rock City", large foresets may comprise ~ 75% of the outcrop exposures with dips of 20^{o} – 30^{o} , no obvious or major reactivation surfaces, and most toesets are tangential (Fig. 11; largest foresets ~ 5.5 m). Some foresets are traceable for +150 m across several blocks and a planar truncation surface at the top of the dunes shows wave influence (e.g., wave ripples with crests parallel to the paleo-shore). About 1 m of low-angle stratification overlies this interface followed with about 2+ meters of gray and red low-angle strata and channels to cap the sequence (Fig. 12; foresets at base, about 1 m, then as described above; note iron seams in redbeds).



An intact 2-3 m dune bedform (a "form-set" with foreset, topset, and stoss beds preserved) is unusual at this scale (Figs. 13-15; note the connected stoss & foresets just above 6' tape in Fig. 14). This form-set informs dune genesis: a medium ebb dune forms the base with directionally-opposed cross-strata aggrading vertically until one flow direction (100^{o} – apparent flood tides, downcurrent from the deltaic" sequence) prevailed about halfway up and coincident with a 15 cm dune (18 cm yellow ruler in Fig. 15-core"photo) which formed the crest of the avalanche face at that point. The uppermost stoss beds are continuous with the topset beds; some topset beds (bedload sheets) flow continuously into foreset beds in places. A small dune is also present in the topset beds but most of its 15-20 cm thick strata appear horizontal.

Figures 13, 14, & 15

Figure 15

Another interesting occurrence in the dune field are small convex “humps” or “piles” of granules and pebbles (average 4 x 10 cm but up to 8 x 20 cm) and spaced erratically along foreset beds as exposed on a vertical joint surface; cross-section along depositional strike showing horizontal foresets (“backside”) of a large dune. And two very coarse (mostly pebbles) and lower profile (~1.5 m) dunes with some opposed cross-strata were observed SE of the main southern dune field at LRC.

INTERPRETATION OF CROSS-STRATA AND DEPOSITIONAL ENVIRONMENTS

Hall’s (1843) explanation of “diagonal lamination”: *“...where the sand is carried on and spread over the surface, sloping off towards one side farthest from its origin. The next deposition covers this sloping side necessarily in the same manner, producing the oblique lines...”* was perhaps the first detailed account of cross-stratification (Allen, 1982); it describes the essential process of sand movement and deposition on an inclined surface. To embellish slightly, currents transport sediment along a gentle stoss slope to the bedform crest where repeated sediment avalanches down the steeper lee slope form cross-strata at or near the angle of repose. The resulting cross-strata are the depositional units formed by the migration of bedforms, dunes of various scales in this case.

Based on mostly shoreward- and some bi-directional-oriented cross-strata, the dominant currents were tidal and predominantly flood tides. Most sediment transport likely occurred during high spring tides of the bi-monthly spring-neap tidal cycle. Bedload transport rates scale roughly with the cube of the current velocity; if the flow rate doubles, bedload transport increases by a factor of roughly eight (Wang, 2012). A mesotidal range of 4 m appears to be a reasonable estimate; similar modern deposits/bedforms are produced in that range such as in the North Sea.

Based on paleohydaulic estimates noted above, the currents required to form dunes ranged from about 0.50 to 1.50 m/s. A similar velocity range (0.5 – 1 m/s) has been reported in the Dutch North Sea where very large simple dunes (like those at LRC) are actively migrating decimeters to a few meters per year (e.g., Tonnon et al., 2007; Passchier and Kleinhans, 2005; Stride, 1982). Allen (1982) reported that on the European continental shelf, sandwaves (large dunes) are found where tidal currents associated with spring tides range between 0.65 and 1.30 m/s. LRC dunes are somewhat coarser than modern examples and perhaps formed in somewhat shallower depths. A shoreface depth of ~ 10 m would conform with the dune height/depth ratio of 0.5 (Allen, 1982) for the largest (~ 5m) LRC dune (note that lower *h/d* ratios are common; Reynaud and Dalrymple suggest ~ 0.2). Tidal transport of coarse sand and pebbles at much greater depths may have been limited by the “littoral energy fence” whereby coarse particles are sequestered nearshore (Allen, 1970; Thorne and Swift, 1989). Even sand is rarely transported offshore by fair weather processes but evidence for Devonian hurricanes in the Catskill basin is strong and modern studies of sediment transport inform the past (e.g., Keen et al., 2012).

The large foresets on large “simple” dunes suggest strong very asymmetric tides. Allen (1980, 1982) depicted four general variants of “sandwaves” (what geologists now call “dunes” per Ashley, 1990) based on tidal current symmetry. Allen’s conceptual “sandwave”/dune generated by the most asymmetric tides (Fig. 16; large simple foresets in bottom frame; note the velocity asymmetry of U* critical, the threshold velocity to move sediment) conforms with the large dunes at LRC. Allen (1982) also depicted superimposed smaller dunes supplying sediment to the large foresets of a larger host dune.

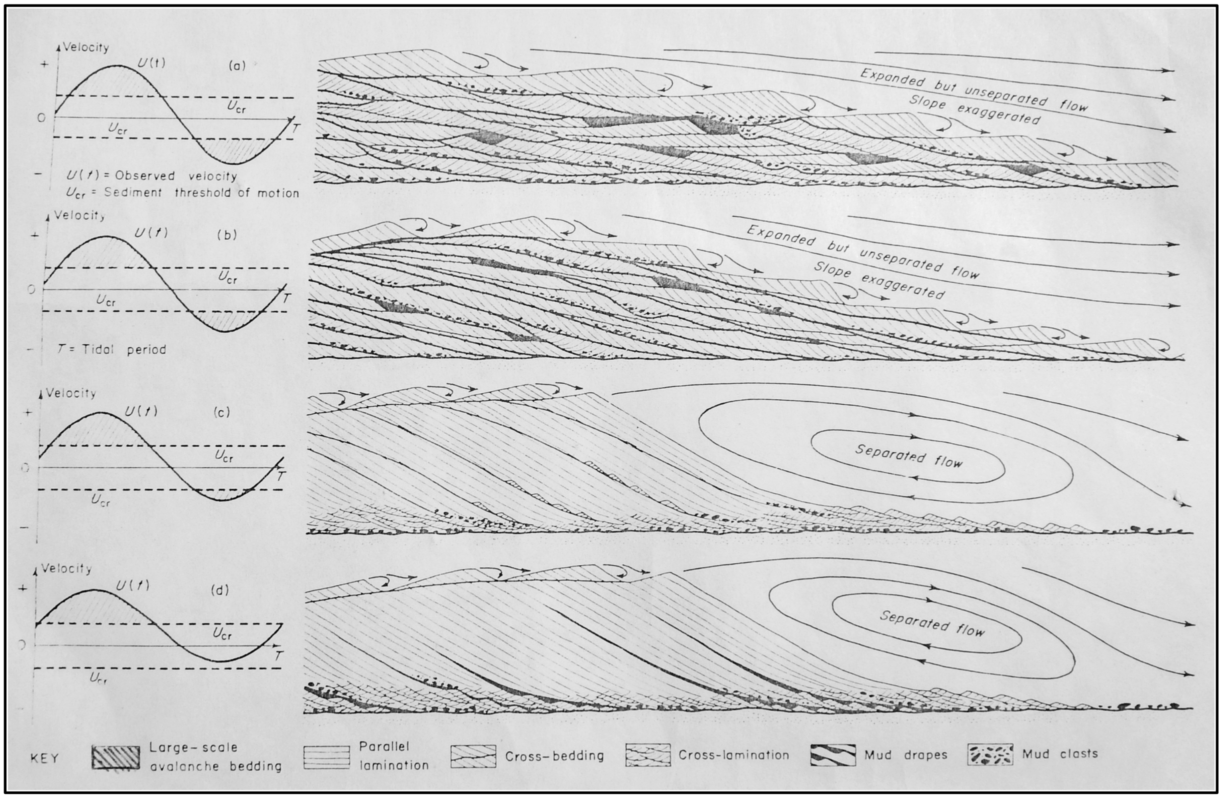


Figure 16 – Conceptual Model for generation of large dunes (Allen, 1982)

The “genesis” dune at the LRC dune field contains a small dune that appears to have “stalled” at the crest and reformed/”sharpened” it (center of Fig. 15; above 18 cm ruler); deposition continued along the aligned stoss and lee of both dunes (cf. Bridge and Demicco, 2008). Superimposed dunes pre-sort and transport sediment to and over large host dunes often in concert with bedload sheets. Pre-sorted wedges of sediment, as formed by smaller dunes (“trains”) advancing over the crest (”cliff”) of large host dunes, move down the flow-separated lee slopes often haltingly, by “grainflow” and finer sediment is distributed by “grainfall” from suspension (Reesink and Bridge, 2007; 2009).

One consequence is the scattered formation of lobate tongues of coarse sediment that may show inverse grading due to kinetic sieving. Otherwise known as “grain striping” (Reynaud and Dalrymple, 2012), a likely result of this process is shown in the dune cross-section along depositional strike (dune “backside”) with scattered and variable convex piles of granules and pebbles along the foresets. Harms et al. (1982) described the process of foreset avalanches at high sediment concentrations as oversteepened areas which slump in places and slide down the lee slope as long “tongues”. A slight scour or channelelized grainflow may form which then “debouches” with a slight positive lobe at the basal portion of the foreset. Some foresets and “grain stripes” at LRC nicely display these subtle structures in dip cross-section (Fig. 17) and the small granule “piles” on a dune backside noted above are interpreted as “grainflows” along the strike of dune foresets.

Figure 17

The largest cross-strata (0.50 – 5 m) increase in size and abundance from north to south across the outcrop belt which may indicate increasing water depth since dunes scale with flow depth. And the LRC dune field is downcurrent of the inferred delta complex which provided an abundant sediment supply and may partially explain the location of the dune field.

Form-set ("genesis") Dune

As shown in Figures 13-15, a medium ebb-dune formed at the base and small competing dunes aggraded vertically (or slightly in the ebb direction) until the flood (shoreward) tides began to dominate about where the small dune is perched in the middle of the bed. With an abundant up-current sediment supply, the flood tidal current began to dominate and the ~ 2–3 m dune began to migrate. In effect, the simple large dune has compound small dunes at its core/start and other superimposed dunes supplying and presorting sediment along with bedload sheets. In addition to a small dune in the topset bed, at least 2 topset locations show continuous strata between inferred bedload sheets and foreset beds. In a review of dune preservation, Reesink et al., (2015) noted that dune sets may climb due to local dominance of deposition over dune migration which generally fits this situation. But more specifically in this case, it appears that the localized balance between ebb and flood dune deposition aggraded a vertical core until the more dominant flood current and sediment supply tipped the balance toward large dune migration.

Wave-truncated Dunes

All of the largest dune (> 3 m) foresets at LRC appear to have been truncated ("beheaded dunes") horizontally at similar elevations (~ 3 m from the top of the sequence). Evidence of waves at this interface is common which suggests storm wave action which is well documented in the North Sea (e.g.,Terwindt,1971; Reynaud and Dalrymple, 2012). The overlying ~ 1 m of low-angle bedding is somewhat cryptic but the scale suggests lateral-accretion deposits of migrating tidal point bars.

Fluvial Channels/Bars

The final 2+ m of outcrop in the dune field contains robust channels and lateral-accretion w/some well-oxidized redbed deposits (Fig. 12) and some seaward-directed paleocurrents which are interpreted as fluvial meandering stream channels and point bars (Slingerland and Loule, 1988) noted a similar transition). This upper sequence appears to have shallowed upward probably from both lowered relative sea level (the beheading appears unique and the dunes never recovered) and active progradation. This upper channel sequence also overrode the rest of the sequence; the top of the beach at the north face and the tidal point bars are at similar elevation. If the beach deposits were largely preserved, then mean sea level should have been about ~1.5 m below the channels (~ half the foreshore) and roughly the same for the inferred tidal channels in the dune field. Other than crisscrossing channels within the delta sequence, surprisingly little incision is evident anywhere in this sequence but basal exposures are limited).

The remaining facies associations/depositional environments are summarized briefly with mention of outstanding issues/ongoing work:

Shoreface to Foreshore (beach) to Deltaic/Fluvial Sequence

coarsening-upward sequence - (“north face” Outcrops #2 & 3 from base)

- ~ 1 m of thin-bedded (5-10 cm) wave cross-laminated strata; mostly buff, medium sand with some coarse sand, granules, and a few fine pebbles.
- ~ 3 m of amalgamated coarse-grained, large (10-20 cm x 50-100 cm), smooth-crested wave ripples with abundant pebbles (some apparent 3-D forms seem without analogues), interbedded in places w/thin fine-grained (rolling-grain) wave ripples; some trough/planar cross-beds near top.
- ~ 3 m of parallel/low-angle strata of gray interbedded coarse sand and pebbles (a mixed gravel/sand beach, Komar, 1998; shape sorting is common whereby flatter clasts are left higher on the beachface. Not observed here but ongoing work will evaluate). Bed inclinations vary somewhat due to tilted blocks but some ~ 5^{o} are seaward (a few degrees is expected). A fairly thick beach sequence and a coarse shoreface have been suggested as indicators of significant tidal range/influence).

Figure 18 is interpreted from the top (red lines ~ 3 m): Fluvial (deltaic?) channels/bars, foreshore (beach), foreshore (wave deposits). Figure 19 is interpreted as amalgamated shoreface deposits. Figure 20 is interpreted as mainly foreshore/beach deposits (all photos from outcrop #2).

Issues: The smooth-crested (“arched”) large wave ripples (noted above) probably form in a similar hydraulic regime as fine-grained hummocky cross-strata but published work (Leckie, 1988; Cummings et al., 2009) indicate such ripples should be sharp-crested in coarse sediment. Some smooth arches (10-15 cm high) weather out at some bedset boundaries at outcrop #2 and appear 3-D but not all follow strata. Perhaps sharp crests are smoothed or different bedforms arise under combined flow conditions as such with tidal interaction (e.g., Perillo et al., 2014; Passchier and Kleinhans, 2005); needs further description.

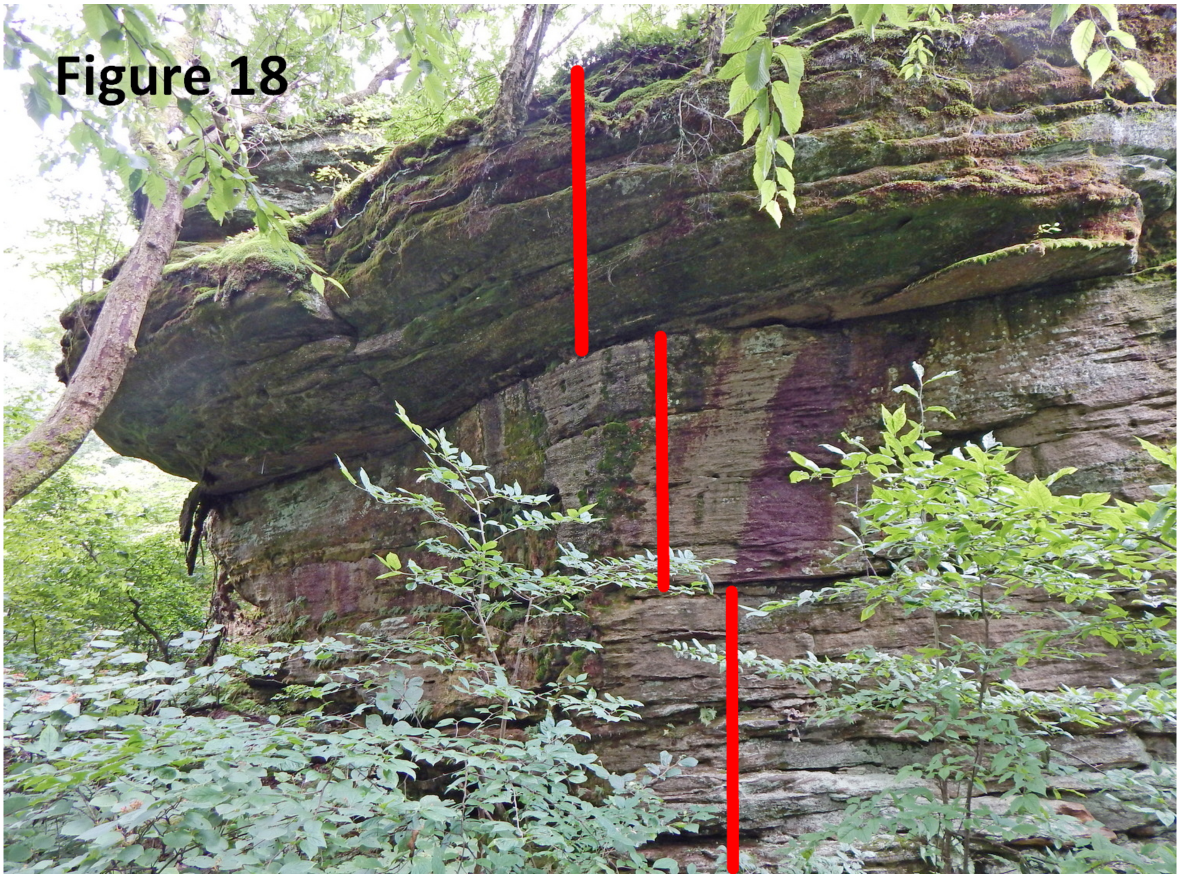


Figure 18

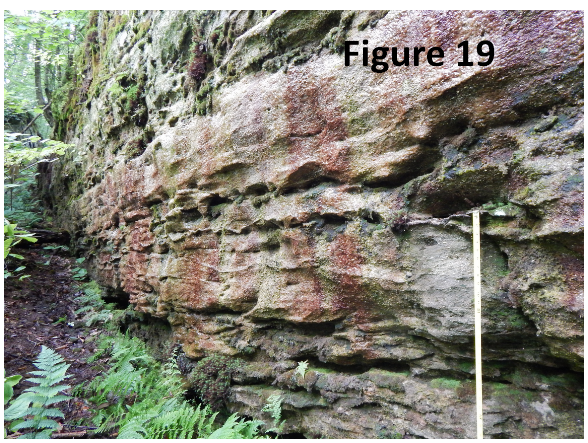


Figure 19

Figure 20

Prograding Tide-dominated Delta

Outcrops # 4, 5, & 6 - Coarse-grained distributaries, Tidal channels, Bars, and Shoals

- Outcrops #4 & 5 - abundant channels (meters to tens of meters wide, ~ 1-2 m deep) and channel point bars (coarse sand to pebble lateral-accretion deposits of tidal and delta distributary channels).
- Outcrop #6 - extraordinary exposures of channels/point bars with direct deltaic evidence (Fig. 26-27). The base of channel complex (three discrete ~ 2 m units or "storeys") directly overlies tan fine to medium-grained wave-ripple laminated sandstones which is interpreted as a distributary channel complex prograding over marine shoreface deposits.
- Cross-bedded strata of various dimensions (~ 0.05 m to +1 m) commonly arranged in cosets, some bidirectional (Fig. 21).
- Current indicators mainly directed shoreward (E-SE); some point bar tops and some truncation surfaces show wave influence (wave ripples).

As noted earlier, recognition of cross-strata in pebbly beds is difficult but most beds have crude bedding. Most of the bars appear to be point bars associated with meandering tidal channels and distributaries. The vertical dimensions of bars and channels are generally small (1-2 m) which reflect channel depths and the finer-grained bar tops may be wave-rippled suggestive of storm waves or slack tidal periods. Channel widths may approach 10+ m; an example at outcrop #5 appears to show back-and-forth lateral bar migration within a shallow channel while slowly aggrading vertically (perhaps slow subsidence due to compaction; Fig. 23).

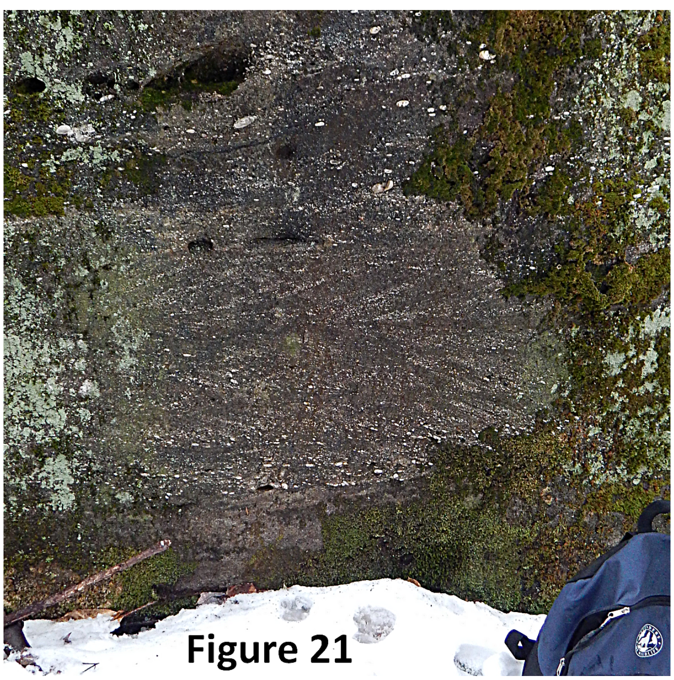


Figure 21

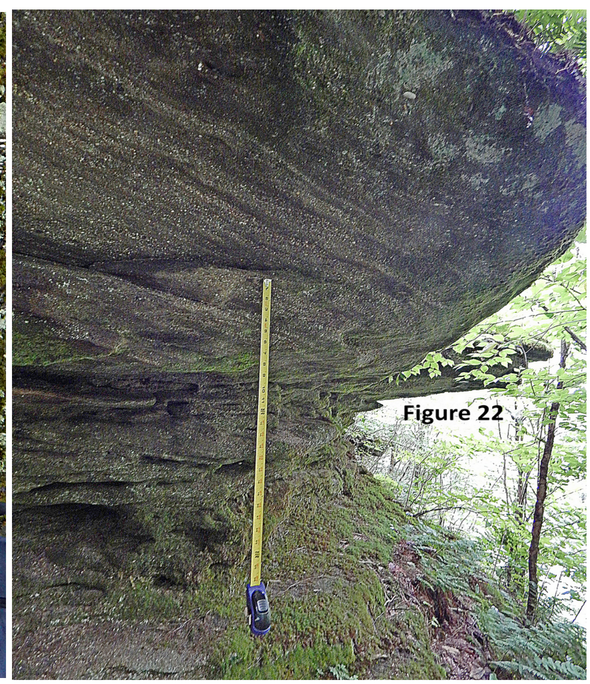
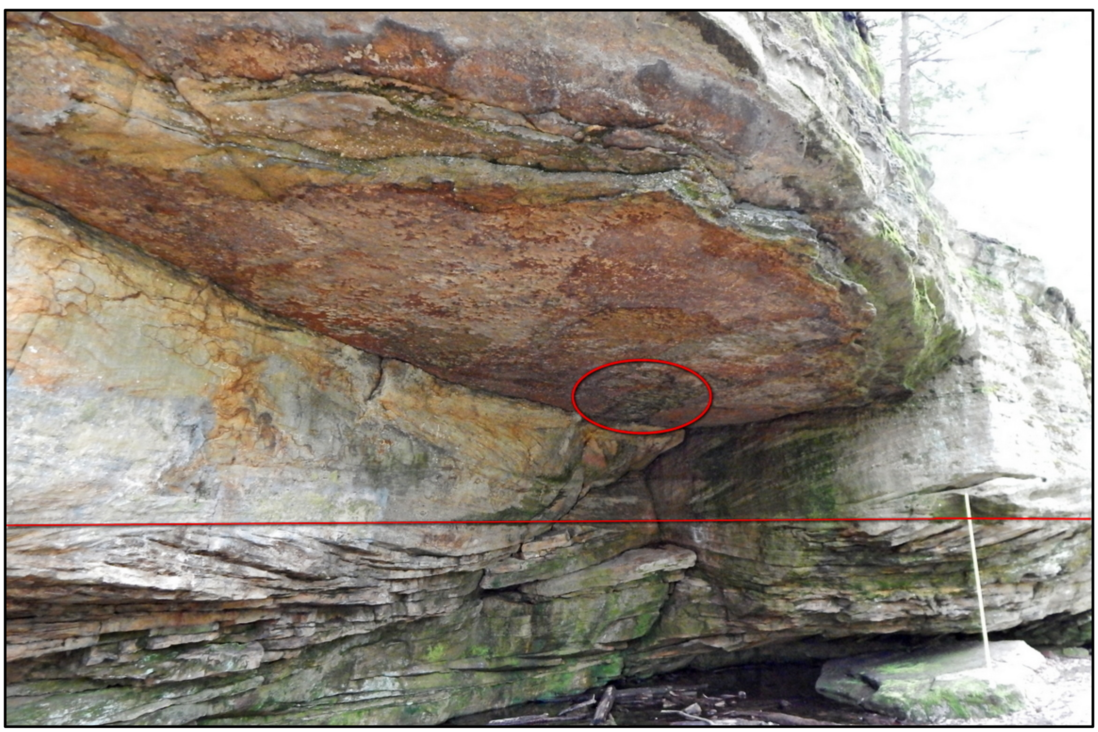


Figure 22

Figures 23 - (channels), **24** (*Productella* found on caprock), **25** (Caprock) - **Outcrop #5**



Figures 26 & 27 - – Outcrop #6 – Two westerly views of main channel complex - Multiple channels, marine deposits at base, deltaic depositional environment inferred. Figure 27 – Red line marks the truncated top of a channel missing a curved section which is defined by orthogonal joints; by length of the sides (arctan 2 m/3 m = 30^{o} tangent) yields a ~ 60^{o} bend. Large fossil “log” circled in large channel base which is much larger than the channels beneath; tape = 1 m).

Lateral Accretion Deposits vs. Cross-bedding

Point bars of a meandering steam migrate roughly perpendicular to channels in response to cut-bank erosion and deposition on the inside bends ("points") of meanders (e.g., Allen, 1982; the study of such deposits began in the 1920s on the tidal flats of the North Sea). A suite of bedforms (largely dunes in this case) may develop on the point bar surfaces in response to prevailing currents and sediment size/supply, and if preserved, form "lateral accretion deposits". As point bars migrate, the base of the depositional units may be preserved as low-angle bedding structures ("lateral accretion surfaces"; LAS) which reflect channel geometry (generally < ~ 15^o) and often display basal scour and pebble/fossil lags. So the dip direction of the larger-scale, low-angle LAS of point bars reflect channel migration direction (~ normal to channel trend) whereas the smaller-scale internal sedimentary structures (e.g., dune cross-strata) reflect current direction(s) which are generally channel-parallel. So a point bar has two scales of sedimentary structures/surfaces: usually much larger, low-angle (< 15-20^o) LAS and usually much smaller, cross-stratification (ideally 30-35^o but much less in coarse deposits and poor exposures). At LRC, the potential for overlap between LAS and foresets must be considered. Confusion is possible especially where channels are small, shallow and steep-sided and the lateral accretion deposits are thin and where coarse cross-strata can be difficult to recognize and may range into meter-scale.

Channels and point bar deposits are well exposed in places. Channel cross-sections are common at outcrops #4 and #5 as displayed on strike-joint (paleo-shore parallel) surfaces. The sense of flow direction is less clear but bi-directional cross-beds are present (Fig. 21). At outcrop #2 and #6, some channels weather out and overhang dramatically with steep (> 45^o) sides; some joint surface exposures are much less obvious. Developing criteria to distinguish tidal vs. distributary vs. fluvial channels would be useful especially for exposures on the western side of the hill yet to be studied.

Fluvial (Deltaic?) - Meandering Stream Deposits - Uppermost sequence

This sequence has been described at each outcrop; to summarize ~ 2 – 3 m thick channel/lateral accretion deposits with some reddish, well-oxidized strata and plant remains interpreted as prograding/meandering upper delta plain (beyond tidal influence) or coastal plain streams associated with the delta. No delta analogues are suggested since the exposures are not extensive and it's not clear how the three depositional sequences interrelate. However, if all of the small delta is exposed at LRC, it's at least two orders of magnitude smaller than the Niger delta (much larger drainage area and wave-dominated but with extensive tidal channels and flats; van Cappelle et al., 2016 provide an excellent review of tide-wave influenced deltas).

Figure 28 – "Caprock" at Outcrop #4

The caprock varies spatially and is generally similar to the underlying deposits with some reworking evident. The erosional remnants at outcrop #4 appear to be distinct fluvial deposits above deltaic deposits. At the base of a remnant here is an unusual deposit of large pebbles sandwiching a grayish sandstone with visible pores and abundant plant debris (Fig. 28 above). Given the position of the coarse pebble layers and possible air escape pores, rapid deposition by powerful unsteady currents is suggested; a tsunami, very large storm, or fluvial processes are possibilities. This pebble layer likely correlates with the caprock at outcrop #5 and elsewhere and needs further study. Within the deltaic sequence (outcrop #5), the uppermost caprock contains large (average 2-4 cm; up to 6+ cm) densely/randomly-packed vein-quartz pebbles (with some large red and brown sandstone and metamorphic clasts and red mudstone rip-up clasts not seen elsewhere) with abundant generally-aligned plant remains. And finally, with diligent search, wave-ripple laminated buff-colored sandstones with marine fossils (not seen insitu elsewhere) can be found draping the caprock in this area suggestive of a major flood event and a subsequent abrupt marine transgression.

CONCLUSION

The Salamanca Conglomerate records a high-energy Upper Devonian seacoast, with at least a meso-tidal range, as indicated by a pebbly beach, a tide-dominated delta prograding over marine wave-rippled fine sands, and a sub-aqueous large-scale dune field formed by strong flood tides. Most of the sequence records delta progradation and sediment transport/redistribution along shore to dunes and beaches by tides and waves. Well-exposed channel deposits at the top (which overlie wave-truncated dunes and beach deposits at a similar elevation) suggest a transition to a coastal plain environment including a major flood event as suggested by localized large clasts of quartz, sandstone, mud rip-up clasts, and abundant plant fossils followed by an apparent abrupt rise in relative sea level and a transgression as indicated by subsequent fine-grained wave-formed strata with an abundant marine fossil fauna.

ROAD LOG

Meeting Point: Rock City State Forest – Little Rock City Rd. at the DEC sign/State Forest boundary (two small parking lots; carpool if possible)

Meeting Point Coordinates: 42.225830, -78.710587

Meeting Time: 10:00 AM

Rock City State Forest is off Hungry Hollow Rd. which can be approached from US route-219 or State route-353.

All outcrops are short hikes on trail or just off the road.

Bring a lunch.

Be prepared for no facilities.

Latitude	Longitude	Stop or View Description
42.2280	-78.7092	STOP 1. The “Sentinels” - obvious from the first crest of Little Rock City Rd.; well visited but respect private property. Well-weathered, isolated blocks (two photos shown with the epilogue and Figs. 9 & 10); interpreted as tidal flats and shoals.
42.2265	-78.7137	STOP 2. The “North Face” - the highest outcrops (+10 m; which includes 3-4 m of marine strata) – shoreface/foreshore/channels; see Figs. 18, 19 & 20)..

42.2259	-78.7175	STOP 3. NW corner of outcrop belt is better exposed (sunny, less moss; "cleaner" rock faces), same interpretation as #2.
42.2217	-78.7113	STOP 4. just east of the first rise (escarpment) on Little Rock City Rd. within RCSF. A multitude of point bars (lateral accretion deposits) and channel fills. Interpreted as part of the deltaic sequence; likely mostly tidal channels and deposits.
42.2187	-78.7127	STOP 5. A few 100 meters south of #4, east side of the road, just inside treeline, extends for 100s of meters (nicely shown on as curvy strike joints on online orthoimagery (htps://orthos.dhses.ny.gov/). Outcrops appear largely filled with glacial debris – just 2 m of channel deposits exposed but with the coarsest caprock and fossiliferous sandstone.
42.2094	-78.7108	STOP 6. Follow the North Country National Scenic Trail (NCT) which enters the woods east of first campsite/shelter; perhaps the most interesting (confusing?) outcrop area. A large channel complex overlying fine-grained wave-rippled sandstone dominates the area (delta interpretation) with adjoining cross-strata of all scales.
42.2088	-78.7075	STOP 7. The NCT joins outcrop area #6 with #7 and following it south (white blazes) through Little Rock City provides a representative sampling of large scale cross-strata (tidal dune field) and overlying channel deposits (meandering streams).

EPILOGUE

Hall (1843) poetically summed up these formative geologic processes and the passage of time:

"Here was an ocean supplied with all the materials for forming rocky strata: in its deeper parts were going on the finer depositions, and on its shores were produced the sandy beaches, and the pebbly banks.

All, for aught we know, was as bright and beautiful as upon our ocean shores of the present day; the tide ebbed and flowed, its waters ruffled by the gentle breeze, and nature wrought in all her various forms as at the present time, though man was not there to say,

How Beautiful!"

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